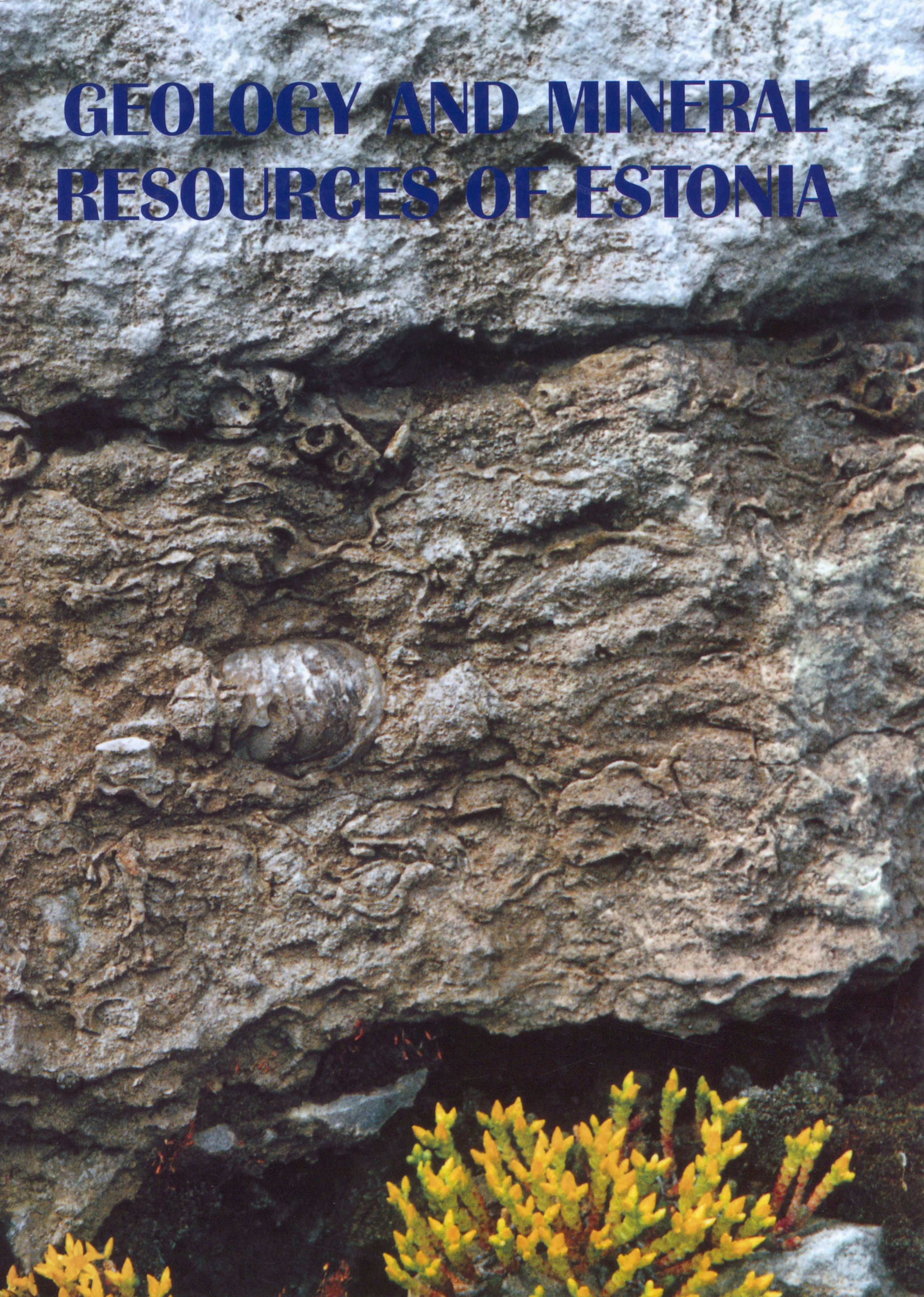
GEOLOGY AND MINERAL RESOURCES OF ESTONIA



INSTITUTE OF GEOLOGY

GEOLOGY AND MINERAL RESOURCES OF ESTONIA

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FOREWORD

The territory of Estonia has been inhabited at least ten thousand years. The first tribes to settle in the area were hunters and fishers who could already use local raw materials, such as crystalline erratic boulders, gravel, sand and clay. At about 6000 BP inhabitants learned to make earthenware from clay and around 5000-4000 BP to apply carbonate rocks to building of townlets and fortified settlements. Since 1230, lime has been widely used as a binder. Red bricks, made of local clays and used as building material for strongholds and churches, provide Estonia's historical buildings and architectural monuments with a specific geological splendour.

The first geological studies were carried out in Estonia more than 150 years ago. The long tradition of geological research in the area is due to the large and representative bedrock exposures providing excellent conditions for the study of Lower Palaeozoic rocks, and making Estonia a key region for solving several principal stratigraphic problems. The Palaeozoic rocks in Estonia enclose extraordinarily rich communities of well-preserved fossils, and a great number of new species and higher taxa have been established here. Ancient coastal formations of the Baltic Sea and relief forms, left behind by the last glaciation, are represented here more completely than in other regions. Within Estonia are found excellent examples of meteorite craters and the largest erratic boulders in northern Europe. All this makes Estonia's geology unique in several aspects.

For centuries Estonia has served as an economic, scientific and cultural bridge between the East and the West. Already in the Middle Ages it was an arena of serious ideological conflicts. At the end of the Livonian War (1558-83), southern Estonia fell under Polish rule. In the interest of restoring Catolicism, Jesuits opened a gymnasium in Tartu in 1583. After Estonia was taken over by Sweden, the Swedes founded a Protestant gymnasium in Tartu in 1630 to counterbalance the Jesuit school. In 1632, the Protestant gymnasium was changed into a university, which is one of the oldest and most prominent higher educational establishments in northern Europe. In the 17th century, Tartu University became a principal centre of education, science and humanistic ideas in the region. After re-opening in 1802, it developed into an outstanding centre of geological education and science in the former Russian Empire. The corresponding topics were also advanced in the Tartu (later Estonian) Society of Naturalists, founded in 1853.

In 1920, Estonian became the medium of instruction at Tartu University. In 1937, the Geological Committee of Estonia was founded. After the occupation and incorporation of Estonia into the Soviet Union, the most prominent geologists and a lot of promising young scientists left homeland and a new generation of geologists was trained. In 1947, the Institute of Geology of the Estonian Academy of Sciences was established and ten years later the Geological Survey of Estonia was founded. Both these institutions developed into important centres for geological, geophysical and environmental research in the northwestern portion of the Soviet Union and neighbouring countries.

The essential results of the research carried out during more than two centuries were summarised in multi-volume issues showing the directions and level of geological studies in Estonia (Geological Studies of the USSR, 50, Estonian SSR, Tallinn, 1968, 1972, 1973, 1974, 1977, 1984, 1987; History of Geological Sciences in Estonia, 1986). These, like most monographs in the field of geology issued in Estonia during the last decades, are in the Russian language and practically unknown to our western colleagues. Due to intensive drilling programmes and medium- and large-scale geological mapping, a lot of new geological information has been obtained. As there are currently no published general surveys on the geology and mineral resources of Estonia, the present monograph attempts to fill this gap. Its main purpose is not only to impart scientific information about Estonia's natural environment, but to serve also for industrial and agricultural purposes encouraging the sound use of mineral resources in the present-day Estonia.

Mining of mineral resources has inflicted incurable wounds on Estonia's nature. Another task of the present issue is to assist in drawing up main outlines of the strategy addressing improvement of the environment.

Most distinguished specialists of the Republic have participated in the compilation of this monograph. Its publishing has been made possible by the financial help of the Estonian Science Foundation (grant No. 1661), which is gratefully acknowledged. Thanks are due to the authors and all persons who have contributed to finalizing of this book. Special thanks go to Mrs. Helle Kukk for the revision of the English text, to Mr. Jüri Nemliher for the layout of this book and to Mr. Paul Pärkma for the drawings.

Anto Raukas and Aada Teedumäe

I LOCATION AND TOPOGRAPHY

The Republic of Estonia, the northernmost of the three Baltic States, is situated in the North-East of Europe, on the east coast of the Baltic Sea. The name *Estonia* is probably derived from *Aists*, the name the ancient Germans used to denote the Baltic tribes, living to the northeast of the Vistula River. In a written record the Aists *(Aesti, Aestorium gentes)* were first mentioned by the Roman historian Tacitus in the first century AD. The first written reference to the land of Estonians dates from 1154. On the order of Roger II, the king of Sicily, the Arab geographer and traveller Abu Abdallah Muhammad al-Idrisi designed a map of places in the world known in those times including Qalewany (Tallinn) in Astlanda (Estonia).

The territory of Estonia in nowadays boundaries extends from 57°30'34" to 59°49'12"N and from 21°45'49" to 28°12'44"E (Fig.1). The northernmost point of Estonia is on the Island of Vaindloo (the Cape of Purekkari on the mainland), the easternmost point in the Town of Narva, the southernmost point is the Naha farmstead at Mõniste, and the westernmost point is on the Island of Nootamaa (the Cape of Ramsi on the mainland). The extreme length of the Estonian territory is 350 km from west to east, and 240 km from north to south. The length of the Estonian coastline is 3,780 km; of this 1,242 km are on the mainland and 2,540 km are divided among the islands.

Estonia has an area of 45,215.4 sq km of which 9.2% is taken up by islands and 4.6% is under inland bodies of water. Climatically, Estonia belongs to the mixed-forest subregion of the Atlantic continental region of the temperate zone, which is characterized by warm summers and moderately mild winters.

Geologically, Estonia is situated in the northwestern part of the East-European Platform. Structurally, it lies for the most part within the boundaries of the southern slope of the Fennoscandian Shield with only its extreme southwestern and southern parts forming the wings of the Baltic Syneclise and the Valmiera-Lokno Uplift, respectively.

As part of the vast East-European Plain, Estonia is a generally flat country (Photo 1), where uplands and plateau-like areas alternate with lowlands, depressions and large valley-like forms. The average height above sea level is approximately 50 m, relative heights of landforms do not as a rule exceed 20 m, being only seldom 50 and more metres. About 40 per cent of Estonia's territory is at an absolute height of 50 to 100 m, and only one tenth has an elevation over 100 m above sea level (Fig. 1). The highest point in Estonia, the Suur Munamägi Hill (nearly 318 m), is located in the Haanja Heights.

Estonia displays a large variety of landscapes (Fig. 2). The northern part of the country consists of an extensive limestone plateau (Fig. 2), the northern edge of which forms a steep escarpment (Photo 2), known as the North-Estonian Klint (relative height up to 56 m). The narrow Fore-Klint Coastal Plain is situated in front of the Klint. The highest areas in the northern part of Estonia are the Pandivere Upland (166 m a.s.l.) and the Jõhvi Upland (81 m a.s.l.). To the south of the Pandivere Upland lies the gently sloping Vooremaa watershed (the Saadjärv Drumlin Field, with elevations up to 144 m a.s.l.).

Relatively high areas of North Estonia border on the Kõrvemaa and Alutaguse lowlands. To the south-west of the Pandivere Upland lies the Central-Estonian Plain which, gently sloping, passes over into the Võrtsjärv Depression. The Alutaguse Lowland turns into the Peipsi Depression.

In western Estonia the absolute height seldom exceeds 20 m and large areas are entirely flat. This is the region of the

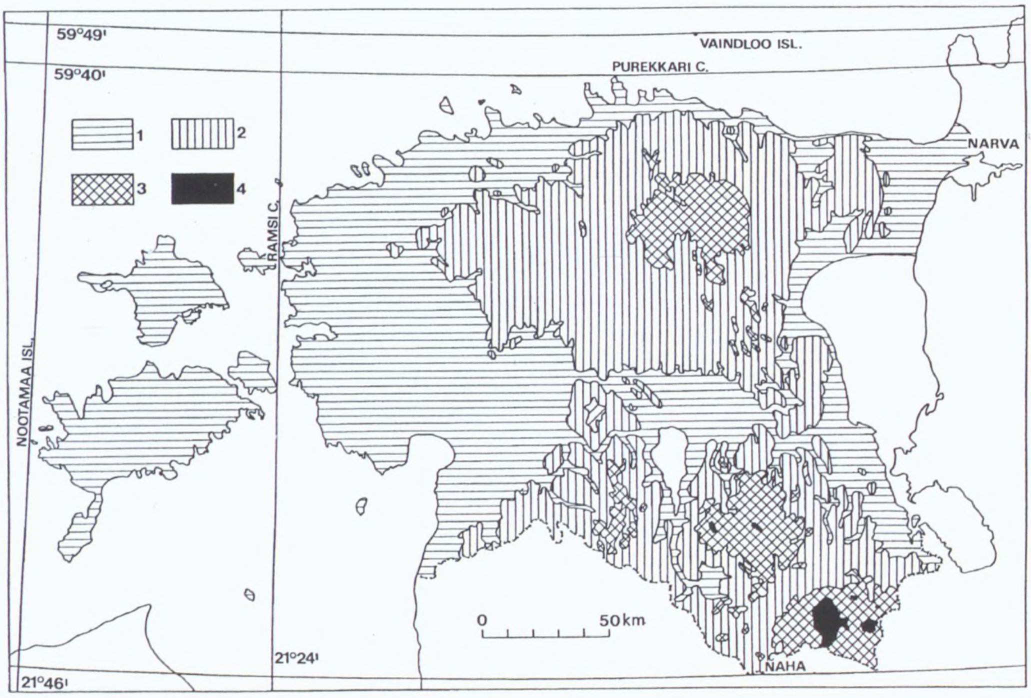

Fig. 1. Scheme of Estonian contemporary topography with isolines drawn at intervals of 50 m: 1 - 0...50; 2 - 50...100; 3 - 100...200; 4 - over 200.

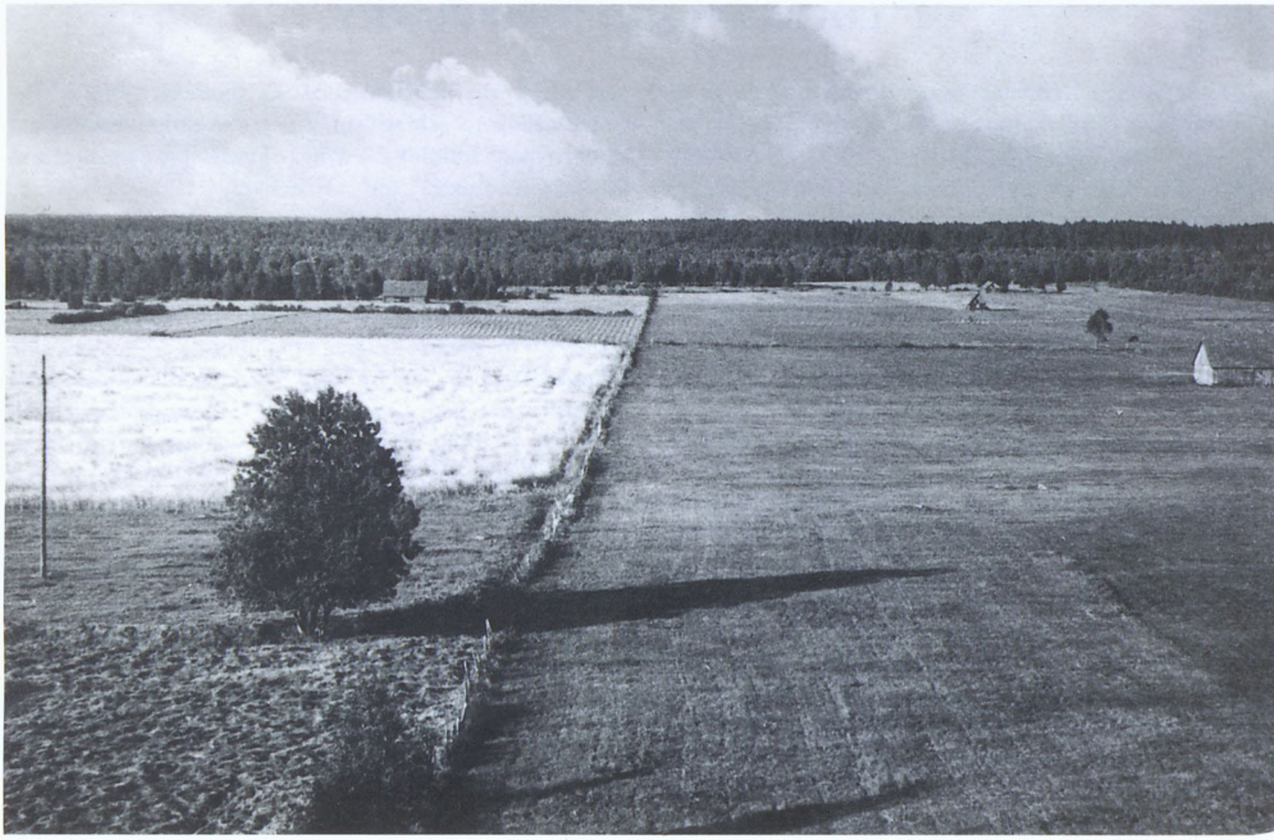

Photo 1. Estonia is a generally flat country. Coastal plain at Varangu, NE Estonia. *Photo by A. Raukas.*

Photo 2. The North-Estonian Klint at Ontika. *Photo by A. Rõõmusoks.*

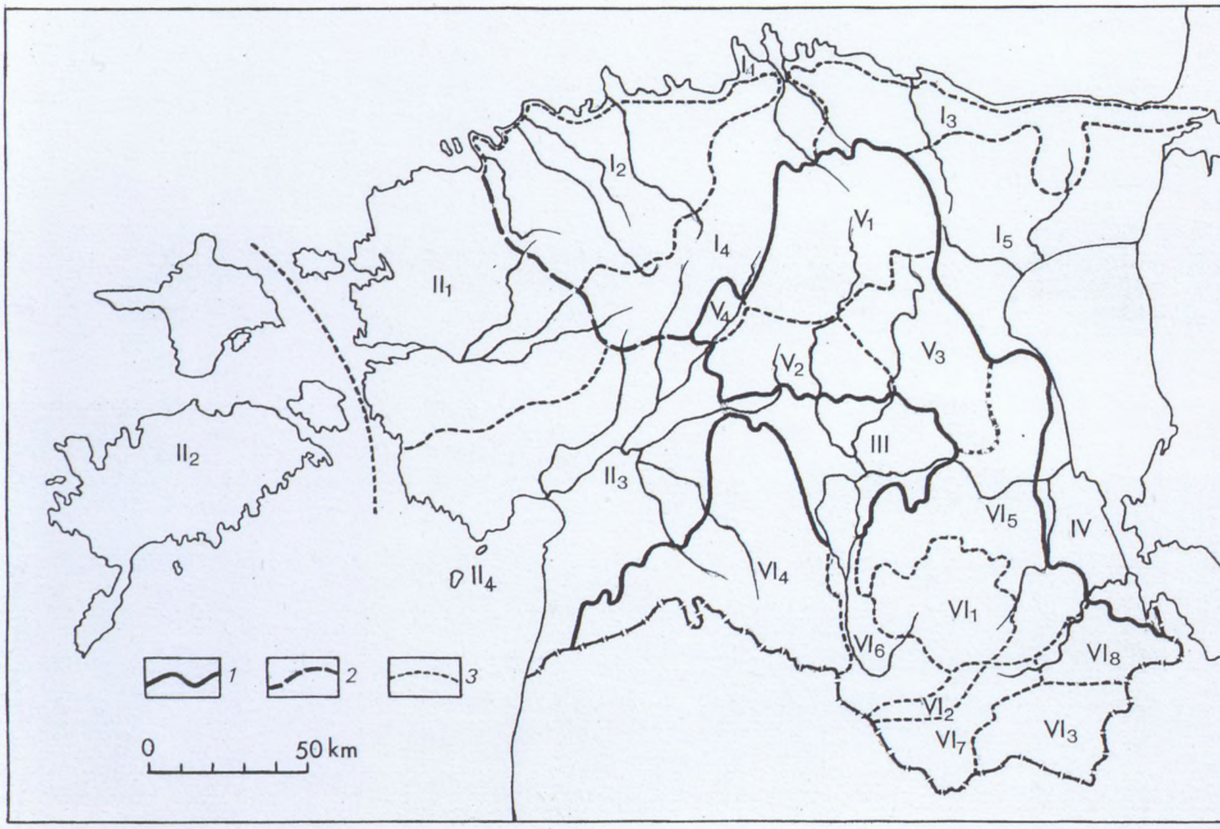

Fig. 2. Landscapes regions of Estonia after E. Varep: 1 - The boundary between Lower and Upper Estonia; 2 - the boundary between North and West Estonia; 3 - boundaries between landscape regions. **Lower Estonia**. I - North Estonia: I_1 - North-Estonian Coastal Plain with the Gulf of Finland's islands; I_2 - North-West Estonian Plateau; I_3 - North-East Estonian Plateau; I_4 - Kõrvemaa; I_5 - Alutaguse Lowland; II - West Estonia: II_1 - West-Estonian Lowland; II_2 - West-Estonian Archipelago; II_3 - Pärnu Lowland; II_4 - Islands of the Gulf of Riga; III - Võrtsjärv Lowland; IV - Peipsi Coastal Lowland. **Upper Estonia**. V - Intermediate Estonia: V_1 - Pandivere Upland; V_2 - Central-Estonian Plain; V_3 - Vooremaa; V_4 - Türi Drumlin Field; VI - South Estonia: VI_1 - Otepää Heights; VI_2 - Karula Upland; VI_3 - Haanja Heights; VI_4 - Sakala Upland; VI_5 - South-East Estonian Plain; VI_6 - Valga Depression with the Väike-Emajõgi Valley; VI_7 - Hargla Depression with the Võru Valley; VI_8 - Palumaa Plain.

lowlands of West Estonia and West- Estonian (Moonsund) Archipelago. Some small elevations are Kõpu (63 m a.s.l.), Middle-Saaremaa (54 m), Sõrve (36.6 m) and Tõstamaa-Varbla (44 m); the scarps of the islands of Saaremaa and Muhu and those in the western part of the mainland form the West-Estonian Klint (up to 21 m a.s.l.).

In South Estonia the topography is more varied and differences in the altitude are greater than elsewhere in Estonia. The area has four topographic highs (Fig.2): Sakala (up to 146 m a.s.l.), Otepää (217 m), Karula (137 m) and Haanja (318 m). They are separated from one another by the Valga and Hargla depressions and Võru Valley. The South-Estonian medium-height terrain (Ugandi Plateau, 40-100 m a.s.l.) is occupied by the South-West Estonian Plain.

The largest relief forms — plateaus, uplands, lowlands, depressions, the North-Estonian and West-Estonian escarpments (Aaloe & Miidel 1967) were formed in Pre-Quaternary times as a result of the long-term continental erosion (Таваст и Раукас 1982). Monoclinal bedding of bedrock strata and their different resistance to erosion resulted in the questa-like ancient topography (Орвику 1955). During all ice ages glacial erosion prevailed in North and West Estonia. These areas are characterized by a thin Quaternary cover and wide distribution of alvars against the background of Estonia's generally flat topography. The erosional relief forms here are represented by both small (glacial scratches, *etc.*) and large (rock drumlins, hollows and troughs of glacial ploughing) ones.

In the transitional zone between the prevailing glacial erosion and accumulation areas in central Estonia, the most characteristic relief forms are of the erosional-accumulative type. Among them are drumlins (Photo 3) and drumlin-like ridges, including megaflutings, which may reach 13 km in length, and 80 m in height (Saadjärv Drumlin Field). The accumulation area in southern Estonia features gently sloping and undulating till plains, and morainic hills, with the latter being especially common on accumulative insular heights (Otepää, Haanja). In places, dump and push moraines stretch some tens of kilometres in length with relative heights up to 50 m (West-Saaremaa Elevation).

Glaciofluvial accumulative relief forms are widely distributed in Estonia, with classic eskers and kame fields formed, as a rule, in passive or dead ice (Karukäpp & Raukas 1976). Radial eskers are most common on the Pandivere Upland (Photo 4) and marginal eskers on the West-Estonian Lowland (Раукас и др.1971). Fluvio- and limnoglacial kames either form separate fields or are scattered in hilly topography. As for the genesis, the glaciofluvial gravel and sandy plains are for the most part glaciofluvial deltas or outwash deltas. Less frequent are kame and glaciofluvial terraces and outwash cones.

Genetically and morphologically, the valleys of glacial meltwater discharge are diverse. These relief forms are most

Photo 3. Drumlins are distinctive relief forms in central Estonia. Within the Saadjärv Drumlin Field they sometimes exceed 10 km in length. *Photo by A. Raukas.*

typical of southern Estonia where in places they form orthogonal valley systems. They include both radial and marginal valleys, some of which are formed under the ice (e.g. rills of discharge), while others came into being by epigenetical superimposing on subglacial topography under the conditions of jointed passive or dead ice, but also due to glacial breaks and intense joining of ice-dammed lakes. Glacial meltwaters often flowed along ancient valleys which had developed before the last glaciation (Тавaст и Раукас 1982).

Characteristic of glacial terrain are also funnel- and saucer-shaped closed depressions — kettle holes, formation of which is associated with the melting of buried dead ice blocks (glaciokarst). Undoubtedly, in many cases, the process came to an end in the Late-glacial or at the beginning of the Holocene. However, it seems that some of the kettle-holes formed considerably later, with the process having started in the Boreal and coming to an end only in the Atlantic climatic period (Raukas & Rõuk 1995). Quite often kettle-holes are filled with peat, the thickness of which may reach 17 m.

In all stages of deglaciation considerable areas in front of glacier margins were occupied by glaciolacustrine basins of different ages (Raukas 1992a). These bodies of water have left behind deposits (mainly varved clays) and coastal relief forms (abrasional scarps, beach ridges, *etc.*) which are traceable at different levels, such as those on the slopes of the Otepää and Haanja heights.

The extensive glaciolacustrine plains, which were of great landscape-forming significance, occur only in the lower parts of the territory, particularly on the Alutaguse, Võrtsjärv, Peipsi and Kõrvemaa lowlands (Раукас и др. 1971). Due to the wide distribution of clayey deposits on the surfaces, modest absolute height and unfavourable drainage conditions these plains have undergone paludification, and many of them have turned into bog plains.

The effects of a variety of coastal processes can be found along both modern (Орвику 1974) and ancient coasts of the Baltic Sea (Раукас и др. 1965) and large lakes (Raukas & Tavast 1989). These features include wave-cut notches, scarps, abrasional platforms and plains, accumulative terraces, spits, barrier beaches, arrow-shaped spits, tombolos, bars, beach ridges, *etc.*). The development of the largest ancient coastal formations is connected to the transgressive phases of the Baltic Sea development (Раукас 1966).

More prominent aeolian relief forms (ridge-like longitudinal dunes, parabolic dunes, *etc.*) are spread along the ancient transgressive coastlines of the Baltic Sea. The height of dunes seldom exceeds 20 m. Formation of higher dunes was hindered by the small supply of sand, humid climate, continuing uplift of the Earth's crust and by several other circumstances (Raukas 1968). Beside contemporary and ancient coastal dunes there are also inland aeolian formations.

Karst topography and underground features (Photo 5), resulting from the solution of rocks and leaching processes, are common on the outcrops of carbonate rocks in northern, western and central Estonia, and of limited distribution in the southeasternmost part of the Republic (Heinsalu 1977). However, because of the relatively small thickness of soluble rocks, low absolute heights of the terrain, short duration of the postglacial evolution of the territory and for several other reasons, karst relief forms have rather modest dimensions in Es-

Photo 4. Roela esker on the Pandivere Upland. *Photo by A. Miidel.*

Photo 5. Karst topography at Uhaku above the Erra River, which flows underground along tectonic joints. *Photo by A. Aaloe.*

tonia. Nevertheless, several features have been identified here, including karren, karst holes, open jointings, karst relicts, funnel-sinks of absorption, angular subsidence sink-holes, underground and disappearing rivers, small caves, and other karst phenomena. In some places one may find peculiar temporary karst lakes.

Beside karst caves there are suffosion caves (Heinsalu 1987). For the most part, they are found along the slopes of the ancient South-Estonian valleys and scarps of the North-Estonian Klint. Calcium carbonate precipitates out of spring water forming dome-shaped travertine terraces on valley slopes.

Gravitational relief forms composed of debris are for the most part distributed at the foot of the North-Estonian and West-Estonian klints. In the hilly topography of southeastern Estonia, deluvial processes are ongoing due to human impact.

Biogenic (organogenic) relief forms were formed by the plant growth (phytogenic forms) and animal activities (zoogenic forms). The largest phytogenic relief forms having great landscape-forming significance are the bog or telmatogenic plains, with their hummocks, mochezhinas and other relatively small features (Masing 1977).

II HISTORY OF GEOLOGICAL RESEARCH

The essential natural resources of Estonia have been used for centuries. The first geological studies were carried out in the 17th century by members of staff at Tartu University. Publications of that time include G. Mancelius' paper on earthquakes (1619) and M. J. Herbinius' survey of waterfalls (1678). After reopening of Tartu University in 1802, the foundation was laid for the systematic scientific research into Estonia's geology.

Orviku and Viiding (Орвику и Вийдинг 1986) differentiated several stages, substages and periods in the history of geological research in Estonia. Within the first stage, which covered the time span from the beginning of human settlement in Estonia up to the reopening of Tartu University, they distinguished three substages: (1) gathering of elementary empiric knowledge about local geological monuments; (2) limited use of local mineral resources; (3) rapidly expanding use of local building materials after Estonia had been occupied by Germans, Danes and Swedes.

During the second stage, the most noteworthy events were the establishment of the Cabinet of Mineralogy at Tartu University in 1820, the foundation of the Tartu (later Estonian) Society of Naturalists in 1853, the University reform in 1892, the October Revolution in 1917, the birth of the independent Republic of Estonia in 1918, the incorporation of Estonia into the Soviet Union in 1940 and the restoration of the independent Republic of Estonia in 1991.

This chapter will deal only with the main outlines of the historical studies.

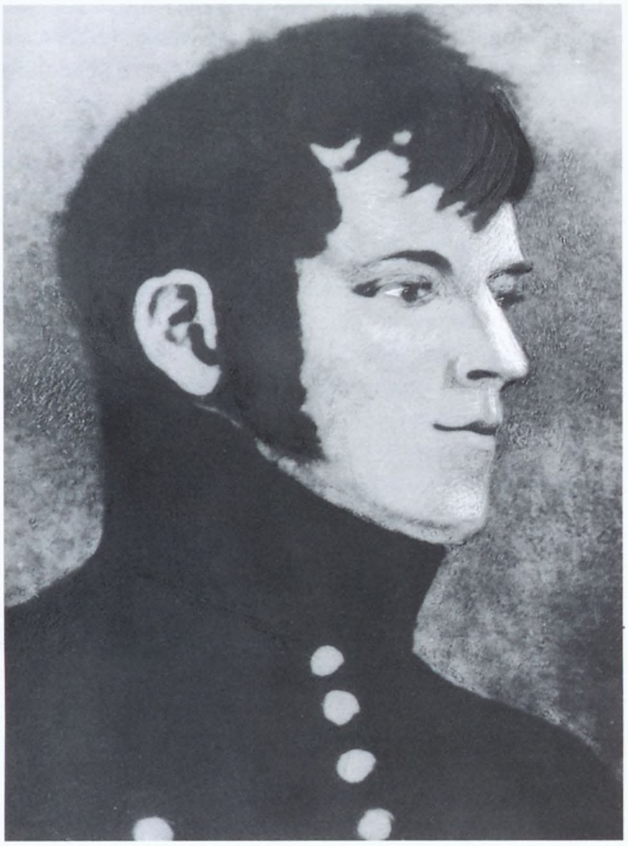

Photo 6. Moritz von Engelhardt (1779-1842) became the first professor of mineralogy at Tartu University in 1820.

Organisation of geological education, research and exploration

After reopening in 1802, the old (1632) Tartu (Dorpat, Derpt, Jurjev) University developed into an outstanding centre of geological education in the former Russian Empire. Moritz von Engelhardt (1779-1842), who was born in Estonia but received special education in Germany (Leipzig, Freiberg), became the first professor of mineralogy at Tartu University in 1820 (Photo 6). Students came to Tartu beside Estonian and Livonian districts also from Poland, Lithuania and other countries. Lecturing in the Estonian language started in 1920. Hendrik Bekker (1891-1925), the first Estonian professor of geology, defended his Ph.D. degree at London University in 1921 (Photo 7).

During 1920-40, only a few geologists were trained at Tartu University. A special governmental body - the Geological Committee (Survey) of Estonia was formed in 1937, but soon after Estonia was occupied by the Soviet Union in 1940, it ceased to exist. During the German occupation (1941-44) it operated as the Department of Geology of the Institute of Industrial Research and afterwards, with the similar staff, as the Department of Mineral Resources of the Central Research Institute of Industry of the Estonian S.S.R. In 1947, the latter gave its staff to the new institution - the Institute of Geology of the Estonian Academy of Sciences (Jaanusson 1994). Up to the beginning of the fifties this institute carried out both geological survey and basic research. The small scientific staff of the geological department of Tartu University concentrated its efforts in some special fields of research.

Photo 7. Hendrik Bekker (1891-1925) was the first professor of geology of Estonian nationality.

Only a few Estonian geologists survived through World War II. Since the end of the war, some 350 geologists have graduated from Tartu University. Several tens of Estonian-born geologists were trained at the universities in Moscow, Leningrad (St. Petersburg), Irkutsk, Tomsk (all in Russia) and Vilnius (Lithuania).

In 1957, the Geological Survey of Estonia was founded and the Institute of Geology became mainly engaged in basic studies. Since then, the geological, hydrogeological, geophysical and other studies, initiated by branch establishments of central institutions of the former Soviet Union after World War II, were gradually turned over to local organisations. In the course of complex studies aimed at geological mapping, mineral prospecting and exploration, tens of thousands boreholes penetrating deep into the sedimentary cover and more than 500 boreholes (Fig. 3) passing right through the sedimentary cover to reach the crystalline basement were made. Large-scale prospecting and exploration studies on oil shale and phosphorite were performed.

The stratigraphic studies aimed at compiling regional legends for geological mapping have always been of high priority in Estonia. Based on the results of large-scale field and laboratory studies and drilling programmes, researchers from the Geological Survey of Estonia began compilation of basic maps separately for the Palaeozoic bedrock and the Quaternary sediments, on a scale of 1 : 200 000. These were completed in 1975. The mapping on a scale of 1 : 50 000 is under way. A large number of complementary maps dealing with the geomorphology, hydrogeology, engineering geology, distribution of mineral resources, geological structure and several other aspects, and provided with the relevant comments, have also been compiled.

During the late 1970s and the 1980s, as a result of joint efforts of scientists of Estonia, Latvia, Lithuania and the Kaliningrad District of the Russian Federation, a set of 15 general geological maps of the East Baltic (1:500 000) was compiled and published (Григялис и Пуура 1980).

Evolution of scientific ideas

At the dawn of biostratigraphical studies in Estonia, E. Eichwald (1795-1876), first and foremost in several papers from 1840, introduced the terms Cambrian, Silurian and Devonian (*sensu* Murchison 1839) in the area (Photo 8). In 1856-58, Fr. Schmidt (1832-1908) proposed a reasonably detailed, palaeontologically and lithologically motivated subdivision of the northern Baltic Lower and Upper Silurian (now Cambrian, Ordovician and Silurian) rocks on stages level for the outcrop area, and an excellent geological sketch map.

In the last century it was already understood that the extraordinary survival of sediments, defined today as Vendian and Palaeozoic, in a nearly original low lithified level with perfectly preserved rich assemblages of fossils of calcitic, phosphatic and chitinous skeleton could be due to a very low tectonic compression and a small depth of burial which, according to recent estimation, never reached deeper than some 800 m during the whole geological history. In the middle of the last century, also the general feature of the regional geological structure - a homocline with a very gentle southward inclination of the weakly disturbed Palaeozoic strata overlying the crystalline ("granitic") basement, was described by Fr. Schmidt (Photo 9), C. Grewingk (1819-87) and several others.

Since the pioneer work of palaeontologists of several gen-

Photo 8. Eduard Eichwald (1795-1876) introduced the terms Cambrian, Silurian and Devonian into the Estonian scientific literature. He was the first in the Baltic provinces to consider continental glaciations.

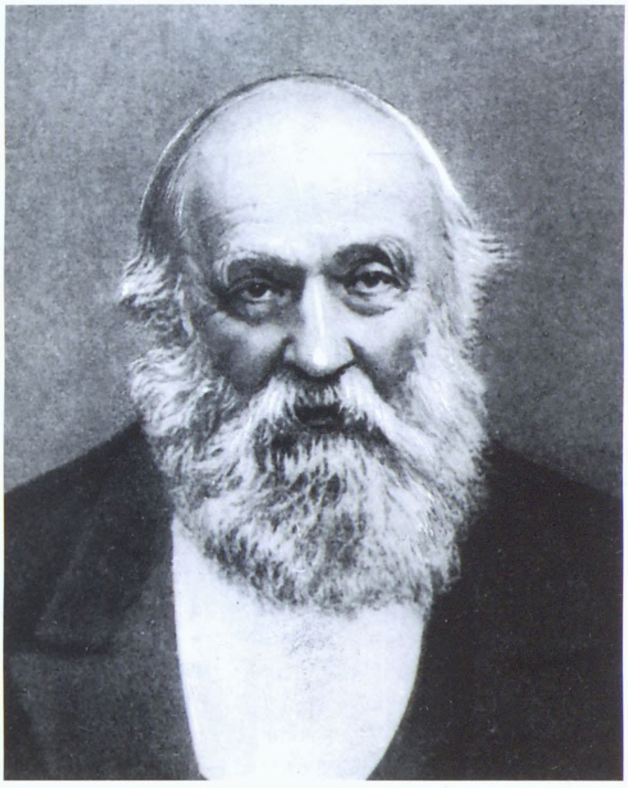

Photo 9. Friedrich Schmidt (1832-1908), the "father of Estonian geology", was the author of many outstanding papers dealing with the bedrock and Quaternary geology.

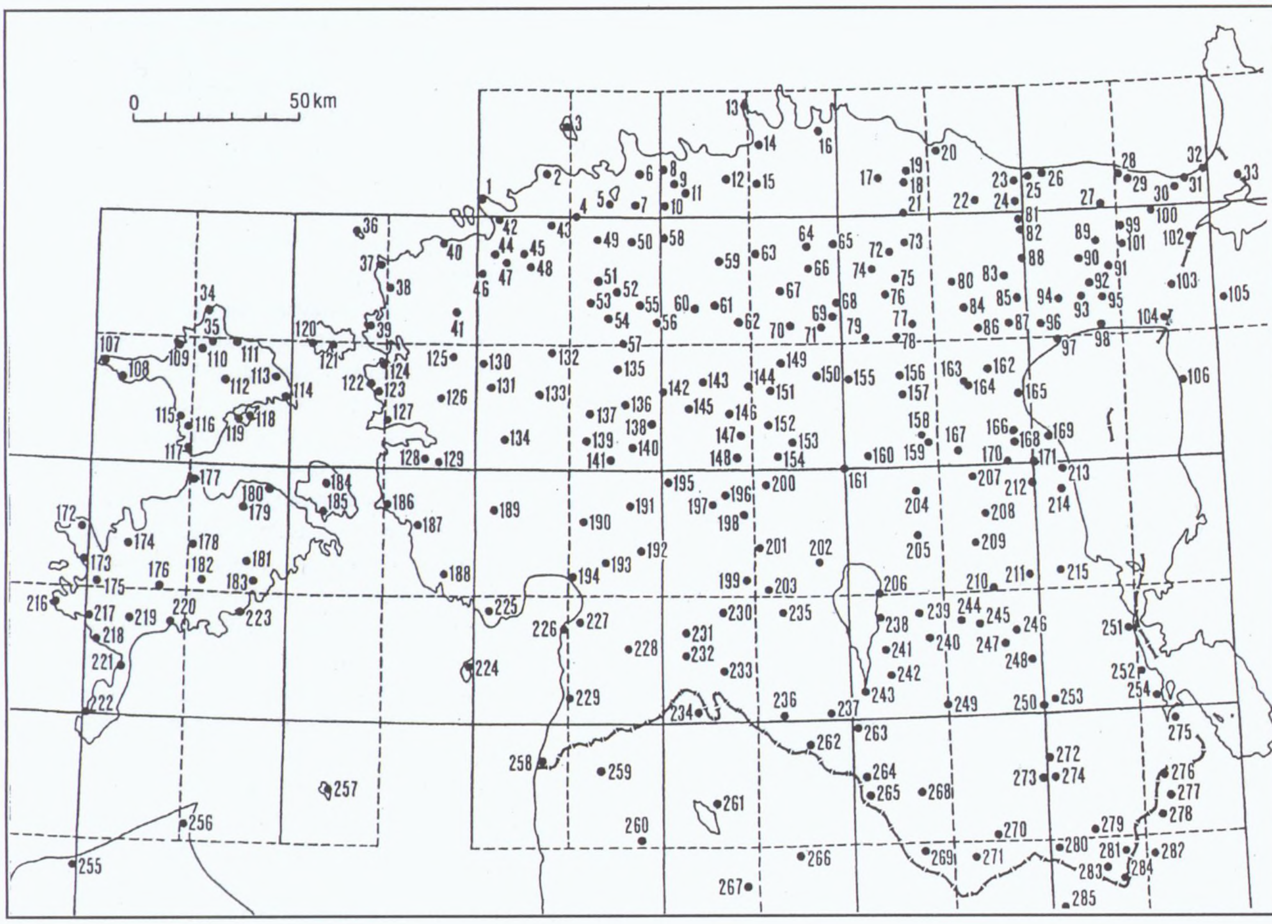

Fig. 3. Location map of main boreholes mentioned in the text. 1 - Pakrineeme, 2 - Muraste, 3 - Naissaare, 4 - Saue, 5 - Ülemiste, 6 - Lasnamägi, 7 - Assaku, 8 - Maardu, 9 - Kostivere, 10 - Aruküla, 11 - Jägala, 12 - Kiiu, 13 - Juminda, 14 - Pudisoo, 15 - Hirvli, 16 - Korjuse, 17 - Vanamõisa, 18 - Põdruse, 19 - Essu, 20 - Kunda, 21 - Rakvere, 22 - Uljaste, 23 - Kestla, 24 - Kiviõli, 25 - Purtse, 26 - Aa, 27 - Jõhvi, 28 - Toila, 29 - Oru, 30 - Sinimäe, 31 - Utria, 32 - Meriküla, 33 - Narva, 34 - Tahkuna, 35 - Kidaste, 36 - Osmussaar, 37 - Põõsaspea, 38 - Riguldi, 39 - Noarootsi, 40 - Vihterpalu, 41 - Seljaküla, 42 - Põllküla, 43 - Keila, 44 - Padise, 45 - Vasalemma, 46 - Audevälja, 47 - Rummu, 48 - Laitse, 49 - Saku, 50 - Nabala, 51 - Hageri, 52 - Lohu, 53 - Adila, 54 - Koigi, 55 - Härgla, 56 - Juuru, 57 - Rapla, 58 - Seli, 59 - Pihkva, 60 - Habaja, 61 - Ardu, 62 - Mustla, 63 - Vetla, 64 - Lehtse, 65 - Tapa, 66 - Ambla, 67 - Järva-Madise, 68 - Vajangu, 69 - Karinu I, 70 - Roosna-Alliku, 71 - Järva-Jaani, 72 - Assamalla, 73 - Koeravere, 74 - Porkuni, 75 - Pandivere, 76 - Väike-Maarja, 77 - Simuna, 78 - Kamariku, 79 - Puhmu, 80 - Roela, 81 - Savala, 82 - Oandu, 83 - Tudu, 84 - Paasvere, 85 - Oonurme, 86 - Venevere, 87 - Enniksaare , 88 - Talumaa, 89 - Pagari, 90 - Mäetaguse, 91 - Kaidma, 92 - Sõrumäe, 93 - Iisaku, 94 - Taga-Roostoja, 95 - Vaikla, 96 - Tudulinna, 97 - Rannapungerja, 98 - Alajõe, 99 - Rausvere, 100 - Viivikonna, 101 - Illuka, 102 - Mustajõe, 103 - Kuningaküla, 104 - Jaama, 105 - Slantsõ, 106 - Verholjane, 107 - Kalana, 108 - Kaleste, 109 - Kõrgessaare, 110 - Kurisuu, 111 - Paluküla, 112 - Määvli, 113 - Värsso, 114 - Soonlepa, 115 - Nurste, 116 - Lassi, 117 - Emmaste, 118 - Kassari, 119 - Orjaku, 120 - Förby, 121 - Hullo, 122 - Pusku, 123 - Asuküla 2, 124 - Haapsalu, 125 - Palivere, 126 - Martna, 127 - Kiideva, 128 - Kirikuküla, 129 - Lihula, 130 - Koluvere, 131 - Kullamaa, 132 - Vaimõisa, 133 - Paeküla, 134 - Rumba, 135 - Raikküla, 136 - Lihuveski, 137 - Valgu, 138 - Lokuta, 139 - Nurme, 140 - Kõnnu, 141 - Kaisma, 142 - Lelle, 143 - Piiumetsa, 144 - Paide, 145 - Käru, 146 - Türi, 147 - Vilita, 148 - Vaki, 149 - Vodja, 150 - Kahala, 151 - Kirila, 152 - Äiamaa, 153 - Taadikvere, 154 - Kabala, 155 - Oostriku, 156 - Vägeva, 157 - Kärde, 158 - Jõgeva, 159 - Ellavere, 160 - Sulustvere, 161 - Põltsamaa, 162 - Ulvi, 163 - Sadala, 164 - Tähkvere, 165 - Tiirikoja, 166 - Ruskavere, 167 - Palamuse, 168 - Halliku, 169 - Aadama, 170 - Saare, 171 - Pala, 172 - Undva, 173 - Jaagarahu, 174 - Mustjala, 175 - Viki, 176 - Kaarmise, 177 - Murika, 178 - Eikla, 179 - Järveküla, 180 - Pulli, 181 - Sakla, 182 - Valjala, 183 - Putla, 184 - Muhu, 185 - Suuremõisa, 186 - Virtsu, 187 - Paatsalu, 188 - Varbla, 189 - Koonga, 190 - Are, 191 - Tootsi, 192 - Jõhve, 193 - Sindi, 194 - Pärnu, 195 - Vändra, 196 - Vanaõue, 197 - Kootsi, 198 - Lahmuse, 199 - Heimtali, 200 - Võhma, 201 - Auksi, 202 - Survaküla, 203 - Viljandi, 204 - Kursi, 205 - Laeva, 206 - Väike-Rakke, 207 - Haavakivi, 208 - Saadjärve, 209 - Kärkna, 210 - Tartu, 211 - Kaagvere, 212 - Nõva, 213 - Kallaste, 214 - Alatskivi, 215 - Kavastu, 216 - Elsa, 217 - Kipi, 218 - Riksu, 219 - Kõrkküla, 220 - Kuressaare, 221 - Kaugatuma, 222 - Ohesaare, 223 - Nässumaa, 224 - Kihnu, 225 - Seliste, 226 - Tahkuranna, 227 - Uulu, 228 - Ristiküla, 229 - Häädemeeste, 230 - Asuküla, 231 - Kanaküla, 232 - Tõlla, 233 - Abja, 234 - Ipiku, 235 - Holstre, 236 - Taagepera, 237 - Tõrva, 238 - Tamme, 239 - Karijärve, 240 - Elva, 241 - Valguta, 242 - Aakre, 243 - Pikasilla, 244 - Kemeri-Loku, 245 - Tatra, 246 - Uniküla, 247 - Sirvaku, 248 - Vastse-Kuuste, 249 - Otepää, 250 - Kioma, 251 - Mehikoorma, 252 - Raigla, 253 - Põlva, 254 - Karisilla, 255 - Ovisi, 256 - Kolka, 257 - Ruhnu, 258 - Ikla, 259 - Staicele, 260 - Buikule, 261 - Burtnieki, 262 - Holdre, 263 - Jõgeveste, 264 - Valga 10, 265 - Valga 324, 266 - Strenči, 267 - Valmiera, 268 - Karula, 269 - Laanemetsa, 270 - Vartu, 271 - Mõniste, 272 - Väimela, 273 - Võru, 274 - Võru-Kubija, 275 - Värska, 276 - Petseri, 277 - Deksino, 278 - Panikovichi, 279 - Tsiistre, 280 - Luutsniku, 281 - Vungi, 282 - Laura, 283 - Hino, 284 - Parmu, 285 - Alūksne.

erations, tens of thousands specimens of fossils have been collected during systematically arranged field studies. The collections stored at the Institute of Geology and at the Geological Museum of Tartu University include about 1000 type-specimens of new species of fossils. A great number of new genera and families have been established. The most numerous groups of fossils, mostly Ordovician and Silurian, which have been studied in particular detail, are marine invertebrata: articulate and inarticulate brachiopods (studied by A. Rõõmusoks, M. Rubel, L. Hints, I. Puura during the last decades), molluscs of different classes, arthropoda like trilobites (Reet Männil), ostracods (L. Sarv, T. Meidla), merostomata and different echinodermata (Ralf Männil), tabulate (E. Klaamann) and rugose (D. Kaljo) corals and stromatoporoids (H. Nestor), bryozoans (Ralf Männil), graptolites (D. Kaljo), conodonts (V. Viira) *etc*. Besides, chitinozoans of problematic origin (Ralf Männil, J. Nõlvak, V. Nestor) as well as fossil elements and fragments of endo- and exoskeleton of early representatives of vertebrata from the Late Silurian and Devonian (E. Mark-Kurik, T. Märss) have also been studied.

Based on the data on the distribution of fossils in the Vendian - Middle Palaeozoic sedimentary record, a stratigraphic scheme for the Baltic States and the East-European Platform, as a whole (Ralf Männil, D. Kaljo, K. Mens, E. Mark-Kurik, H. Nestor, L. Hints *et al.*) and a detailed biostratigraphic subdivision of the Ordovician and Silurian rocks in the North Baltic area were elaborated. The ecostratigraphic studies, taking into account peculiarities in the distribution of different fossils in lateral changes of lithologies, provided a basis for the correlation of the successions of different facial belts of the Early Palaeozoic pericratonic Baltic Basin (Reet Männnil, D.Kaljo a.o.). An impact of global and regional geologic/tectonic and palaeoenvironmental processes, like the continental drift of the Baltic Continental Plate from the south polar position to the northern tropics during the Late Vendian - Devonian, the continental glaciation in Gondvana in the Late Ordovician, causing ocean-level changes, and pericratonic tectonic activities along the continental palaeomargins with specific influence on sedimentation and fossil communities, has been fixed in the geological sedimentary record of the Baltic Palaeozoic marine basin.

Estonia was among the first regions where the theory of continental glaciation in the middle of the last century was applied (E. Eichwald, Fr. Schmidt) and later the structure and formation of different landforms and the evolution of the Baltic Sea were described in detail (Karl Orviku, H. Kessel, A. Raukas, E. Rähni a.o.).

Structural studies

Up till the 1950s, the area of Estonia was structurally classified as a simple, almost horizontal homocline. Detailed investigations of the sedimentary cover and the studies carried out within large-scale drilling and geophysical survey programmes revealed a typical fault-and-block tectonic pattern of both the sedimentary cover and the underlying crystalline basement (Вахер и др. 1962, Пуура 1979). Classifications of the fault-related linear structures were suggested during the detailed structural studies of the oil shale (Пуура 1986) and phosphorite basins (Пуура 1987). The recent local crustal movements and weak earthquakes occurring against the background of the postglacial crustal uplift (Орвику 1960b) are, to a certain extent, related to faulting lines (Sildvee & Vaher 1995). The tectonic map with the accompanying explanatory text for the whole East Baltic territory was published by Suveizdis *et al.* (Сувейздис 1979).

Research into the crystalline basement and the Vendian-Cambrian strata

The existence of the crystalline ("granite") basement under the sedimentary cover was recognized in Estonia in the middle of the 19th century (Schmidt 1881), but systematic geological research into the basement was initiated not until the 1960s. The studies showed that in northern and north-eastern Estonia the basement consisted mainly of migmatized metamorphic rocks of the amphibolite facies and in southern and southwestern Estonia of granulite facies (Пуура и др. 1976). The recent Sm-Nd dating of folded rock assemblages indicated Palaeoproterozoic age of the orogenic continental crust in southern Estonia (Puura & Huhma 1993) and in the whole Byelorussian - Baltic granulite province (Gorbatschev & Bogdanova 1993). Thus, the basement of Estonia was considered as a continuation to the Svecofennian Domain of the Fennoscandian Shield.

The first evidence about the Vendian rocks was obtained in 1842-45 from a 90-m-deep borehole in Tallinn (Helmersen 1851). The first borehole, which reached the crystalline basement at a depth of 162.8 m, was drilled in 1898-99 at Aseri. According to Sokolov (Соколов 1953), the thickness of the Vendian complex was 92.5 m there (Rüger 1923). As a result of recent studies (Менс и Пиррус 1971, 1974, 1980, 1986), a great lateral and vertical facial variability has been established and a detail lithostratigraphical subdivision presented.

As early as the first half of the 19th century, the Cambrian sequence was divided into a lower, Blue Clay Unit and an upper, Sandstone Unit (Engelhardt 1820, Strangways 1822, Eichwald 1825). Following Murchison's concept of the Silurian system, Eichwald (Эйхвальд 1840) included these two units into the Silurian, and Schmidt (1858), more specifically, into the Lower Silurian. Within the Cambrian sandstone upon the blue clay, Linnarsson (1873) suggested to distinguish units equivalent to the Swedish Euphyton Sandstone and Fucoid Sandstone units. These two terms were widely used in Estonia until denomination on the basis of geographical names (Öpik 1933).

A great step forward was A. Mickwitz's discovery of malacofauna in the Eophyton Sandstone (Schmidt 1888) and the finds of an olenellid trilobite which proved Lower Cambrian age of these beds, and also the finds of the brachiopod *Mickwitzia* which indicated the same age with the correlatable beds in Sweden. Öpik (1925, 1926, 1929) contributed further important biostratigraphical data on the Lower Cambrian and revised the terminology (Öpik 1933). He (Öpik 1956) was convinced that in places the terrigenous beds dated as the basal Ordovician could be of Late Cambrian age. An increased access to numerous cores of borings all over Estonia enabled the subsurface Cambrian to be studied in more detail. Since the 1960s, many papers dealing with various aspects of the Lower Cambrian sequence, including a detailed lithostratigraphic classification (Менс и Пиррус 1977), have been published. In southeastern Estonia, some parts of the terrigenous sequence were supposed to belong to the Upper Cambrian (Волкова и др. 1981). Intense biostratigraphical studies aimed at determining a convenient basis for the Ordovician System indicate that a comprehensive lower portion of the sandstone, previously regarded as basal Ordovician, is actually Late Cambrian in age.

History of Ordovician research

The study of the Ordovician strata in Estonia was commenced with the descriptions of the North-Estonian Klint (Севергин 1808, 1809). Engelhardt (1820), Strangways (1821, 1822) and Eichwald (1825) pointed out the similiarity of the North Estonian and Scandinavian sequences. Strangways (1822) produced the first geological-lithological map which included also Estonia.

In the history of Cambrian-Silurian stratigraphy of Estonia, the second half of the 19th century is known as the Schmidt's Epoch (Мяннил 1986). Friedrich Schmidt (1832-1908), a descendant of Baltic Germans, published a comprehensive survey of the Cambro-Silurian outcrop area in northern Estonia (Schmidt 1858) which included the stratigraphic classification of the Ordovician sequence in the East Baltic area and a bedrock map. The units of Schmidt (Table 1) are neither strictly lithostratigraphical nor based on guide fossils, but reflect the stages of development of the regional benthic fauna without any preference for any particular group of fossils. In modern terms, they could be compared to topostratigraphical units. This approach, together with the use of geographical names for the units, was followed by subsequent generations of geologists in the research of the Estonian bedrock.

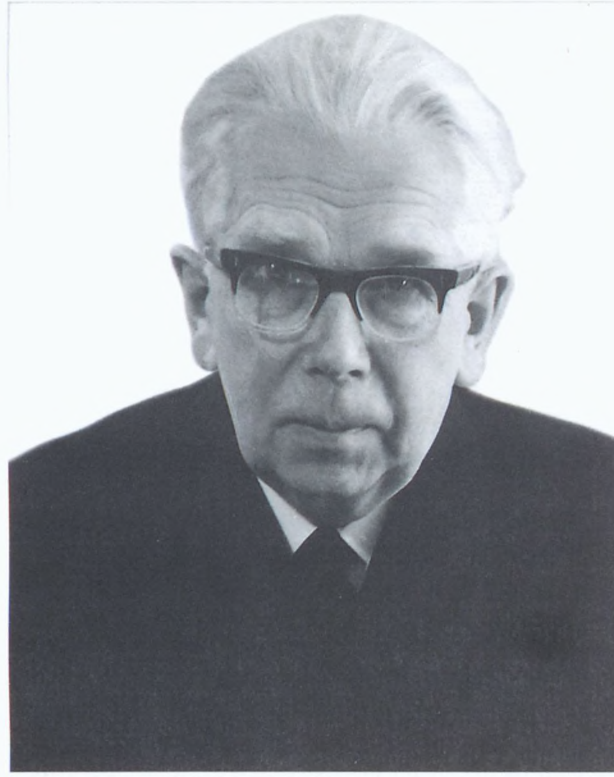

Photo 11. Karl Orviku (1903-81) studied the lithostratigraphy of Estonian bedrock, Quaternary deposits and landforms. He was the main initiator of research into the history of geology, and the teacher of most of the authors of this monograph.

Photo 10. Armin Öpik (1898-1983) was the successor of Bekker on the chair of geology at Tartu University. He was a distinguished stratigrapher, but his main interest was in the field of palaeontology, documented by a number of famous monographs on brachiopods, trilobites, ostracodes and other groups.

Based on a detailed study of the post-Tremadocian Lower Ordovician in Ingria, Lamansky (Ламанский 1905) introduced a stratigraphic classification. It was primarily based on ranges of trilobite species which he could follow into northern Estonia. An important result of his study was the recognition of numerous breaks in the correlatable North-Estonian sequence.

In the late 19th and early 20th century, a number of important taxonomic monographs were published by both native and foreign specialists (Rosen 1867, Dybowski 1877b, Schmidt 1874, Pahlen 1877, Holm 1886, Mickwitz 1896, Stolley 1896b, Koken 1897, 1925, Hoyningen-Huene 1899, Jaekel 1899, Bonnema 1909), and the data on the Ordovician of Estonia was included in several other publications as well. Of particular importance was the series of monographs on trilobites by Schmidt (1881, 1885, 1894, 1898, 1901, 1904, 1906, 1907). With Bassler's (1911) monograph on bryozoans the former Lower Silurian Series was, as far as the Estonian sequence was concerned, definitely elevated to the rank of a system, termed Ordovician.

In 1914, the renowned American geologists P.E. Raymond and W.H. Twenhofel examined Cambro-Silurian exposures in Ingria, Estonia and Scandinavia with the purpose of attempting a correlation with eastern North America. Raymond (1916) proposed five new terms (Table 1) and divided the Ordovician System in the Baltoscandian Basin into three series: Lower, Middle and Upper Ordovician. According to current correlations, the boundaries of these series correspond very closely to those between the Ordovi-

Table 1. Development of the Ordovician stratigraphical scheme in Estonia (see also Tab. 7)

Schmidt 1858	Schmidt 1879, 1881	Schmidt 1897,1898	Raymond 1916	Bekker 1922, 1925	Orviku 1940	Jaanusson 1944b	Männil and Meidla 1994	
Not known								
Borkholm	Borkholm F_2		Borkholm	Porkuni Stage		Porkuni Stage F_2	Harju Series	Porkuni Stage F_{II}
Lyckholm	Lyckholm F_1	F_1b	Lyckholm	Saaremõisa Stage		Pirgu Stage F_Ic		Pirgu Stage F_Ic
						Vormsi Stage F_Ib		Vormsi Stage F_Ib
		F_1a				Saunja Stage F_Ia		Nabala Stage F_Ia
Not known								
Wesenberg	Wesenberg E		Wesenberg	Rakvere Stage			Viru Series	Rakvere Stage E
Jewe	Wassalem D_3		Kegel	Vasalemma Keila Stage				Oandu Stage D_{III}
	Kegel D_2							Keila Stage D_{II}
	Jewe D_1		Jewe	Jõhvi Stage				Jõhvi Stage D_I
	Itfer C_3		Itfer	Idavere Stage				Idavere Stage C_{III}
Brand-Schiefer	Kuckers C_2		Kuckers	Kukruse Stage				Kukruse Stage C_{II}
Vaginaten-kalk	Echino-sphaeriten-kalk C_1	Echino-sphaeriten-kalk C_1b	Reval	Tallinna Stage	Uhaku Stufe			Uhaku Stage C_Ic
			Dubowiki		Lasnamägi Stufe			Lasnamägi Stage C_Ib
		Obere Linsen-schicht C_1a		Aseri Stage	Aseri Stufe			Aseri Stage C_Ia
	Vaginaten-kalk B_3	B_3b / B_3a	Kunda	Kunda Stage			Oeland Series	Kunda Stage B_{III}
Chloritkalk	Glaukonit-kalk B_2	B_2b / B_2a	Walchow	Paldiski Stage				Volkhov Stage B_{II}
Grünerde	Glaukonit-sand B_1							Latorp Stage B_I
Tonschiefer Unguliten-sandstein	Dictyonema-schiefer A_3		Packerort	Pakerorti Stage				Varangu Stage A_{III}
	Unguliten-sand A_2							Pakerort Stage A_{II}

cian series of North America. Bekker (1921) introduced the term Stage (Table 1) for the basic chronostratigraphic unit (= beds in Schmidt 1881) and revised the spelling of geographic names to accord modern maps (Bekker 1922, 1925). Armin Öpik (1898-1983), the successor of Bekker on the chair of geology at Tartu University (Photo 10), focused, as far as the Ordovician stratigraphy is concerned, on studying the Tremadocian sequence (Öpik 1929). However, his main contribution was in the field of palaeontology, documented by a number of monographs on brachiopods (Öpik 1930b, 1933, 1934, *etc.*), trilobites (1937b), ostracodes (1937a) and some other groups. In a manuscript in 1934, he proposed the terms Viru Series for the Middle Ordovician and Harju Series for the Upper Ordovician of the region, which were first published by Luha (1940a). Karl Orviku (1903-81) made a noteworthy contribution to knowledge of the Ordovician in Estonia mainly in the field of detailed lithostratigraphy (Photo 11). His excellent monograph on the lower Middle Ordovician of northern Estonia (Orviku 1940), in which he distinguished the Lasnamägi and Uhaku stages (Table 1), exerted great influence on the succeeding generations of Estonian geologists. His analysis of discontinuity surfaces (hardgrounds), in particular, both with regard to their morphology and importance as markers of stratigraphic breaks, received wide attention. In subsequent papers Orviku (Орвику 1960a, б) gave a detailed lithostratigraphic classification of the Lower Ordovician Volkhov and Kunda stages of northern Estonia.

The scientific activity of the succeeding generation (Jaanusson, Männil, Rõõmusoks, Kaljo, Aaloe and several others) started in an organisation known as the Section of Geology of Gustavus Adolphus Natural Science Circle. As a result of field studies and later publications, the Schmidt's hitherto poorly known Lyckholm beds (Table 1) were divided into three separate stages (Jaanusson 1944b, 1956) and the stratigraphy of the Viru Series (Middle Ordovician) of northern Estonia was revised (Jaanusson 1945). In the upper Middle Ordovician, the occurrence of K-bentonites was observed (Jaanusson & Martna 1948), and their importance for correlation and stratigraphic classification recognized (Юргенсон 1958a, Bergström *et al.* 1995).

The progress in knowledge of the Ordovician sequence in Estonia after the war was largely due to the availability of numerous cores of drill borings. Ralf Männil (1924-90) solved the somewhat confused stratigraphic terminology of the Schmidt's unit F_1a by introducing the term Nabala Stage (Мяннил 1958г). He also showed (Мяннил 1958 б , в, 1960) that information from borings indicated Schmidt's Vasalemma beds to represent a lithostratigraphic unit which also includes the beds of Keila Age. For the chronostratigraphic unit above the Keila Stage Männil proposed the term Oandu Stage, borrowed from the poorly exposed lithostratigraphic unit Oandu beds in northeastern Estonia (Öpik 1934) which was unknown to Schmidt. Kaljo, Rõõmusoks and Männil (Кальо и др. 1958) emphasized the need of regional terms for Ordovician series and proposed the Oeland Series for the Lower Ordovician in the accustomed Baltoscandian usage. The post-Tremadocian Lower Ordovician was renamed Ontika Subseries.

As a result of a thorough examination of the subsurface Ordovician in Estonia, Latvia and Lithuania, Ralf Männil (Мяннил 1966, preliminary report 1964) summarized the development of the Baltoscandian Basin in the Ordovician Period. He showed that in an extensive west-central area of the East Baltic region (Livonian Tongue after Jaanusson 1976), including southern Estonia, the post-Tremadocian Ordovician rocks are basically of the same type, both lithologically and faunistically, as on the Swedish mainland north of Scania, and that they belong to a single Swedish-Latvian Facies Zone (Central Baltoscandian Confacies, Jaanusson 1976). The Estonian Facies Zone (North Estonian Confacies Belt, Jaanusson 1976) was proved to be conspicuously different in many respects. Põlma (Пылма 1967, 1972, 1982) presented a comparative analysis of the carbonate lithology of these main regions and distinguished a lithologically transitional belt, characterised by transitional lithologies and interlocking pattern of various lithofacies from the south and north. The relation of the boundaries of the transitional belt to those of confacies belts is still disputed (Jaanusson 1995, Meidla 1996). Of great palaeogeographical interest was the discovery that the Varangu beds (Мяннил 1958a), which occur in a limited area of northern Estonia, correlate with the upper Tremadocian Ceratopyge Shale in Scandinavia (Вийра и др. 1970, Кальо и Кивимяги 1970). The stratigraphy of the Middle Ordovician (Viru) of North Estonia (stratotype area) was summarised in a comprehensive monograph by Rõõmusoks (Рыымусокс 1970), and that of the Ordovician of Estonia in general, in a chapter of a separate book by the same author (Rõõmusoks 1983).

The rich knowledge of the Ordovician fauna of Estonia has increased the precision of correlations not only within Estonia but also with the Ordovician sequences elsewhere (Nõlvak & Grahn 1993, Männil & Meidla 1994, Jaanusson 1995).

History of Silurian research

M. von Engelhardt (Engelhardt & Ulprecht 1830) was the first to draw attention to the differences between the rocks now classified as Ordovician and Silurian. He noted that on mainland Estonia limestones containing orthoceratite cephalopods and trilobites were succeeded southwards by younger rocks with corals and pentamerid brachiopods.

In the monograph by Murchison, Verneuil and Keyserling (1845) the Silurian sequence of Estonia is briefly but adequately summarised. Murchison distinguished within the Silurian of Estonia (in the current sense) the following three units in ascending order (1) Pentamerus Limestone, (2) Limestone with corals, and (3) Limestone with Terebratulas.

Schrenk (1854) gave the first comprehensive lithological survey of the Silurian localities in Estonia. Schmidt (1858) continued the study in a much greater detail and with emphasis on fossil fauna. His classification of the Silurian sequence of Estonia into three groups is roughly comparable to Murchison's tripartite subdivision and the distinguished groups correspond to the Llandoverian, Wenlockian and Ludlovian + Downtonian Series in the current sense (Table 2). The lower, Schmidt's Group with Smooth Pentamerids (= Llandoverian), was subdivided into smaller units and the base of the Upper Silurian defined, in the outcrop area of northern Estonia, at the level of the present Ordovician-Silurian boundary in the area. Schmidt interpreted the Borealis Banks as local mass accumulation of shells (Muschelbänke) and regarded the unit to be somewhat artificial in a chronostratigraphic sense. Sub-

Table 2. Development of the Silurian stratigraphical scheme in Estonia (see also Table 8)

<table>
<tr><th colspan="2">Schmidt 1858</th><th>Шмидт 1879, Schmidt 1881</th><th>Schmidt 1892</th><th colspan="2">Twenhofel 1916</th><th>Bekker 1922</th><th colspan="2">Bekker 1925</th><th>Luha 1933</th><th colspan="2">Кальо 1970в</th><th colspan="2">Решения... 1987</th></tr>
<tr><td colspan="2" rowspan="5">Obere Oesel'sche Gruppe</td><td rowspan="5">Obere Oeselsche Schicht K</td><td rowspan="3"></td><td rowspan="5">Oesel Formation</td><td rowspan="3">Kaugatoma Zone</td><td rowspan="5">Upper Saaremaa Stage K</td><td rowspan="5">Saaremaa Stage</td><td>Ohesaare Substage</td><td>Ohesaare Stage K_4</td><td rowspan="2">Downton</td><td>Ohesaare Stage</td><td rowspan="2">Přidoli</td><td>Ohesaare Stage</td></tr>
<tr><td rowspan="2">Kaugatoma Substage</td><td rowspan="2">Kaugatoma Stage K_3</td><td>Kaugatuma Stage</td><td>Kaugatuma Stage</td></tr>
<tr><td rowspan="3">Ludlow</td><td>Kuressaare Stage</td><td rowspan="2">Ludlow</td><td>Kuressaare Stage</td></tr>
<tr><td>Ilionia (Didyma) Schichten</td><td rowspan="2">Sagaristi Zone</td><td>Paadla Substage</td><td>Paadla Stage K_2</td><td>Paadla Stage</td><td>Paadla Stage</td></tr>
<tr><td>Eurypterus Schichten</td><td>Rootsiküla-Kaarma Substage</td><td>Kaarma Stage K_1</td><td>Rootsiküla Stage</td><td rowspan="3">Wenlock</td><td>Rootsiküla Stage</td></tr>
<tr><td colspan="2" rowspan="2">Untere Oesel'sche Gruppe</td><td rowspan="2">Untere Oeselsche Schichten-gruppe J</td><td rowspan="6"></td><td colspan="2" rowspan="2">St.Johannis Formation</td><td rowspan="2">Lower Saaremaa Stage J</td><td rowspan="2">Jaani Stage</td><td>Tagamõisa & Vilsandi-Panga Substage</td><td>Muhu-Kurevere Stage J_2</td><td rowspan="2">Wenlock</td><td>Jaagarahu Stage</td><td>Jaagarahu Stage</td></tr>
<tr><td>Suuriku Substage</td><td>Jaani Stage J_1</td><td>Jaani Stage</td><td>Jaani Stage</td></tr>
<tr><td rowspan="4">Gruppe der glatten Pentameren</td><td>Pentamerus ehstonus Zone</td><td>Pentamerus estonus Kalk H</td><td colspan="2">Addifer Formation</td><td>Adavere Stage H</td><td colspan="2">Adavere Stage</td><td>Adavere Stage H</td><td rowspan="4">Llandovery</td><td>Adavere Stage</td><td rowspan="4">Llandovery</td><td>Adavere Stage</td></tr>
<tr><td>Zwischen-zone</td><td>Raiküllsche Schicht G_3</td><td rowspan="3">Tamsal Formation</td><td>Raiküll Beds</td><td>Raiküla Stage G_3</td><td colspan="2">Raikküla Stage</td><td>Raikküla Stage G_3</td><td>Raikküla Stage</td><td>Raikküla Stage</td></tr>
<tr><td rowspan="2">Borealis-bank und Jörden'sche Schicht</td><td>Borealis bank G_2</td><td>Borealis Bank</td><td>Borealis Stage G_2</td><td colspan="2">Borealis Stage</td><td>Borealis Stage G_2</td><td rowspan="2">Juuru Stage</td><td rowspan="2">Juuru Stage</td></tr>
<tr><td>Jördensche Schicht G_1</td><td>Jörden Zone</td><td>Juuru Stage G_1</td><td colspan="2">Juuru Stage</td><td>Juuru Stage G_1</td></tr>
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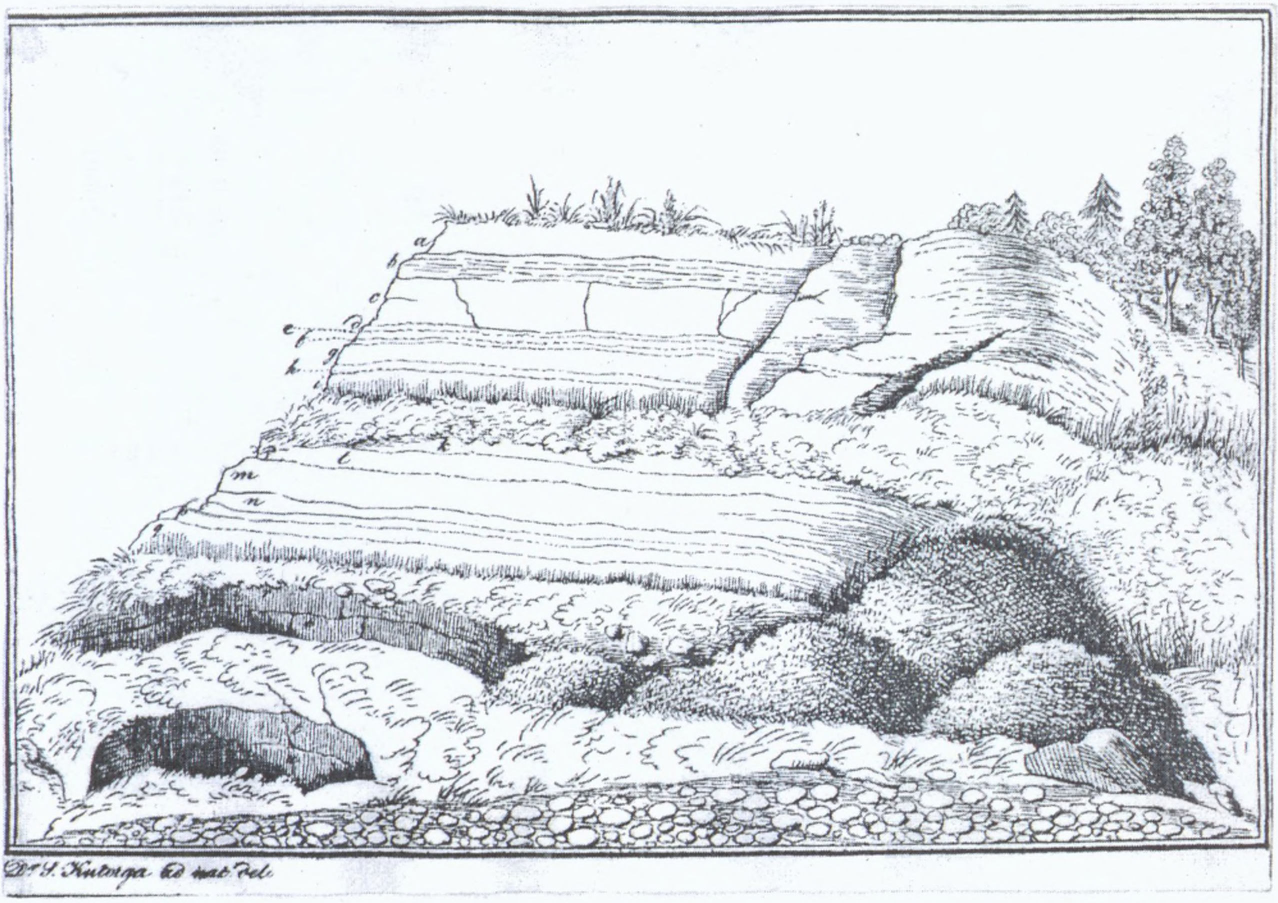


Fig. 4. The oldest sketch to depict the outcrop of Devonian sandstones of the Aruküla Stage at Jaama Street in Tartu. From Kutorga 1835.

sequently, Schmidt (1881, Шмидт 1879) introduced capital Latin letters combined with Arabic numbers (G_1, G_2, G_3, *etc.*) as symbols for this unit. He (Schmidt 1892) further refined his stratigraphic classification by distinguishing within the Upper Oesel Group Eurypterus beds and Ilionia (Didyma) beds which correspond to the Rootsiküla and Paadla stages of the current classification, respectively.

The Silurian fauna of Estonia received early attention, especially by Eichwald, in various papers. A contribution of great international importance was the monograph on Silurian (=Ordovician + Silurian) fishes by Pander (1856). The Silurian (*s.str.*) material was mostly derived from the Schmidt's Upper Oesel beds of Saaremaa. The description of Pander's new group Conodonta was partly based on specimens from Saaremaa.

The exquisitively preserved specimens of merostomes, fishes and some other uncommon groups from exposures at Viita in Rootsiküla on Saaremaa Island were of wide international interest. The exposures were located by Schrenk in 1852 and material from those beds formed the subject of many papers. The monographs by Rohon (1892, 1893) and Holm (1898) should be mentioned in the first place.

The Schmidt's stratigraphic terminology was somewhat inconsistent which made Twenhofel (1916) to propose more adequate terms for some units (Table 2). Bekker (1922) revised the terminology to accord modern maps and started field work with an aim of studying the stratigraphy of the Silurian sequence of Saaremaa in more detail (Table 2). On account of his untimely death, only an outline of his stratigraphical results became published (Bekker 1925). The study was continued by Artur Luha (1892-1953) who assisted Bekker in the field. Regrettably, only a condensed version (Luha 1930) of his voluminous manuscript was published.

In 1929, Luha discovered a locality with a rich agnathan fauna at Himmiste-Kuigu on Saaremaa. Later studies (Аалоэ 1963б) showed that the Himmiste beds were younger than previously believed (about middle of the Paadla Stage).

Luha (1933) improved the chronostratigraphic classification by restricting the term Jaani Stage to beds now known to correspond to the Lower Wenlockian (Table 2). Later Luha (1946) introduced the term Jaagarahu Stage for the unit comparable to the Upper Wenlockian. Teichert's (1928) study on the Lower Llandoverian in the western part of mainland Estonia and on the Island of Hiiumaa is also worthy of mentioning.

Studies of the Silurian resumed in the mid-1950s. Numerous borings made it possible to extend examination of lithofacial relationship also south of the outcrop area, and facilitated establishing of a detailed lithostratigraphic classification (Aaloe 1958, Аалоэ 1960, 1961, Аалоэ и Кальо 1962, Эйнасто 1962, *etc.*). Kaljo and Sarv (Кальо и Сарв 1966) specified equivalents to the Downtonian Series in the Estonian sequence. Examination of graptolites (Кальо 1967, Кальо и Вингисаар 1969) contributed to a more precise correlation of the sequence with other areas. As the Kaugatuma Stage proved to be a composite unit, Klaamann (Клааманн 1970a) distinguished its lower, Ludlovian part as a separate Kuressaare Stage. The discovery of Llandoverian K-bentonite beds in borings on Saaremaa and in the southwestern part of mainland Estonia by Jürgenson (Юргенсон 1958 a,1964) contributed to correlations. Studies of lithology of the Silurian rocks received increased attention (Юргенсон 1966, 1974, *etc*.).

Biostratigraphic correlations were greatly facilitated by monographic descriptions of several groups of microfossils such as ostracodes (Сарв 1968), agnathan scales (Märss 1996), chitinozoans (V. Nestor 1994) and conodonts (Männik 1992b).

Silurian macrofauna was described in numerous papers, some of those monographic in character. Various groups were covered, including stromatoporoids (Нестор 1964, 1966), tabulate corals (Соколов 1951б, 1952б, Клааманн 1961, 1962, 1964, 1966, 1970 б), bryozoans (Астрова 1970), articulate brachiopods (Рубель 1970, Rubel 1970), and trilobites (Männil 1982, 1992).

The impressive array of information from the Silurian of Estonia is systematised and discussed in several books, covering general features (Кальо 1970в), facies and fauna (Кальо 1977), communities and biozones (Кальо и Клааманн 1982), ecology (Kaljo & Klaamann 1982) and lithology (Юргенсон 1988).

History of Devonian research

Engelhardt and Ulprecht (1830) provided the earliest information of the red sandstones and mentioned also finds of vertebrate teeth and bone fragments. Kutorga (1835, 1837) described the exposures at Tartu (Fig. 4). Eichwald (1840a, Эйхвальд 1840) and Buch (1840) recognised very early that the terrigenous sequence was comparable to the Devonian Old Red rocks in Great Britain. Helmersen was the first to show the approximate outcrop area on a map. A similar map was provided by Murchison *et al.* (1845). The available information was summarised by Grewingk (1861, further specifications in 1878 and 1879, Photo 12).

Photo 12. Caspar Andreas Constantin Grewingk (1819-87) was the first to speak about several glaciations in Estonia. He studied Devonian outcrops and the Kunda Stone Age settlement, compiled a geological map of the Baltic States (published in 1861 and 1879) and laid the foundation to the collection of meteorites in Tartu University.

Quenstedt (1838) suggested that the teeth and bones recovered from the sandstone belong to ancient fish-like forms. The fairly substantial material of fish remains was subject to monographic studies by Asmuss (1856) and Pander (1856, 1860).

Based on the field studies of the Upper Devonian carbonate rocks of the southeastern part of the Republic, Bekker (1924a) attempted a chronostratigraphic classification of the sequence of which the term Dubniki Stage is still in use. In the current stratigraphic practice the other terms are based on the sections in the neighbouring areas of Latvia and Russia where carbonate rocks of this age have a wider distribution.

The lithostratigraphy of the basal Old Red lithofacies in the outcrop area was subject to a detailed study by Orviku (1930c, 1932, 1935b). Obruchev (Обручев 1933) distinguished two stratigraphic units, now regarded as the Pärnu and Narva stages. Subsequently, Orviku (1948) published an additional comprehensive paper on the stratigraphy of the Narva River Stage (=Narva Stage).

An important contribution to the knowledge of the Old Red sequence was the biostratigraphic classification, based on fishes from both Estonia and Latvia. The classification was developed by Gross (1933, 1940a, b; 1942, 1950) and later continued by E. Kurik (née Mark). Mark (Марк 1958) distinguished the Eifelian Aruküla Stage. The historically known Aruküla caves near Tartu (Photo 13) where, since 1831 during more than 20 years, H. Asmuss had excavated bones of large placoderms and other fishes, were selected for the type locality.

During several years between World War I and II, V. Paul (1934, 1939) excavated fossil fishes at Tamme near Lake Võrtsjärv and at Haaslava. The excavations, organized by E. Mark-Kurik, started in 1949 and lasted with a few intervals up to 1993 (more than 10 excavations were made at Karksi alone). Taphonomical study was provided during these excavations.

The overlying, lower Givetian unit was distinguished by Mark (Марк 1958) as the Burtnieki Stage with the stratotype in Latvia. The stratigraphy of the Devonian in Estonia was summarised by Mark and Paasikivi (Марк и Паасикиви 1960), including new information obtained from examination of cores from numerous deep borings. Special attention was further paid on studying the mineral composition of terrigenous rocks, largely on the basis of core material. The studies of the kind were initiated by H. Viiding (1929-88) and supplemented particularly by A. Kleesment (née Tamme). Numerous papers were published on the subject (Вийдинг 1962, 1964, 1965, 1976; Тамме 1962, 1964; Марк и Тамме 1964). Based on lithological and mineralogical criteria, Kleesment (Клеесмент 1966) and Kleesment *et al.* (Клеесмент и др. 1975) recognised in the borings of southeastern Estonia the presence of the Lower Devonian Tilžė and basal Middle Devonian Rēzekne stages, previously distinguished in Latvia and Lithuania. Mark-Kurik (1991a) has considered the Rēzekne Stage as a lower Devonian unit.

The Devonian stratigraphy in Estonia was subsequently

Photo 13. The historically known Aruküla caves near Tartu where since 1831 large placoderms and other Devonian fishes have been excavated. Here is the type locality of the Eifelian Aruküla Stage. *Photo by Ü. Heinsalu and E. Mark-Kurik.*

summarised in the papers published by various authors in the book "Devonian and Carboniferous of the Baltic" (Сорокин 1981). Recently, Kleesment (1994, 1995) published important contributions to detailed lithostratigraphy of the Middle Devonian sequence in Estonia. Mark-Kurik correlated Estonian Devonian units with those in other areas, *e.g.* in Scotland and Latvia (Марк-Курик 1981, Mark-Kurik 1991a,b, Курик и др. 1989). Valiukevičius (1994) gave an acanthodian zonation for the Baltic Basin (including Estonia). The fossil fishes of the Devonian sequence continued to receive attention. Beside the above-mentioned papers by Gross, the contributions by Heintz (1930, 1934), Mark (Марк 1953a, б, 1956, 1963), Mark-Kurik (1968, 1973, 1993a,b), Mark-Kurik *et al.* 1991, Obruchev and Mark-Kurik (Обручев и Марк-Курик 1965, Obruchev & Mark-Kurik 1968) deserve special mention. Mark-Kurik has also studied special problems of palaeoichtyology: palaeopatology (Марк-Курик 1966), functional morphology (1984) and trophic relations (1995). Rich collections in Estonia have provided an excellent basis for studying of several fossil fish groups: antiarchs by Karatajūtė-Talimaa (1960, Каратаюте-Талимаа 1963), arthrodires by Obrucheva (Обручева 1962, 1966), acanthodians by Valiukevičius (Валюкявичюс 1985) and crossopterygians by Paul (1940), Vorobyeva (Воробьева 1977). Thomson (1940) pioneered in studying the Devonian plant remains (Photo 14). Later Vaitiekūnienė (Клеесмент и др. 1975) and Kalamees (1988) studied spores and macroremains (including phytoleimma), respectively. A number of papers contain descriptions of the Devonian invertebrates from Estonia: ostracodes (Öpik 1935a, Поленова 1966), lingulates (Батрукова 1960, 1964, 1969, Гравитис 1981) and conchostracans (Миронова 1969).

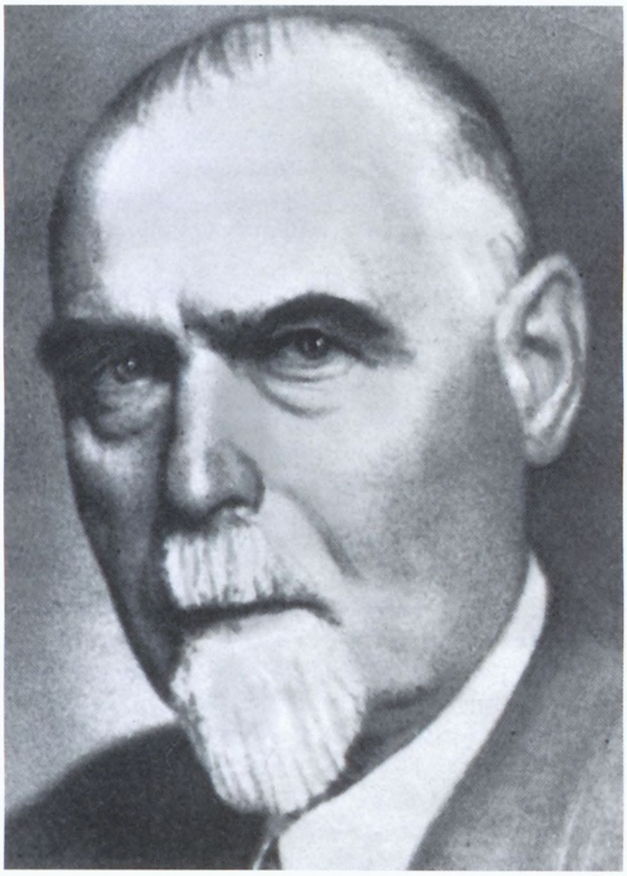


Photo 14. Paul William Thomson (1892-1957) pioneered in studying the Devonian plant remains in Estonia and Eemian interglacial deposits at Rõngu. Based on pollen analysis, he proposed the first biostratigraphical schemes of the Holocene and Late-glacial sediments in the area.

History of Quaternary research

The first stratigraphical scheme of the Quaternary deposits was compiled by Schrenk (1854). Based on the then prevailing drift theory he, like Schmidt (1854) and Grewingk (1861), divided all Quaternary sediments into diluvial and alluvial sediments with several lithological varieties.

Estonia was among the first regions where the theory of continental glaciation was applied. Eichwald (1853) was the first in the Baltic provinces to consider the possibility that at least northern Estonia had once been covered by an ancient active glacier. Already in 1865, Schmidt (1865) clearly spoke about glacial sediments, and a bit later he (Schmidt 1869) described glacial and postglacial formations and differentiated four stages in the development of the territory, including the time of the invasion of big glaciers, the time of the melting of the glaciers, and the time of the final melting of the ice with a wide distribution of fresh-water lakes. Schmidt (1858) was the first who found the shells of *Ancylus fluviatilis* in beach deposits on Saaremaa Island and distinguished a fresh-water stage in the Holocene history of the Baltic Sea.

Helmersen (1869) explained the distribution of erratic boulders and formation of boulder clays and ice scratches with the joint action of the continental ice sheet, floating icebergs and erosional processes on land. Schmidt (1871) proved that in the glacial epoch a unitary glacier moved from Scandinavia over the Baltic Sea depression and Estonia. In 1879, Grewingk already spoke about several glaciations (Grewingk 1879), basing on the study of the different till beds in Tartu.

However, it was not until half a century later that Grewingk's statement was confirmed by palaeontological data. In 1939, Orviku performed the first detailed studies on interglacial organogenous deposits at Rõngu (Orviku 1939) which, according to pollen zones (Thomson 1939a, 1941), were correlated with the typical Riss-Würmian (Eemian) interglacial deposits in Western Europe. By now, tills of five glaciations or big stadials have been identified in Estonia (Раукас 1978) and both Eemian and Holsteinian interglacial sediments described in detail (Liivrand 1991). Official stratigraphical schemes of the Quaternary (Raukas & Kajak 1995), Lateglacial (Pirrus & Raukas 1996) and Holocene (Raukas *et al.* 1995b) have been accepted and published.

Monographic studies of ice-marginal formations (Раукас и др. 1971), bedrock topography (Таваст и Раукас 1982), lithology of Quaternary deposits (Раукас 1978), modern (Орвику Каарел 1974) and ancient (Кессел и Раукас 1967) coastal formations have been published. The overviews about the glacial history (Raukas 1995a) and the history of the development of geomorphology in Estonia (Raukas & Karukäpp 1993), imparting more information about the study history of different landforms and types of sediments, appeared recently in print.

The authors of the chapter would like to thank Prof. Valdar Jaanusson for his numerous valuable comments.

III PRECAMBRIAN BASEMENT

Introduction

The early Precambrian crystalline rocks are covered by the Upper Vendian and Palaeozoic sedimentary rocks. The basic data for studies of the Precambrian has been obtained by means of boreholes and geophysical survey. Of about 500 boreholes passing right through the sedimentary cover, the deepest ones penetrate into the basement to a depth of up to 450 m.

The Precambrian basement in Estonia consists of two megaunits: the orogenic Svecofennian complex of metamorphic and plutonic rocks and the anorogenic complex of plutonic rapakivi granites and related rocks. Earlier views about the age of basement rocks (Пуура и др. 1976, 1983) have considerably changed during the last decade due to the isotopic dating of the South-Estonian granulitic crust (Puura & Huhma 1993) and rapakivi granites of Fennoscandia (Rämö *et al*. 1996). The new standpoints, based on the results of these studies, underlie the recent joint publications on the Precambrian of the Gulf of Finland and surrounding area (Koistinen 1994, 1996; Laitakari 1996). Based on recent results, a stratigraphic chart of Precambrian rocks of Estonia was compiled (Table 3).

PALAEOPROTEROZOIC

Svecofennian orogenic metamorphic rocks

According to the degree of metamorphism and the composition of the metamorphic sequences reflected in the geophysical patterns, Estonia's basement is divided into structural regions (Figs. 5, 6) which differ from each other in the volume of sedimentary and felsic to mafic volcanic rocks.

Comparison of the basement-forming rocks in Estonia, Finland and Sweden has shown that the metamorphosed volcanic and sedimentary rocks in Estonia's basement have many features in common with the rocks in the Svecofennian orogenic complexes. In the first instance, it was established that the rocks of the Tallinn and Alutaguse structural zones and the Svecofennian complex are similar in lithology and

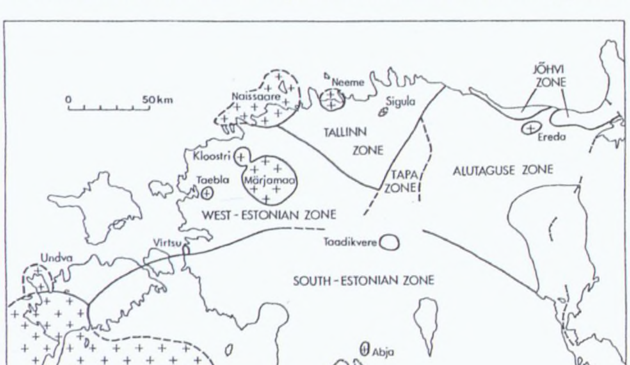


Fig. 5. Relations between two megaunits of the Precambrian basement of Estonia: 1 - Svecofennian orogenic complex (Palaeoproterozoic) with the names of the structural zones and postorogenic massives; 2 - anorogenic complex of plutonic rapakivi granites and related rocks with the names of the massives.

have the same stage of metamorphism (Пуура и др. 1983, Клейн 1986). The orogenic supracrustal rocks of southern and western Estonia differ from the bulk of Svecofennian rocks of Scandinavia by their prevailingly mafic to intermediate composition and high-grade metamorphism. This is in good correlation with high anomalies of gravity and magnetic fields (Пуура и др. 1976, 1983, Коппельмаа и др. 1978). The Palaeoproterozoic age of the granulite complex of Estonia was dated by Sm-Nb isotopic studies (Puura & Huhma 1993). Petrological signatures of mafic rocks in southern and western Estonia are concordant with those in the northern and northeastern parts of the territory.

In the **Alutaguse Zone** (Fig. 5) gneisses containing biotite, cordierite, garnet and sillimanite intercalate with biotite gneisses and form a complex of the same name (Пуура и др. 1976). Within the Alutaguse Zone, in the area of Uljaste, Haljala and Assamalla, the basement comprises sulphidic black schists, quartzites, amphibole and pyroxene gneisses, marbles and pyroxene skarns.The rocks in the Alutaguse Zone derive from clastic successions with minor sequences of volcanic, sandy and carbonate rocks in the above-mentioned areas. The local uplifts of the basement in the Uljaste and Assamalla area consist primarily of quartzites.

The Al-rich, sillimanite-garnet-cordierite gneisses are, for the most part, medium-grained, banded and migmatized by plagioclase-microcline granite or pegmatite. The mineral composition of gneisses varies. Light minerals are represented by quartz, plagioclase (An25-55) and microcline; dark minerals by biotite, cordierite, garnet and sillimanite. Muscovite is rare, while andalusite is occasional and rare. Chemically, the sillimanite- garnet- cordierite gneisses are similar to pelites and originate, in all likelihood, from psammitic to pelitic sediments. Microgneisses, rich in quartz, form interlayers with the highest sand content in these sediments.

The **Tallinn Zone** is characterised by the stratified Jägala Complex of intercalating sillimanite-cordierite and biotite gneisses, intermediate to mafic metavolcanics, and leucocrate gneisses. In the WNW-ESE-trending zones, the primarily psammitic to pelitic metasediments intercalate with metavolcanites. The acidic metavolcanites alternate with more abundant intermediate to mafic metavolcanites.

The intermediate metavolcanic rocks are represented by metamorphosed to biotite-hornblende and biotite gneisses which are fine- to medium-grained and migmatized by microcline- plagioclase granites. The main minerals of biotite gneisses are plagioclase (An 30-45), quartz, biotite and, in places, microcline. The basic minerals of biotite-hornblende gneisses, which are slightly more mafic in composition, are plagioclase (An 35-50), quartz, hornblende and biotite. These gneisses are andesitic (SiO_2 55-63%, $Na_2O + K_2 = 4.5 - 6.5\%$).

Typical felsic quartz-feldspar gneisses are fine-grained, rather massive or schistose granoblastic rocks. In single boreholes, rocks with relicts of blasto-porphyritic texture (phenocrysts of quartz and plagioclase), indicative of their volcanic origin, have been found. Quartz (25-40%), plagioclase (An 20-40) and potassium feldspar form 85-95% of

Table 3. Age of Precambrian formations in Estonia and surrounding area

Mesoproterozoic **1000-1600 Ma**		
~1200 Ma	Post-Jotnian dolerites	
~1400 Ma	Jotnian sandstones	
1540-1580 Ma **Rapakivi granites and related rocks of the Åland-Riga Subprovince:**		
volcanics (Undva Member): quartz porphyry plagioclass porphyrite		plutonic rocks: rapakivi granite anorthosite and other mafic plutonics
Palaeoproterozoic **1600-2500 Ma**		
1620-1670 Ma **Rapakivi granites and related rocks of the Vyborg Subprovince:**		
volcanics (Hoglandian): quartz porphyry plagioclase porphyrite		plutonic rocks: rapakivi granite porphyry granite anorthosite and other mafic plutonics ofite gabbro and dolerite
1800-1900 Ma **Svecofennian orogenic rocks**		
		1800-1840 Ma (?): postorogenic plutonic rocks: quartz monzonite
		~1800-1860 Ma late orogenic plutonic rocks: diabase, microcline granite and granite-migmatite, charnockite (?), granodiorite(?)
1880-1900 Ma metasediments and metavolcanics		~1880-1900 Ma synorogenic plutonic rocks: metagabbroids, and -ultramafics, granodiorite, quartz diorite and charnockite
Jägala Complex: felsic, intermediate and mafic metavolcanics, metapelites and metagreywackes		
Alutaguse Complex: metapelites, metagreywackes, quartzites, carbonates, mafic metavolcanics		
Vaivara Complex: mafic and intermediate metavolcanics, metagreywackes and metapelites, banded iron formation		
Metamorphic rocks of South and West Estonia and Tapa Zone: mafic, intermediate and felsic metavolcanics, metasediments		

quartz-feldspar or granite gneisses. In the chemical composition (SiO_2 65-76%, Na_2O + K_2O = 5.5 - 8%) the quartz-feldspar and granite gneisses are similar to acidic volcanites (dacites, rhyolites) and arcosic sandstones.

Petrographically, the Al-rich gneisses of the **Jägala Complex** are similar to those in the Alutaguse Zone.

The basement in the **West-Estonian Zone** consists predominantly of the same assemblage of rocks as in the Tallinn Zone, although the stratified structure is not so well reflected in geophysical anomalies. The rocks, characteristic of this zone, are rather uniformly medium- and fine-grained biotite-hornblende gneisses and amphibolites, which have been migmatized by microcline granites. The amphibolites mostly occur as layered bodies and are intercalated with gneisses. In the amphibolites, plagioclase (An 35-55) and hornblende are the main minerals, however, they may also contain biotite, clinopyroxene and quartz. The main minerals of the biotite-hornblende gneisses are plagioclase (An 30-50), quartz, hornblende and biotite, rarely potassium feldspar. There are also gneisses, the mafic parts of which consist entirely of biotite. North of Haapasalu, on the Noarootsi Peninsula, the gneisses also contain hypersthene. According to the chemical composition, the amphibolites (SiO_2 45-53 %, Na_2O +K_2O = 3 - 4.5%) are referred to basalt, and the gneisses (SiO_2 55-63, Na_2O + K2O = 5-6.5%) to andesite.

In the **Tapa Zone**, a rock association, analogous to that of the West-Estonian Zone (amphibolites, biotite-hornblende gneisses, in places pyroxene gneisses) occurs. The **Jõhvi Zone** (Fig. 5) is composed of the rocks of the Vaivara Complex. Magnetite quartzites occur in a limited area together with Al-rich and pyroxene gneisses. The latter contain interlayers of quartz-feldspar and biotite-amphibole gneisses. The fine- and medium-grained pyroxene gneisses with a variable mineral composition display charnockitic and granitic migmatization. Orthopyroxene and biotite are always

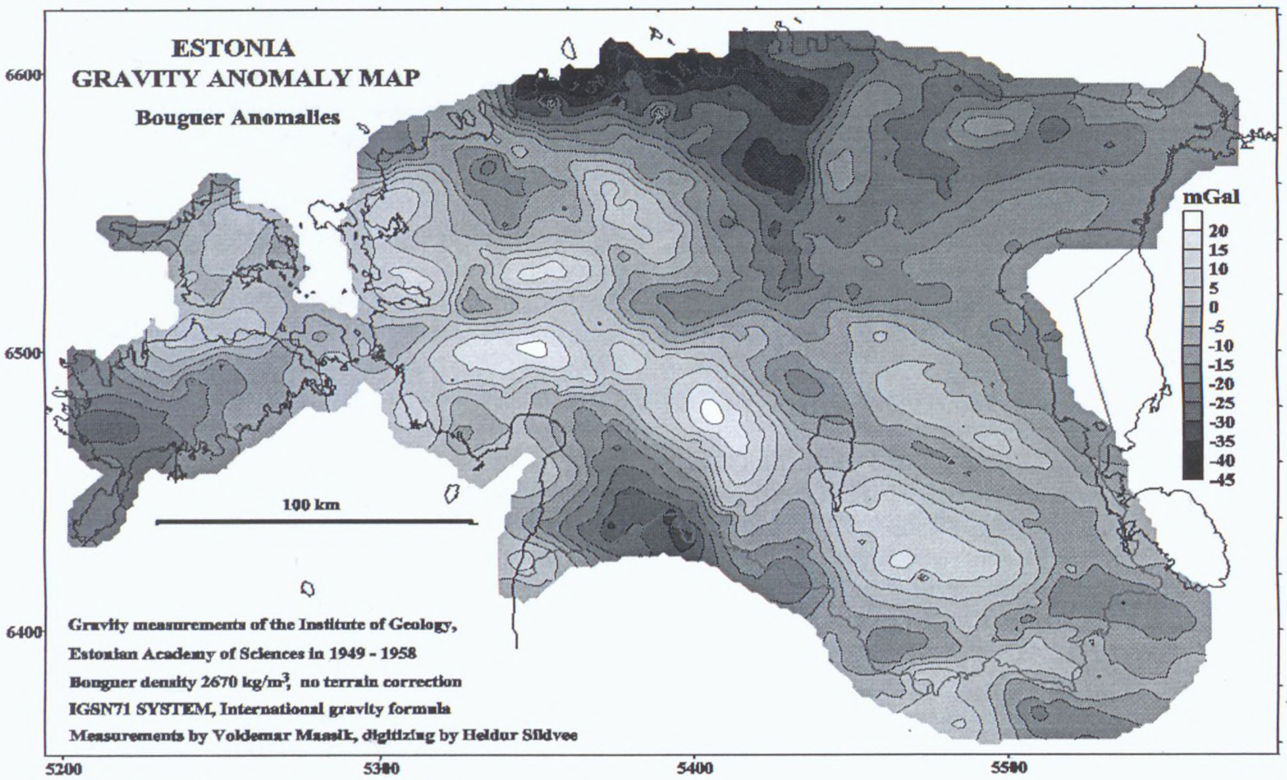


Fig. 6. The gravity anomaly map of Estonia.

present. The content of clinopyroxene and hornblende varies from 0 to 25%. Of light minerals, plagioclase (An 40-55%) predominates, while quartz and potassium feldspar are often absent. Within the Jõhvi magnetic anomaly area, rocks of almost ultramafic composition comprising orthopyroxene, clinopyroxene, hornblende, biotite and plagioclase are occasionally encountered (5-10%). Biotite-hypersthene gneisses containing plagioclase and quartz are also widespread. Chemically, the gneisses correspond to andesite.

The magnetite quartzites, fine-grained banded rocks in the Jõhvi area, contain besides quartz and magnetite, also garnet, orthopyroxene, clinopyroxene, hornblende, cummingtonite and biotite in different quantities. The average content (by microsections) of quartz is 30-40%, with the proportion of magnetite reaching 25-30%. The magnetite quartzites are cut by veins of pegmatoid microcline granite.

In the **South-Estonian Zone**, the metamorphic complex consists of hypersthene, clinopyroxene and amphibole gneisses, originating from mafic to intermediate volcanites, and possibly from greywackes. It also contains Al-rich and minor members of felsic gneisses.

Different fine- to medium-grained pyroxene gneisses, which have undergone charnockitic and granitic migmatization, are characteristic of southern Estonia. The primary structures of these gneisses have been obscured or obliterated. The characteristic mineral assemblage of the hornblende-pyroxene gneisses is orthopyroxene + clinopyroxene + hornblende + biotite + plagioclase +/- potassium feldspar +/-quartz. The plagioclase is mostly antiperthitic, mainly andesine-labradorite, rarely bytownite. The potassium feldspar is orthoclase-microperthite. Quartz is rare. The hornblende-pyroxene gneisses have been found mostly in boreholes in the surroundings of Pärnu and Viljandi where they occur as interlayers in acidic gneisses. The chemical composition of the amphibole-pyroxene gneisses corresponds to basalt or basaltic andesite (SiO_2 47-54 %, Na_2O+K_2O = 3-5 %), but the content of iron, magnesium and calcium differs noticeably. The increased content of magnetite (3-4%) is a specific feature of these gneisses. The essential minerals of the biotite-hypersthene gneisses are plagioclase (mainly An 35-45%, in some cases An 70-80%), hypersthene, biotite and quite often quartz and potassium feldspar. Gneisses of this type occur typically in the vicinity of Tartu, Otepää and Laeva. Among the biotite-hypersthene gneisses, both melanocratic and leucocratic varieties occur (SiO_2 48-60%). Compared to the hornblende-pyroxene gneisses, the biotite-hypersthene gneisses are generally poorer in calcium, but richer in potassium and magnesium, which is evidently due to the weathering of the source rock and mixing with pelitic matter. In the rather rare quartz-feldspar gneisses of southern Estonia, garnet or hypersthene and hornblende occur as accessory minerals.

The gneisses, formed at granulite facies in the South-Estonian and Jõhvi zones, contain hypersthene and accessory spinel, garnet and cordierite porphyroblasts (also in granitic veins) and sillimanite, the latter occurring as inclusions in cordierite. The biotite gneisses occur together with sillimanite-cordierite gneisses. They are medium- to fine-grained, often foliated migmatitic rocks, the main minerals being quartz, andesine, biotite and potassium feldspar. Garnet, cordierite, sillimanite and muscovite occur in small quantities. The content of dark minerals averages 20-25%. Compared with other gneisses, biotite gneisses have the highest content of quartz.

Svecofennian orogenic plutonic rocks

Traditionally, the granitoid rocks of southern Finland have been classified into four groups, based on their relationship to orogenic movements (Koistinen 1994, 1996). These groups are synorogenic (synkinematic), late-orogenic (late-kinematic), post-orogenic and anorogenic rocks (rapakivi granites). Practically, this classification is often expanded on the whole variety of plutonic rocks.

Compared to Finland, the Estonian basement is rather poor in orogenic plutonic rocks. Synorogenic granitoid complex and associated mafics are rare in Estonia. The late-orogenic potassium granites which form the extensive W-E-striking belt in southern Finland, occur as small bodies and migmatite veins in the basement of northern Estonia.

In Finland, the granite migmatites fall into two distinct age groups which are related to early (1.9-1.87 Ga) and late orogenic (1.84-1.83 Ga, Koistinen 1996) granitoids. Like in southern Finland, where the both age groups of migmatites occur in the same metamorphic complex around the Potassium Granite Belt (Koistinen 1996), age classification of migmatites in Estonia's basement is extremely complicated and, therefore, they are treated as one orogenic group. Occasionally, the classification of other plutonic rocks into the early and late orogenic groups is possible. Small bodies of synorogenic gabbronorite and gabbro, or metagabbro, cut by granite veins, occur in the Tapa Zone. The mafic rocks contain abundant hornblende and biotite of later origin. Similar rocks in northeastern Estonia form the considerably large Pada Pluton, which contains also diorite.

Drilling in southern and western Estonia has revealed some mafic, probably synorogenic rocks. There are small gabbronorite plutons including Võru, Laeva, Pärnu, Vanaküla and several others, some of those with structural orientation. Gabbro-norite is a massive rock of coarse or medium grain size, which contains plagioclase (An 50-70), ortho- and clinopyroxene, hornblende and biotite of secondary origin. In northern Estonia, small granodiorite and quartz-diorite bodies have been found in single boreholes (Aruküla, Letipea). The quartz-diorite is orientated, medium-grained and cut by veins of microcline granite. The small Utria body in the northeastern coastal area consists of massive medium-grained gabbrodiorite.

The granite rocks of the Estonian crystalline basement are mainly migmatite granites: plagioclase-microcline granites in northern and western Estonia, and charnockites and plagioclase-orthoclase granites in the granulite facies area of southern Estonia. The charnockites consist of potassium feldspar, plagioclase (andesine) and, to a lesser extent, of biotite, hypersthene and hornblende, the latter three forming 5-10 % of the rock. Quartz, microcline, oligoclase-andesine and biotite are the main minerals in the migmatite granites of northern and western Estonia.

Svecofennian post-orogenic plutonic rocks

In southern Finland, the 1.82-1.78 Ga tonalitic to monzonitic and granitic post-orogenic intrusions are neither voluminous nor numerous (Koistinen 1996). However, they mark the final stage of the Svecofennian orogeny when the temperature of the crust was still high. Recently, a group of post-orogenic granites was identified in the Estonian basement as well (M. Niin, unpublished report).

The Taadikvere body in Central Estonia consists of granodioritic - quartz-monzonitic rocks which are of preferred orientation and contain plagioclase and potassium feldspar phenocrysts. The medium-grained groundmass of the rock comprises quartz, plagioclase (An 32-36), potassium feldspar, biotite and hornblende.

The Virtsu body in western Estonia, consists of porphyritic rocks of quartz monzonite composition that are strongly crushed within the west-east oriented central Estonian cataclastic zone. The medium-grained groundmass of the rock consists of plagioclase (An 31-42), potassium feldspar, quartz and biotite with some admixture of hornblende, and numerous potassium feldspar and plagioclase phenocrysts, up to 2-3 cm in diameter.

Palaeoproterozoic metamorphism

The mineral paragenesis of metamorphic rocks and, partly, of early orogenic plutonic rocks is due to regional metamorphism which in the Svecofennian orogen in Finland occurred in several stages during 1.885-1.81 Ga as dated by isotopic studies (Koistinen 1996).

Metamorphic zoning (Fig. 7) is typical of the Svecofennian orogen. In the province as a whole, andalusite-muscovite mica schists prograde into potassium feldspar-sillimanite gneisses and migmatitic garnet-cordierite gneisses. The neosomes in the migmatites differ markedly in type. In central Finland the granitoid area is surrounded by tonalitic and trondhjemitic migmatites, while potassium granitic neosomes occur in migmatites of southern Finland (Koistinen 1996). Potassium granitic migmatites extend into the North-Estonian zones of amphibolite metamorphism. In the southern part of Estonia, charnockite and enderbite migmatites characteristic for granulite facies area have been described.

The Precambrian basement of Estonia consists of rocks which have been subjected to high-grade metamorphism. In northern Estonia (Figs. 8, 9), amphibolite facies gneisses are most abundant, while granulite facies mineral assemblages occur locally as in the Jõhvi and Tapa zones. Assemblages marking transition from amphibolite to granulite facies occur in the vicinity of Uljaste and Assamalla (Клейн 1986). The amphibolite facies gneisses in Estonia serve as an extension to those spread in southern Finland. The dominant metamorphic grade here is a high- temperature amphibo-

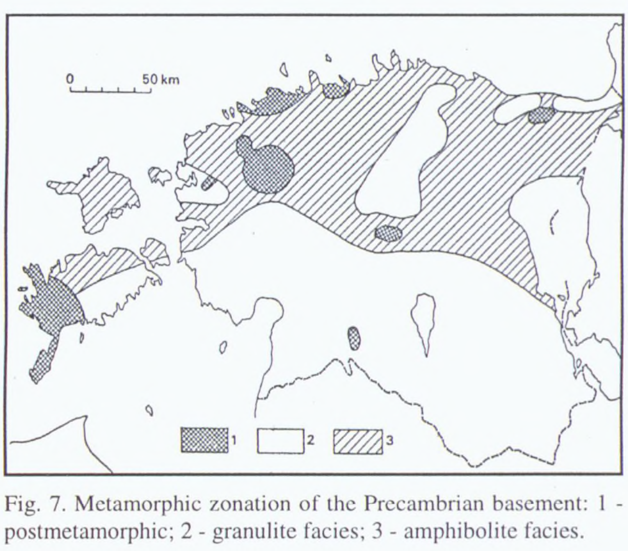


Fig. 7. Metamorphic zonation of the Precambrian basement: 1 - postmetamorphic; 2 - granulite facies; 3 - amphibolite facies.

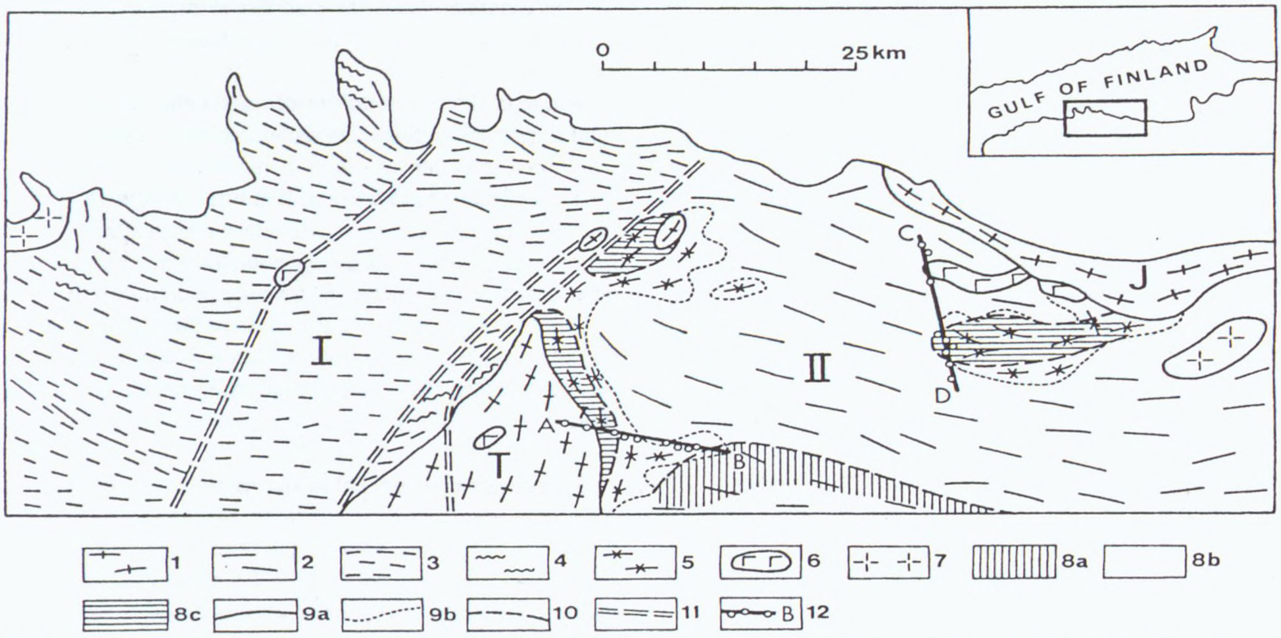


Fig. 8. Zonation of the regional metamorphism in northern Estonia: I - Tallinn Zone; 2 - Alutaguse Zone, T - Tapa and J - Jõhvi blocks. 1 - mafic granulite complexes (Py- and Amph-rich gneisses); 2 - Al-rich (Sill-, Gr-, Cord- and Bi-bearing) gneisses; 3 - acid, intermediate and mafic metavolcanites; 4 - graphite- and sulphide-bearing aluminiferous and micaceous gneisses; 5 - graphite- and sulphide-bearing gneisses (black schists), quartzites, marbles, pyroxene-rocks; 6 - gabbro and gabbronorite; 7 - rapakivi granite; 8 - low/medium (a) and upper (b) amphibolite facies and granulite (c) facies; 9 - boundaries of structural zones (a) and subzones (b); 10 - boundaries between subfacies; 11 - faults; 12 - cross-sections (AB - Assamalla, CD - Uljaste, Fig. 9).

lite facies with a local PT zoning from sillimanite-potassium feldspar subfacies to granulite facies (Uljaste, Haljala). Geothermobarometry, mostly of the biotite + garnet + /- sillimanite assemblage and of cordierite, estimates prograde metamorphism at 600-700° C and 3-5 kbar.

In the Jõhvi Zone, there are characteristic granulite mineral assamblages in cordierite-garnet gneisses (hypersthene) and in mafic gneisses (two pyroxenes and spinel). In the Tapa Zone, the traces of granulite metamorphism have probably been partly removed by high-temperature retrograde metamorphism.

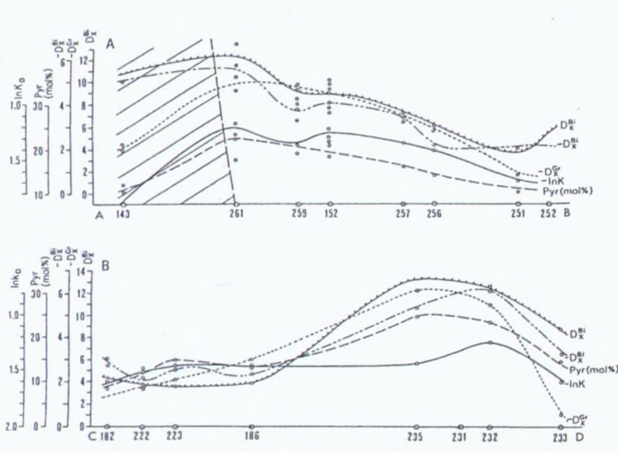


Fig. 9. Metamorphic parameters of Assamalla (A) and Uljaste (B) areas (for location of AB and CD profiles see Fig. 7). 143, 261..., *etc.* - numbers of drill cores. lnK according to Glebovitsky and Drugova (Глебовицкий и Другова 1979). -DxGr = 8.2 Mn - 8.56Mg - 0.37Ca and -DxBi = 19.56Ti + 1.26Al - 3.89Mg according to Nikitina and Drugova (Никитина и Другова 1977), DxBi = 7.876Si - 10.251Al + 17.173Ti - 5.661Fe_3 + - 1.404 (OH) - 4.286Fe_2+ + 4.524Mg + 4.661K according to Ushakova (Ушакова 1971).

In southern Estonia, granulite facies gneisses form a large domain (Fig. 7), which extends from the Middle-Estonian fault zone (Saaremaa-Peipsi Zone) to northern Latvia and further south through the beltiform Belarussian-Baltic granulite domain. The conditions of metamorphism have been mainly studied from two drillcores - Kõnnu 300 and Varbla 502 (Коппельмаа и др. 1978, Hölttä & Klein 1991). The mineralogy of granulites varies. The widespread garnet and cordierite, formed by breakdown of biotite and sillimanite, indicate prograde metamorphism. Hypersthene coexists with garnet and cordierite although, so far, the sillimanite-hypersthene assemblage has not been found. The PT-conditions have been calculated using several geothermometers and geobarometers (Hölltä & Klein 1991). These give temperature estimates for the prograde stage of metamorphism of 700-800°C and pressure estimates of 5-6 kbar or more.

The age relations between the described zones are still ambiguous. The southern Estonian granulite area has been correlated with those in Finland. In the Haukivesi-Kiuruvesi Complex metamorphism has been dated at 1.88 Ga (Korsman *et al.* 1984). High-temperature metamorphism in southern Finland, 1.83-1.81 Ga ago (Korsman *et al.* 1984), may be correlated with similar metamorphism in northern Estonia.

It is emphasized that the southern Estonian granulites (Fig. 10) formed under higher pressure than is characteristic of Svecofennian metamorphism. This suggests that the South-Estonian region represents a deeper crustal section (Koistinen 1996).

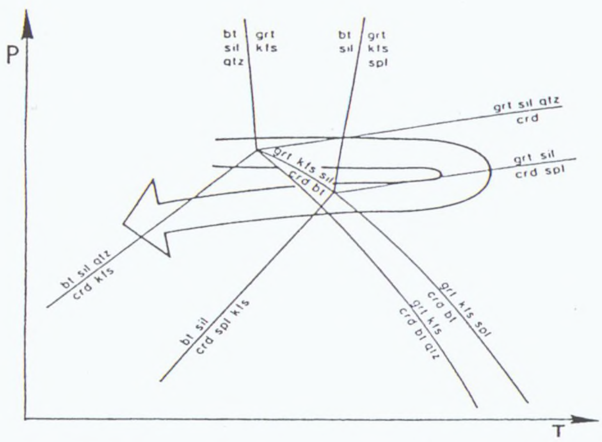


Fig. 10. A schematic locus of PT development of southern Estonian granulites (Höltta & Klein 1991).

PALAEOPROTEROZOIC TO MESOPROTEROZOIC – RAPAKIVI AND RELATED ROCKS OF THE FENNOSCANDIAN PROVINCE

A large time span for the Fennoscandian rapakivi and related rocks' plutonism at 1.65-1.54 Ga was established by isotopic studies (Rämö *et al.* 1996). Recently, it was stated that the province consists of four subprovinces separated areally and differing in age (Puura & Flodén 1996). The large plutons have a central position in the subprovinces, while the stocks and dike swarms occur in peripheries of the subprovinces. Volcanic rocks have survived as remnants in the vicinity of the main plutons. In the basement of Estonia, plutonic and related rocks of two, Vyborg and Riga-Åland subprovinces occur (Table 4).

Palaeoproterozoic rocks of the Vyborg Subprovince, the 1.62-1.67 Ga age group

The Vyborg Pluton in southeastern Finland and adjacent offshore area has a central position in the oldest rapakivi subprovince. Its southern satellite spreads in the bottom of the Gulf of Finland, near the northeast coast of Estonia (Koistinen 1994).

Characteristic structures of this subprovince in Estonia are stocks of porphyritic potassium granites, chemically only little differing from the proper rapakivi (Куусппалу 1975, Puura & Flodén 1996). Presuming that also the westernmost and smallest known but undated Taebla Stock (Fig. 5) belongs to the Vyborg Subprovince, then practically the whole mainland Estonia was influenced by rapakivi magmatism at about 1.65-1.64 Ga.

The **Naissaar** Pluton (55km x 25 km), the northern part of which extends under the Gulf of Finland, is composed of porphyritic granites cut by aplites. According to the chemical and mineral composition, this pluton is divided into two phases (Soesoo & Niin 1992). The more melanocratic granites of the first phase form the periphery of the pluton. The central part of the pluton is composed of leucocratic granites of the second phase, which have some similarities with the second and third phases of the Märjamaa Stock. The structure of the rocks is massive, in places (second phase) trachytoidic.

There are two generations of quartz (25-35%): crystals within microcline phenocrysts and anhedral crystals in the groundmass. In places, tabular microcline (35-45%) contains inclusions of plagioclase. Euhedral to subhedral zoned plagioclase (20-25%) is of oligoclase composition. Biotite forms small flakes containing euhedral crystals of zircon and apatite as inclusions. Muscovite and fluorite of post-magmatic origin replace plagioclase in the second phase of the intrusion. Hornblende occurs sporadically in granites of the first phase. The other accessory minerals are apatite, titanite, zircon and epidote. The opaque minerals are represented by magnetite and ilmenite.

The **Neeme** Pluton (25km x 20 km), the northern part of which is under the Gulf of Finland, is composed of coarse- and medium-grained pinkish-grey porphyritic granites cut by aplites (Soesoo & Niin 1992). By chemical and mineral composition, the rocks form two groups, possibly two phases. Two small bodies in the central and northeastern parts of the intrusion are more melanocratic, their chemical composition being partially similar to granodiorite. Partially assimilated xenoliths of surrounding gneisses with a diameter of about 20-30 cm have been found in some drill cores. The structure of rocks is massive, in places trachytoidic.

Quartz (20-25%) is generally euhedral, smaller grains are anhedral. Microcline (35-45%) occurs in groundmass in the form of rare phenocryst, up to 3-5 cm in diameter. Plagioclase (15-25%) is represented by euhedral crystals of oligoclase-andesine composition. Biotite (2-10%) forms unhedral flakes that contain small crystals of apatite, titanite, fluorite and zircon as inclusions. Hornblende occurs sporadically. Muscovite, apatite, fluorite, titanite, zircon, epidote and opaques are accessory minerals.

The **Ereda** Pluton (5km x 15 km) is composed of homogeneous pinkish-grey coarse-grained porphyritic granites (Soesoo & Niin 1992). As there are only two drill cores available, it is difficult to correlate the Ereda rocks with those of other stocks. The mineral composition and structure of the Ereda granites and the Märjamaa and Neeme leucocratic type of granites have some similar features.

Two generations of quartz (30-35%) have been distinguished. Microcline (35-45%) is present as tabular crystals; large phenocrysts with a dimater of 3-4 cm are zoned. Plagioclase (20-30%) forms various euhedral, tabular and prismatic crystals of andesine composition. Biotite (5-10%) is altered. The accessory minerals are fluorite, apatite, zircon, epidote and rutile. Magnetite and hematite opaques occur.

The **Märjamaa** Pluton (40km x 25 km) is composed of coarse-grained pink-grey porphyritic granitoids, sometimes cut by aplites (Soesoo & Niin 1992). According to the geophysical and drilling data, the pluton has features of ring structure and is accompanied by a smaller satellite. The contacts between the granites and surrounding Palaeoproterozoic gneisses are sharp. The round central part of the Märjamaa composite stock is represented by the most melanocratic and basic type of porphyritic granodiorites (first phase). In places it contains gneiss xenoliths, up to 20 cm in diameter. The second, intrusive phase, is represented by biotite and hornblende-bearing granites. The granites of the third phase (pos-

Table 4. The structures of the Fennoscandian Late Palaeoproterozoic to Early Mesoproterozoic rapakivi-anorthosite province

Name of subprovince, magmatic body*	Isotopic age, Ma	Size: area (km²) or extent (km)	Petrography** Felsic rocks	Mafic rocks	Location
VYBORG Subprovince	1615-1665				Finland, Estonia Russia
Complex pluton:					
Vyborg	1615-1646	21,100 km²	G, PG, EG, QP	GA, MV	Finland, Gulf of Finland
***Satellite*:**					
Gulf of Finland (GF)		2,730 km²	probably as Vy		Gulf of Finland
Adjacent body:					
Suursaari		33 km²	QP	MV	Gulf of Finland
Small bodies (stocks):					
Ereda (ER)		70 km²	Gs		Estonia
Neeme (NE)		127 km²	Gs		Estonia
Naissaar (NA)	1624	985 km²	Gs		Estonia
Märjamaa (MÄ)	1629	735 km²	Gs		Estonia
Kloostri (KL)		85 km²	Gs		Estonia
Taebla (TA)		35 km²	Gs	Gb	Estonia
Abja (AB)	1622-1635	40 km²	Gs		Estonia
Dike:					
Sigula (SI)		4 km	Gb		Estonia
RIGA-Åland Subprovince	1540-1585				Latvia, Estonia, Finland, Sweden, Baltic Sea
Complex Pluton:					
Riga (RI)	1584	41,300 km²	G, PG, EG, QP		Latvia, Gulf of Riga
	1576			GA, Gb, UM	Saaremaa, Ruhnu Island
Volcanics:					
Undva (UN)		>100 km²	QP	MV	Saaremaa

*Indices correspond to those in Chapter VII

**Type of rocks (see below)

I. Rocks, typical for large magmatic bodies and their satellites:
Felsic plutonic rocks, typically coarse-grained:
granites with rapakivi structure .... G
porphyritic granitoids .... PG
evengrained granitoids (tirilitic, syenitic, biotite granite *etc.* varieties) ... EG
Felsic subvolcanics:
granite porphyrites and granophyres as apical part of plutons .... GP
Felsic volcanics:
quartz porphyries .... QP
Mafic plutonic rocks:
gabbro-anorthosites .... GA
gabbro-norites, leucogabbros, *etc.* .... Gb
ultramafics (peridotites, *etc.*) .... UM
Mafic volcanics: .... MV
II. Rocks, typical for stocks
Medium- to coarse-grained porphyritic granitoids
(also granodioritic and syenitic varieties) .... Gs
III. Rocks, typical for dike swarms
Granite porphyres .... GP
Diabases (dolerites) .... D

sibly the small individual Kloostri Massif in the northwestern part of the Märjamaa Intrusive) are more leucocratic in composition and, in places, have a trachytoidic structure.

Two generations of quartz (20-30 %) have been distinguished. The first generation consists partly of anhedral crystals between potassium feldspar individuals and partly of euhedral inclusions within microcline. The second generation occurs as anhedral crystals in the groundmass. Microcline (20-40 %) is present as phenocrysts (diameter about 2-3 cm) and in the groundmass. The phenocrysts are often perthitic and contain inclusions of quartz, biotite and rare, titanite. Plagioclase (20-40 %), which forms euhedral tabular or prismatic crystals, is represented by oligoclase-andesine. Anhedral crystals of biotite (2-10 %) are often clustered together as swarms of small or large flakes. Hornblende (mainly in the first and, partly, in the second intrusive phase) and muscovite (in the third phase) occur sporadically. The content of the main opaque minerals, magnetite and ilmenite may reach 3 - 5 %. Apatite, fluorite, zircon, titanite and epidote are accessory minerals.

The **Taebla** body, 6-7 km in diameter, is the smallest one distinguished by drilling of two wells. It is composed of homogeneous leucocratic porphyritic granites (Soesoo & Niin 1992). In terms of the mineral and chemical composition, the Taebla granites are similar to the rocks of the third phase of the Märjamaa Pluton and to the second phase of the Naissaar and Neeme plutons.

Geophysical data indicate that the **Abja** body (8 km x 5 km), which occurs in the structural zone of South-Estonian granulites, is ellipse-shaped. In the only drill core available, in the depth interval 550-635 m, medium-grained greenish-grey gabbrodiorites (SiO_2 49-52%) with massive texture are cut by fine- and medium-grained pinkish-red potassium granites. The isotopic age (U-Pb, zircon of the gabbrodiorites is 1635 ± 7 Ma (Kirs & Petersell 1994). The age of the intersecting granites is 1622 ± 7 Ma. The gabbrodiorites comprise relatively euhedral plagioclase (An 33-39, 40-50 %), hornblende (10-20 %) and biotite (10-20%) with minor quartz and potassium feldspar. The content of accessory and opaque minerals is quite high; the most important being apatite (2-5%), titanite (1%) and titanomagnetite (2-6%).

Geophysical data suggests that the **Sigula** body of mafic rocks is a NE-trending dike (1.5km x 4 km). The one drill core available to date (depth interval 223.2-316.6 m) reveals that the pluton consists of inequigranular dark-grey massive diabase (SiO_2 47-49%) with ophitic structure (Пуура и др. 1983). The relatively large prismatic plagioclase crystals provide the rock with a slightly porphyritic outlook. The amount of plagioclase (An 55-63) is remarkable reaching 50-60%. Other minerals are hornblende (8-10%) and clinopyroxene (8-10%), biotite, orthopyroxene, quartz and potassium feldspar (all < 5%). The content of apatite (2-5%) and, especially, titanomagnetite (7-10%) is noticeably high.

Mesoproterozoic rocks of the Åland-Riga Subprovince, the 1.54-1.59 Ga age group

The largest, Riga complex rapakivi-anorthosite pluton (Богатиков и Биркис 1972, Кууспалу 1975) spread in the basement of western Latvia, southwestern Estonian Archipelago in the Gulf of Riga and Central Baltic proper, and the Åland Pluton belong to the Åland-Riga Subprovince. Defined and supposed rapakivi bodies in the northern Baltic seabed occur near the West-Estonian Archipelago (Пуура и др. 1992, Koistinen 1994). It has been mentioned that the most intense rapakivi plutonism area coincides with the junction of the Baltic proper with the gulfs of Bothnia, Finland and Riga (Puura & Flodén 1996).

The rapakivi plutons and stocks are characterised by changeable magnetic and stable density properties and, as a whole, they have a considerably massive internal structure. Thus, they are easy to identify and contour by geophysical mapping and drilling.

The Riga Pluton is an essential representative of bimodal rapakivi-anorthosite complexes. The mafic part of the pluton locates in its southern part, in southwestern Latvia. In the central part, on the Kurzeme Peninsula, a variety of both typical vyborgite- and pyterlite-like and even-grained granites occurs. In the central and southern parts of the Riga Pluton, also quartz mangerites, mangeritic granosyenites and quartz-syenites occur among mangeritic granitoids (Богатиков и Биркис 1973). All these rocks have been formed in deep levels of the crust. However, in cores of two wells penetrating into the Riga Pluton in southwestern Estonia, on the islands of Ruhnu (3.4 m of crystalline rocks) and Saaremaa (at Kuressaare Town, 28.4 m), the rocks are represented by subvolcanic granite porphyries (granophyres) with micropegmatite matrix (Кууспалу 1975, Пуура и др. 1983).

Exterior to the northern part of the Riga Pluton (Fig. 5), the Undva well penetrates into a suite of rapakivi-related volcanic rocks (Puura *et al.* 1983). The lower part of the sequence consists of plagioclase porphyrites, and the upper part of quartz porphyries.

Plagioclase porphyrites have some similarity with the rapakivi-related, more felsic porphyrites on Hogland, and with Dala porphyrites from the Transscandinavian Igneous Belt in central Sweden. The rocks are dark grey or black, in places, with pink-shaded massive and dense rocks. Their groundmass is fine- or very fine-grained, with microophitic texture, and consists of plagioclase (An 50-65, 65-75%), clino- and orthopyroxene (15-25%), hornblende (<5%), biotite, titanomagnetite, hematite and apatite. Prismatic and tabular phenocrysts of plagioclase (An 45-70), with an average size of 4mm x 5 mm (occasionally 30-40 mm), are rare forming about 3-10%. In terms of the chemical composition, plagioclase porphyrites are similar to andesites.

The quartz porphyries are brownish-red or pink massive rocks. Their fine-grained groundmass consists of quartz (30-40%), feldspar (40-50%), opaque minerals (hematite and magnetite up to 15%), chlorite, apatite and glass, and they have granophyric, radially fibrous and spherolitic texture. Small rounded phenocrysts of dark grey quartz are about 3-4 mm in diameter and make up 3-10% of the rock. Prismatic phenocrysts of plagioclase (An 1-7) and microcline-perthite are a bit larger (diameter about 5-8 mm, rarely 10-20 mm); their quantity varies between 20 and 30%.

The quartz porphyries of the Undva Member (Table 3) differ from those on Suursaari (Hogland) Island in colour and texture of groundmass. They comprise less and smaller phenocrysts, but the content of opaque minerals and apatite is higher which makes them more similar to some Dala porphyries in the Transscandiavian Igneous Belt in central Sweden.

IV SEDIMENTARY COVER

VENDIAN

The Vendian (Vendian Complex) as an independent stratigraphic unit, probably in the category of system, was distinguished in the early 1950s by B. Sokolov (Соколов 1952a, 1953). Its stratotype area is in the western part of the East-European Platform.

The Vendian of the stratotype area includes three subdivisions in the rank of regional series, which in ascending order are Vilchan, Volyn and Valdai (Келлер и Розанов 1979 б). The Valdai regional series consists of the Redkino (below) and Kotlin (above) stages, of which only the latter occurs in Estonia and forms the lowermost part of the sedimentary cover overlying the Proterozoic crystalline basement.

The current stratigraphic scheme of the Estonian Vendian was accepted in 1976 at the Baltic Stratigraphic Conference in Vilnius (Table 5).

Kotlin Stage

The unit in the rank of stage was defined as the upper part of the Valdai Series corresponding to "Laminarites" Clay on the East-European Platform (Решения... 1965). The name Kotlin was proposed by Sokolov (Мяннил 1958a) after Kotlin Island in the eastern part of the Gulf of Finland. Mens and Pirrus (Менс и Пиррус 1974, 1980, Гниловская и др. 1979, Келлер и Розанов 1979б) determined the present stratigraphic extent of the stage and worked out its classification for the East Baltic area.

The Kotlin Stage is widespread in mainland Estonia, lacking only in its southwestern part and in some local structures, including Assamalla and Uljaste (Fig.11). Stratigraphically, the most representative and thickest sections are situated in northeastern Estonia. In a westerly direction, the sections thin out rather rapidly and change in lithology.

The lower boundary of the stage coincides with the base of the sedimentary cover in Estonia, and is easy to determine. Some complications occur if the core yield is low or the core is distorted. The upper boundary is clear in eastern and central Estonia where the overlying rocks contain glauconite and mineralized skeletal fossils. In the northwestern part of mainland Estonia and on Hiiumaa Island the boundary is less distinct due to the lithological similarity with the overlying Lower Cambrian rocks. Identification of the Kotlin Stage is most complicated in the sections west of Keila (Fig. 12) where the lower part of the sedimentary cover comprises light-coloured loose quartzose sandstones with occasional lenses or interbeds of compacted multicoloured argillaceous rocks. As the latter rock type is lacking in the overlying Cambrian beds, this part of the sequence is conditionally regarded as the Kotlin Stage.

The Kotlin Stage is represented by siliciclastic rocks which accumulated under cool and humid climatic conditions (Pirrus 1992). This extensive, high-order cycle of deposition covered three shorter cycles divided as successive Gdov, Kotlin and Voronka formations. Multicoloured sandy-silty sediments consisting of low maturity and poorly sorted detrital material accumulated at the beginning of the Kotlin depositional cycle (Gdov Formation). Upwards in the section, the coarse-grained red-coloured deposits change into grey-coloured clayey sediments which accumulated during the stable phase of the Kotlin depositional cycle (Kotlin Formation). The cycle ends with the reappearance of multicoloured sediments of high maturity (Voronka Formation).

In recent years, acritarchs as the most abundant and widespread fossil group in the deposits of the Kotlin Stage have underlain the subdivision and correlation of the Vendian rocks (Volkova *et al.* 1983). Acritarchs are represented by a taxonomically simple assemblage consisting mainly of representatives of the genus *Leiosphaeridia* (Пашкявичене 1980).

Table 5. The Vendian of Estonia

Regional Standard			Index-species		Lithostratigraphic units	
Subsystem	Series	Stage	Vendotaenids	Acritarchs	Formation	Member
UPPER VENDIAN	VALDAI	KOTLIN			VORONKA	Kannuka
						Sirgala
			Vendotaenia antiqua Gn. *Aataenia reticularis* Gn.	*Leiosphaeridia pelucida* (Schep.) *L. aperta* (Schep.) *L. culta* (Andr.) *L. effusa* (Schep.)	KOTLIN	Laagna
						Meriküla
						Jaama
					GDOV	Uusküla
						Moldova
						Oru

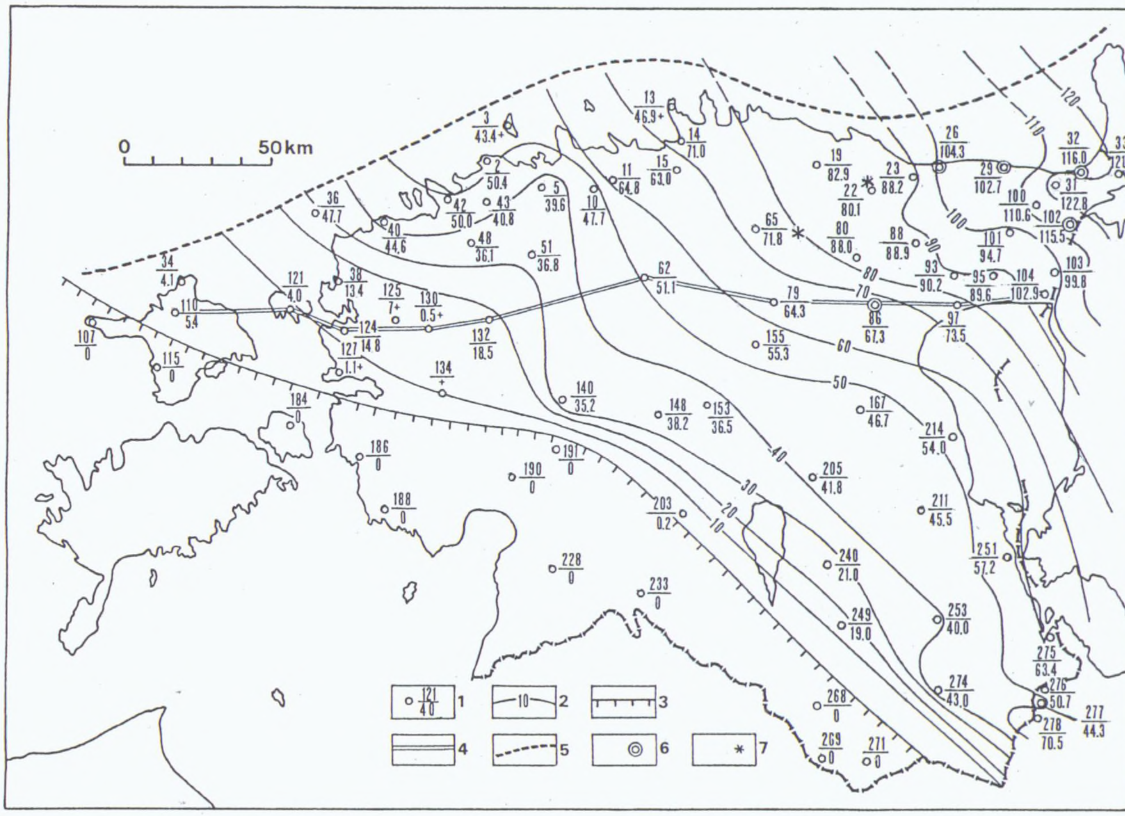


Fig. 11. Distribution and thickness of Vendian rocks: 1 - borehole, the number of borehole (see Fig. 3) in numerator and the thickness of the Vendian rocks in denominator; 2 - isopachs; 3 - limit of the present distribution of rocks; 4 - line of the cross-section (Fig. 12); 5 - line of erosion (northern limit of the present-day distribution of successive Cambrian rocks); 6 - stratotype sections; 7 - local structures where the Vendian rocks are absent.

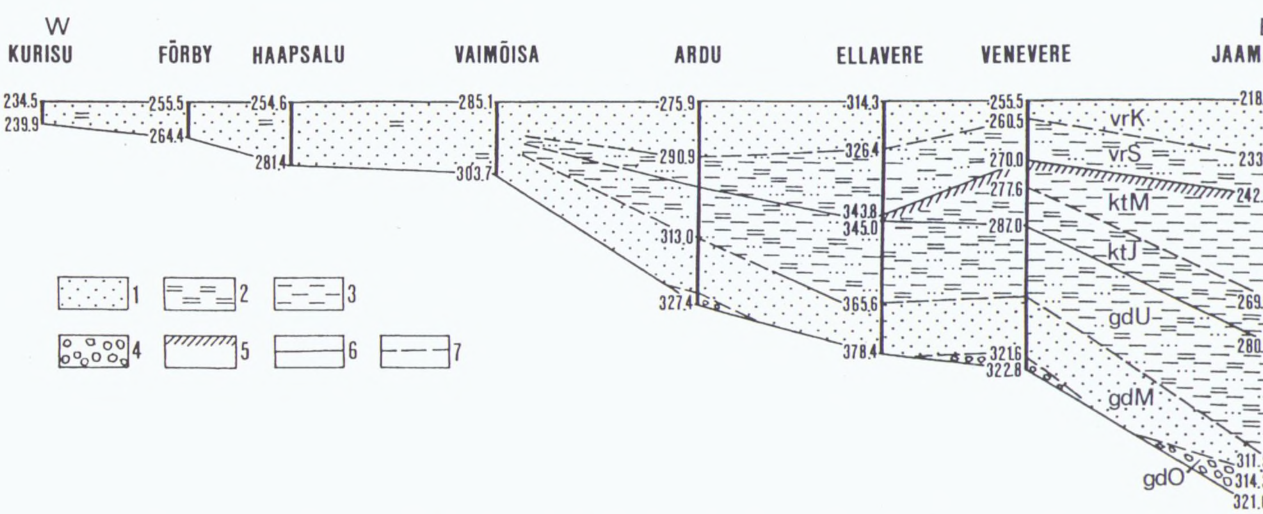


Fig. 12. Latitudinal cross-section of the Vendian strata in northern Estonia: 1 - sandstone; 2 - siltstone; 3 - claystone; 4 - unsorted clastic rocks with coarse-grained constituents; 5 - phenomena of weathering; 6 - boundary between formations; 7 - boundary between members. Indices: gdO - Oru Member, gdM - Moldova Member, gdU - Uusküla Member, ktJ - Jaama Member, ktM - Meriküla Member, vrS - Sirgala Member, vrK - Kannuka Member (see Менс и Пиррус 1980).

Microfossils are accompanied by vendotaenids, of which *Vendotaenia antiqua* Gn. is most common, while *Aataenia reticularis* Gn. is rare (Гниловская и др. 1979). Besides microfossils and vendotaenids, fragments of shapeless organic matter occur on the bedding surface. All the above-listed palaeontological finds occur in the rocks of the Kotlin Formation which have promoted their accumulation and preservation. In some sections in the northeasternmost part of Estonia (Meriküla, Sinimäe), the grey argillaceous rocks of the upper member of the Gdov Formation comprise acritarchs and organic matter of irregular form.

The **Gdov Formation**. Asatkin (Асаткин 1937) derived the name from the Gdov beds used as a division to denote the sandy strata between the "Laminarites" Clay and the crystal-

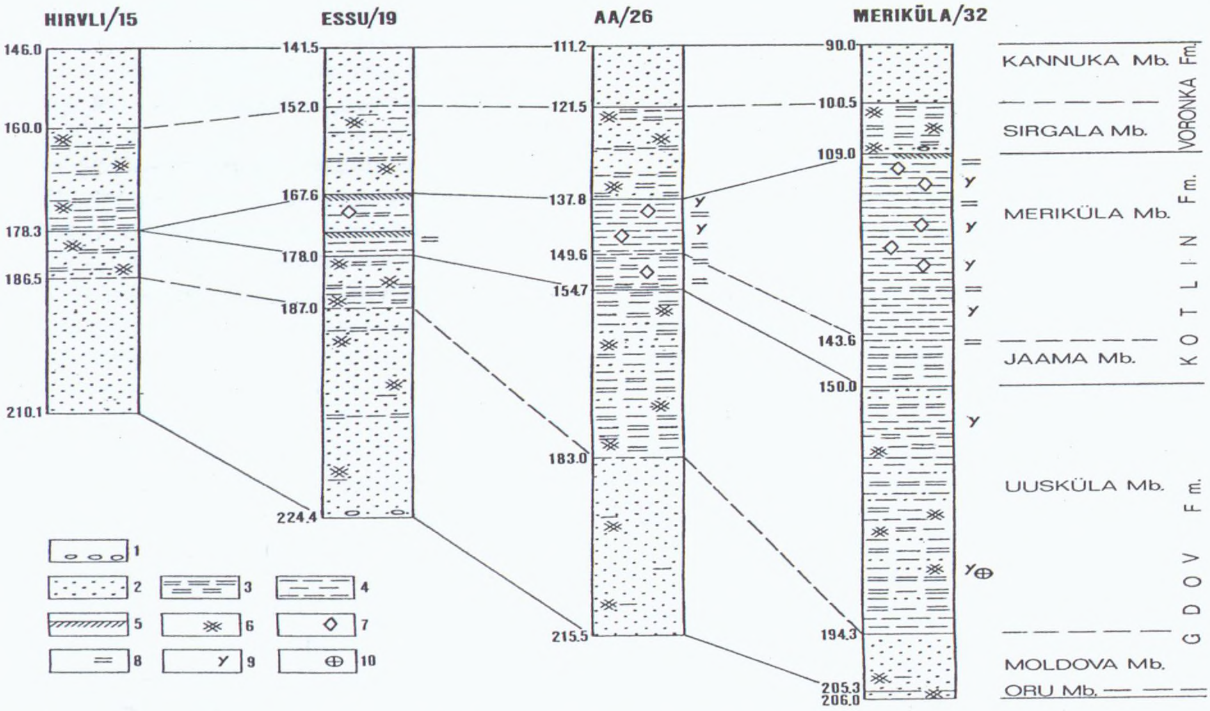


Fig. 13. Correlation of the Vendian boreholes in northern Estonia and the occurrence of fossils: 1 - admixture of coarse-grained constituents in rocks; 2 - sandstone; 3 - siltstone; 4 - claystone; 5 - phenomena of weathering; 6 - variegated (red-coloured) rocks; 7 - siderite nodules; 8 - shapeless films of organic matter; 9 - vendotaenids; 10 - the Kotlin acritarch assemblage.

line basement in the northwestern part of the East-European Platform. The Gdov Formation is considered as the lower part of the Kotlin Stage accumulated during the initial phase of the Late Valdaian transgression over the northwesternmost part of the East-European Platform, including the present-day Estonia, Latvia, and the western part of the Leningrad Region.

In Estonia, the Gdov Formation rests immediately upon the crystalline basement and spreads in subsurface lying in the northern, eastern and central parts of the Republic. Its thickness ranges from 0.2 to 58.3 m (Fig. 11 - *203, 102*). The Venevere (Fig. 11 - *86*) drill core in the interval of 287-322.8 m has been selected as a hypostratotype for Estonia (Fig. 12). The formation prevalently consists of multicoloured sandstones of various grain-size. The uppermost and, locally, also the lowermost part comprises a considerable quantity of reddish and purplish argillaceous rocks. The sandstones are represented by arkose and feldspatic varieties comprising besides quartz up to 50 per cent of feldspars. Micas, both muscovite and green altered biotite, are occasionally found. The mineral composition of the clay fraction is rather stable throughout the formation, being characterised by illite-kaolinite suite (Пиррус 1970). On the basis of lithological features, the Gdov Formation is subdivided into three members which in ascending order are Oru, Moldova and Uusküla (Table 5).

The ***Oru Member*** occurs locally in the base of the Gdov Formation. It accumulated in depressions of the crystalline basement on account of its weathering crust. The greatest thickness of the Oru Member (6.7 m) has been recorded in the Jaama borehole (Fig. 11 - *104*). The member consists of red unsorted clayey-sandy-gravely deposits (mixtite). Unlike the overlying members, the deposits of the Oru Member differ both structurally and mineralogically. Quartz is the prevailing mineral in sand and gravel (up to 90%); its grains are angular or subangular with nonsorted size distribution. Feldspar is not common (less than 10%). The clayey matrix consists mostly of kaolinite. All this suggests that these deposits were formed in deluvial fans as a result of intense weathering of acid crystalline rocks.

The ***Moldova Member*** overlies either the Oru Member or the crystalline basement. Its thickness is usually 30-40 m, maximum 50 m. The member consists of yellowish or pinkish arkose and/or feldspatic sandstones of various grain size and a few interlayers of multicoloured, frequently reddish, argillaceous rocks. Hence, the clastic material transported into the sedimentary basin originates from a close-lying area with a low degree of weathering of rocks.

The ***Uusküla Member*** is the most fine-grained and multicoloured part of the Gdov Formation. It consists of silty claystones intercalated with silt- and sandstones. The number and thickness of claystone layers increases eastwards. In the same direction the rocks gradually loose the red colour until in the easternmost sections they are predominantly grey. The proportion of sandstones is low and they are mainly represented by arkose and feldspatic types. Micas occur in remarkable quantities, but rock fragments are uncommon.

The composition and structure of the rocks suggest a rather low hydrodynamic energy of the sedimentation basin.

The **Kotlin Formation**. The name was introduced by Sokolov to designate the "Laminarites" Clay (Мянниил 1958a). The formation is spread in eastern Estonia, in a more typical form in its northeastern part attaining a thickness of 52.6 m in the Narva borehole (Fig. 11 - *33*). In the west direction the thickness decreases quickly pinching out on the Tapa - Ellavere line (Figs. 12, 13). The formation is known only from core sections, and the interval of 109-150 m of the Meriküla core

(Fig. 11 - *32*) has been defined as the hypostratotype. The lower boundary is drawn at the level where multicoloured deposits turn grey. At the base of the formation, gravel and coarse-grained sand occur locally.

The dominant components of the formation are thinly laminated grey claystones with intercalating light-coloured very fine-grained sandstones or siltstones, or both. The lamination is complicated by the occurrence of dark-brown films of organic matter.

The rocks of the formation are low in sand and silt, the content of which in the upper- and lowermost parts only locally exceeds 50%. Quartz and feldspars (particularly K-feldspar) are the main detrital minerals of sand and gravel grain-size. The content of micas, including biotite and muscovite, is also notable. Their ratios depend on the type of rock. The rocks are characterised by a small content of both opaque and transparent allogenic minerals. Heavy minerals are dominated by siderite and pyrite of authigenic origin. Illite is a dominant clay mineral, the content of kaolinite ranges from 15 to 40%, the latter value being fixed in the lowermost part. Chlorite is common, in the middle part of the formation its average content is 15-20% (Пиррус 1970).

On the basis of the lithological composition, the Kotlin Formation is divided into three members (Table 5).

The ***Jaama Member*** is made up of alternating grey-coloured massive siltstones and thinly laminated claystones. A few siderite nodules and organic matter films occur. This is the first stage in the large-scale clay accumulation in the eastern part of Estonia.

The ***Meriküla Member*** is the most typical unit of the Kotlin Formation. It is represented by the "Laminarites" Clay consisting of rhytmically alternating 0.5-0.8 mm thick pairs of dark-grey fine-dispersed clay layers and light-grey laminae higher in silt. The bedding plane is covered by dark-brown shapeless organic films. Vendotaenides, small flakes of mica and siderite nodules are common.

The fairly uniform mineral composition shows that the source areas must have been located relatively far from the depositional basin.

The ***Laagna Member*** has the most restricted distribution area, compared to other members of the formation. In the northeasternmost part of Estonia it is up to 6 m thick. The member consists of grey clayey and silty-clayey argillaceous rocks with many up-to-20-cm-thick intercalations of siltstone. The typical "Laminariates" Clay layers are absent. Scarce organic matter films and small siderite nodules occur suggesting the terminal phase of clay accumulation under weak hydrodynamic conditions.

The **Voronka Formation** was established by Mens and Pirrus (Менс и Пиррус 1971). Earlier, this part of the sequence was treated as two lower units of the post-Laminarites Sandstone or as the lower and middle parts of the Lomonossov Formation (Мардла и др. 1968). The type section of the formation is an outcrop on the lower reaches of the Voronka River, Russia (Менс и Пиррус 1971). Beyond the stratotype area, the formation is of subsurface occurrence being known in eastern and northern Estonia and in eastern Latvia. The Meriküla (Fig. 11 - *32*) drill core in the interval of 90-109 m serves as a hypostratotype for the Voronka Formation (Менс и Пиррус 1980). The formation occurs between the overlying Lontova Formation and the weathering crust of the underlying Kotlin Formation (Менс и Пиррус 1969, 1970). In Estonia, the thickness of the formation ranges from 10 to 40 m. The Voronka Formation consists of variable siliciclastic rocks and represents a single upwards coarsening cycle from argillaceous rocks to well-sorted sandstones. The lower boundary of the formation is drawn on the basis of the change in colour. Based on lithological evidence, the formation is divided into the Sirgala and Kannuka members (Table 5).

The ***Sirgala Member*** consists of alternating multicoloured clays and siltstones with interlayers and lenses of light-coloured sandstones, the share of which increases upward the section. Most of detrital grains are subrounded quartz with a small quantity of feldspars (up to 10%) and micas (mainly muscovite). In the clay fraction, kaolinite slightly prevails over illite. Chlorite is uncommon.

The mineral composition suggests that these deposits derived from the weathered zone of sedimentary rocks.

The ***Kannuka Member*** consists entirely of light weakly cemented fine- to medium-grained quartzose sandstones with a few thin interlayers of multicoloured clayey siltstones, which are similar to the underlying deposits of the Sirgala Member. The clay fraction is dominated by kaolinite. Increase in the maturity in minerals upward the section is indicative of the redeposition of older sediments.

CAMBRIAN

Cambrian rocks are widespread in Estonia. They are missing on the crest of the Valmiera-Lokno swell and on some peninsulas on the southern coast of the Gulf of Finland. Exposed Cambrian rocks are encountered in outcrops along the Baltic Klint, but mostly they are overlain by younger rocks and the basic data for studies has been obtained by means of boreholes.

The main pioneering work towards the subdivision of the Estonian Cambrian (then the Lower Silurian) was done by Eichwald (1854) and Schmidt (1888) who worked out the lithostratigraphical subdivision and described the first fossils found in these rocks. A zonal division and modern nomenclature were introduced by Öpik (1933, 1956).

Up to the middle of this century, the Cambrian stratigraphy was based on outcrop sections and embraced the lowermost part of the Lower Cambrian succession of the studied area and the problematic *Acrotreta* Zone of the Upper Cambrian (Öpik 1956).

During the last fifty years, numerous deep borings were made which revealed the full thickness of the Cambrian rocks. Elaboration of Cambrian stratigraphy was greatly promoted by identification of plant microfossils (now known as acritarchs) by Naumova, Timofejev and Volkova on the East-European Platform. Palynological studies provided valuable data for establishing distinct acritarch assemblages, their ranges in the sequence and relationship with trilobite zones.

Elaboration of the present-day stratigraphical subdivision of the Estonian Cambrian (Table 6) was favourably influenced by international cooperation with researchers from neighbouring countries, and supported by IGCP projects No. 29 and 86. The results obtained were summarized in the stratigraphic scheme accepted in 1976 at the Baltic Stratigraphic Conference in Vilnius (Решения... 1978) and improved in 1983 by the Stratigraphic Conference on the Cambrian of the East-European Platform (Решение... 1986).

As the global standard is still under preparation, the boundaries between the Lower/Middle and Middle/Upper Cambrian are not strictly formal. There are no generally agreed names for the stages and their boundaries are unclear.

The Cambrian stratigraphic scale in the East-European Platform from the base of the *Sabellidites cambriensis* Zone to the top of the *Acerocare* Zone is based mainly upon the succession of trilobites, except the lowermost part where trilobites are lacking (Mens *et al.* 1990). Only part of the Cambrian is present in Estonia (Table 6). The lower boundary of the system is distinct in the studied area, and coincides with regional changes in the sedimentary conditions which led to the accumulation of normal marine sediments (Pirrus 1993). This level is marked by the appearance of primitive skeleton-forming organisms and changes in the composition of ichno- and phytofossils.

The upper boundary of the system is not obvious although, biostratigraphically, the Cambrian-Ordovician transition in Estonia is relatively well studied (Mens *et al.* 1993). This is due to the circumstance that the IUGS has not yet passed the final decision on the Ordovician lower boundary.

Conventionally, the Cambrian is subdivided into three subsystems: Lower, Middle and Upper. The Lower/Middle

Table 6. The Cambrian of Estonia

SUBSYSTEM	Biozones of the East-European Platform		Regional Stage	Group	LITHOSTRATIGRAPHIC UNITS: WEST ESTONIA	LITHOSTRATIGRAPHIC UNITS: NORTH ESTONIA	LITHOSTRATIGRAPHIC UNITS: SOUTH-EAST ESTONIA
UPPER CAMBRIAN	*Acerocare*				KALLAVERE Fm (lower part)	KALLAVERE Fm (lower part)	KALLAVERE Fm (lower part)
UPPER CAMBRIAN	*Peltura*						
UPPER CAMBRIAN	*Leptoplastus*					TSITRE Fm	
UPPER CAMBRIAN	*Parabolina spinulosa*					ÜLGASE Fm	
UPPER CAMBRIAN	*Olenus & Agnostus (Homogn.)*						PETSERI Fm
UPPER CAMBRIAN	*Agnostus pisiformis*						
MIDDLE CAMBRIAN	*P. forchhammeri*	*Lejopyge laevigata*					
MIDDLE CAMBRIAN	*P. forchhammeri*	*Solenopleura brachymetopa*					
MIDDLE CAMBRIAN	*P. paradoxissimus*	*Ptychagnostus punctuosus*					PAALA Fm
MIDDLE CAMBRIAN	*P. paradoxissimus*	*Ptychagnostus atavus*					PAALA Fm
MIDDLE CAMBRIAN	*P. paradoxissimus*	*Ptychagnostus gibbus*	PANERIAI				PAALA Fm
MIDDLE CAMBRIAN	*E. oelandicus*	*Ptychagnostus praecurrens*	DEIMENA	Deimena	RUHNU Fm		
MIDDLE CAMBRIAN	*E. oelandicus*	*Eccaparadoxides insularis*	KYBARTAI				
LOWER CAMBRIAN	*Proampyx linnarssoni*		RAUSVÉ	Aisčiai			
LOWER CAMBRIAN	*Holmia kjerulfi*		VĒRGALE	Aisčiai	IRBEN Fm		VAKI Fm
LOWER CAMBRIAN	*Holmia inusitata*		LJUBOML'	Aisčiai	SOELA Fm		
LOWER CAMBRIAN	*Mobergella & S. mickwitzi*		DOMINOPOL'	Liivi	TISKRE Fm LÜKATI Fm	TISKRE Fm (Rannamõisa Mb, Kakumägi Mb) LÜKATI Fm	
LOWER CAMBRIAN	*Rusophycus parallelum*		DOMINOPOL'	Liivi	SÕRU Fm		
LOWER CAMBRIAN	*Platysolenites antiquissimus*		LONTOVA	Balti	VOOSI Fm (Paralepa Mb, Kasari Mb, Taebla Mb)	LONTOVA Fm (Tammneeme Mb, Kestla Mb, Mahu Mb, Sämi Mb)	LONTOVA Fm (Kestla Mb, Mahu Mb, Sämi Mb)
LOWER CAMBRIAN	*Sabellidites cambriensis*		ROVNO				

Cambrian boundary is at the base of beds with *Paradoxides*, more exactly *Eccaparadoxides insularis*, and the Middle/Upper Cambrian boundary is at the base of the *Agnostus pisiformis* Zone for the East-European Platform (Table 6).

Rocks of all three subsystems are encountered in Estonia, but the degree of completeness varies. Compared to other subsystems, the Lower Cambrian rocks are most widespread and thickest. Their two-folded structure results from remolding of the basin prior to the Liivi transgression. Of the six regional stages established on the ground of the succession of acritarch assemblages in the Lower Cambrian on the platform, four are present in Estonia (Table 6).

The Middle Cambrian sequence in Estonia is entirely devoid of fossils and the regional stages have been established on the basis of lithological criteria.

The Upper Cambrian is documented on the basis of palaeontological evidence, derived from both the shelly fauna and acritarchs.

No regional stages (except the Kybartai and Deimena in the lowermost Middle Cambrian) have yet been differentiated in the rest of the Middle and through the Upper Cambrian in Estonia. The relevant rocks have been treated only in general lines by lithostratigraphic units.

Based on the stratigraphical completeness of the sections and facies conditions, Estonia's territory is subdivided into the northern, western and southeastern regions (Брангулис и др. 1974, 1975, Table 6), with their characterisitic formations and members.

LOWER CAMBRIAN

Lontova Stage

The oldest Cambrian rocks in Estonia were formed during the Baltic evolutionary stage in the pre-trilobite Early Cambrian (Менс 1981). The onset of sedimentation in the Early Cambrian in Estonia corresponds to the *Platysolenites antiquissimus* Zone defined as the Lontova Stage. The Rovno Stage, composed of the lowermost rocks of the Baltic evolutionary stage, is lacking in Estonia (Table 6).

The name Lontova was introduced by Öpik (1933) in the rank of beds to designate the "Blue Clay" proper. It corresponds to the upper part of the blue clays by Schmidt (1888) and Mickwitz (1911), the Lontova beds by Öpik without the uppermost layers with *Volborthella tenuis* (Öpik 1933, 1956) and to the *Platysolenites antiquissimus* Zone in the current use (Mens *et al.* 1990).

The rocks of the Lontova Stage crop out at the foot of the Klint and extend as a narrow belt from Tallinn to Narva. The main localities are the quarries at Kopli, Tammneeme, Kolgaküla, Kunda and Aseri (Fig. 14).

The stratotype of the stage is the Kunda quarry (Öpik 1933), subsequently complemented by the Lontova drill core in the interval of 14.0 to 88.3 m (Менс и Пиррус 1977). The stage in the stratotype section is incomplete since the deposits of the regressive phase of the Baltic sedimentary cycle are lacking. The stratigraphical completeness and thickness of the stage varies with regions: the rocks are at their thickest (ca 90 m) in northeastern Estonia (Fig. 14) and thin in a south-

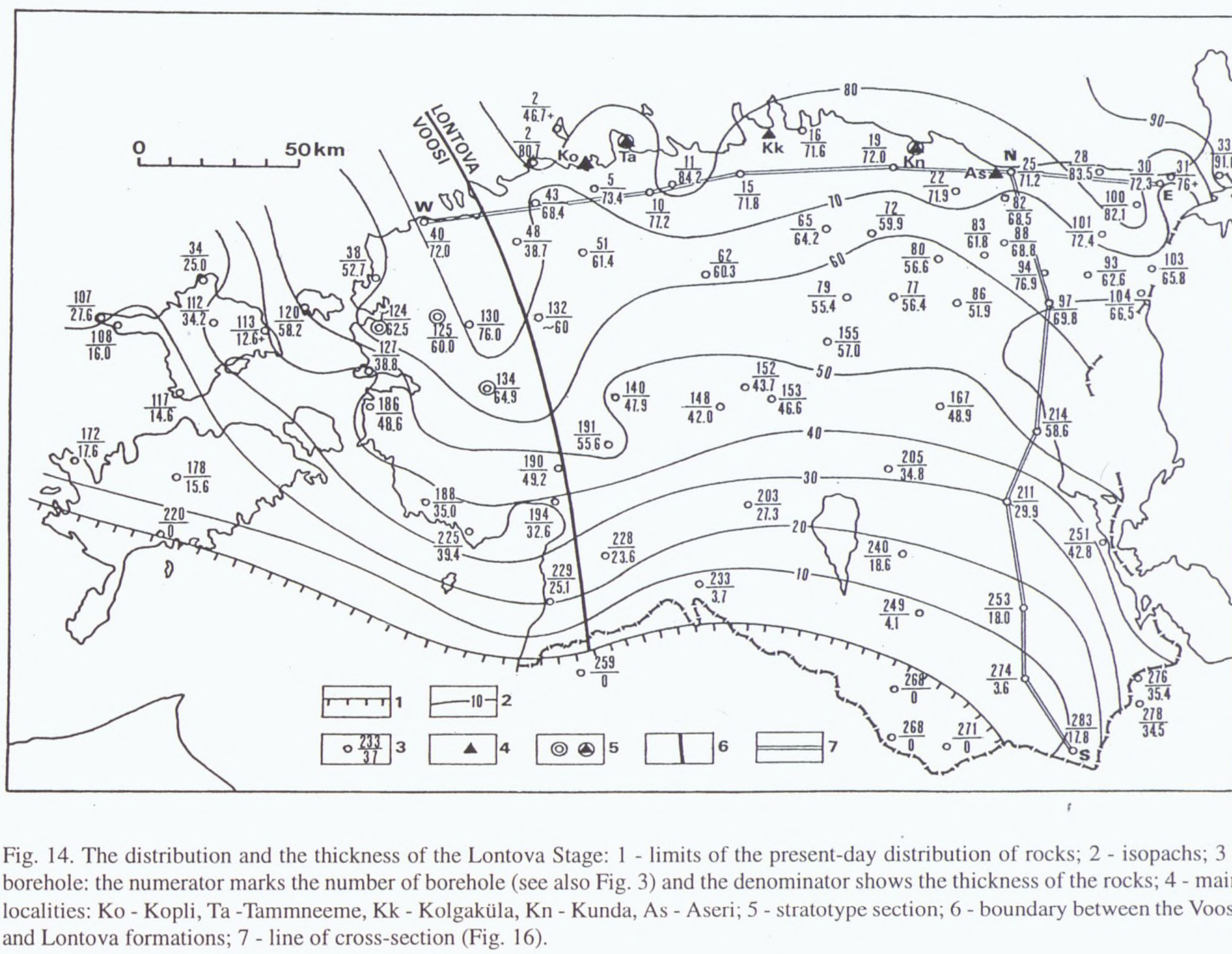


Fig. 14. The distribution and the thickness of the Lontova Stage: 1 - limits of the present-day distribution of rocks; 2 - isopachs; 3 - borehole: the numerator marks the number of borehole (see also Fig. 3) and the denominator shows the thickness of the rocks; 4 - main localities: Ko - Kopli, Ta -Tammneeme, Kk - Kolgaküla, Kn - Kunda, As - Aseri; 5 - stratotype section; 6 - boundary between the Voosi and Lontova formations; 7 - line of cross-section (Fig. 16).

erly direction due to post-sedimentary denudation. In Estonia, the Lontova Stage overlies, with a break in sedimentation, the Kotlin Stage. Its lower boundary, known only from core sections, coincides with the appearance of typical marine sediments in the succession (Менс и Пиррус 1977, Pirrus 1993).

The Lontova Stage is represented by siliciclastic rocks with a clear lateral variation of the ratio of rock types. Argillaceous rocks are prevailing in eastern and central Estonia, while sandstones dominate west of the Vihterpalu - Häädemeeste line (Fig. 14).

A relatively diverse assemblage of skeletal fossils containing *Sabellidites cambriensis* Yan., *S.* sp., *Platysolenites antiquissimus* Eichw., *P. lontova* Öpik, *P. spiralis* Posti, *Yanichevskyites petropolitanus* (Yan.), *Aldanella kunda* (Öpik) together with pyritized casts of hyolithids and hyolitelmintes, hornlike chitinous (?) sklerits, fragments of brachiopods and agglutinated tubes of *Onuphionella* has been identified from the Lontova Stage (Менс и Пиррус 1977, Менс и Пости 1984). Some of the above-mentioned species like *P. antiquissimus*, *Y. petropolitanus* and casts of hyolithids occur throughout the stage, whereas the vertical range of the rest is more limited. On the basis of the earliest appearances of the index taxa, the Lontova Stage is subdivided into four parts, which in ascending order are the *Sabellitides cambriensis* beds, *Platysolenites lontova* beds, *Aldanella kunda* beds and *P. spiralis* beds (Менс и Пости 1984).

The distribution of some species shows a distinct facies control. Thus, the hornlike chitinous (?) sklerites have been found in the argillaceous rocks of the eastern part of the territory only. The tubes of *Platysolenites* and *Yanichevskyites* are rather rare in the well-sorted sandstones in the western part of the studied area.

Acritarchs from the Lontova Stage have been described by several investigators (Наумова 1960, Волкова 1968, 1973; Янкаускас и Пости 1973, a.o.), who all agree that the stage has an acritarch assemblage of its own, which contains besides leiosphaerides and tasmanites also marginats forms. The frequency and diversity of acritarchs in the assemblage clearly depend on palaeoenvironmental conditions and facies changes in the basin of sedimentation (Менс и Пашкявичене 1981).

Trace fossils from the Lontova Stage are diverse and comprise numerous ichnospecies, among them *Phycodes pedum* Seilacher (Palij *et al.* 1983).

In conformity with the ratio of rock types in the succession, two formations laterately replacing each other, have been distinguished in the Lontova Stage (Кала и др. 1981б).

The Lontova Formation was identified in the rank of formation by Männil in 1958. Its type section in the Kunda quarry has been selected for the stratotype of the Lontova Stage (see above). The formation occurs in northern, eastern and central Estonia (Fig. 14), and is westwards laterally replaced by the Voosi Formation. The Lontova Formation is represented by greenish-grey and variegated argillaceous rocks with interbeds of coarse- to fine-grained sandstone in the lowermost and fine-grained sandstones in the uppermost part. The formation is subdivided in ascending order into the Sämi, Mahu, Kestla and Tammneeme members (Кала и др. 1970, Менс и Пиррус 1977).

The Sämi Member consists of alternating sandstones and argillaceous rocks containing glauconite and, occasionally, also flattened phosphatized pebbles.

The Mahu Member is made up of greenish-grey sandy or silty claystones with thin interlayers of sandstone.

The Kestla Member is characterized by homogenous multicoloured claystones with greenish-grey, reddish-brown and purplish interbeds, strips and spots. The admixture of sandy material is limited.

The Tammneeme Member consists of fine-grained sandstones and greenish-grey claystones and occurs in a limited area (Fig.15).

The Voosi Formation was defined on the basis of lithological evidence. Sandstones account for more than 50% of its composition (Кала и др. 1981б). Its stratotype is the interval of 237.5 to 300 m in the Haapsalu-3 drill core (Fig. 14-*124*)

The formation is distributed in the northwestern part of Estonia; in a easterly direction it is gradually replaced by the

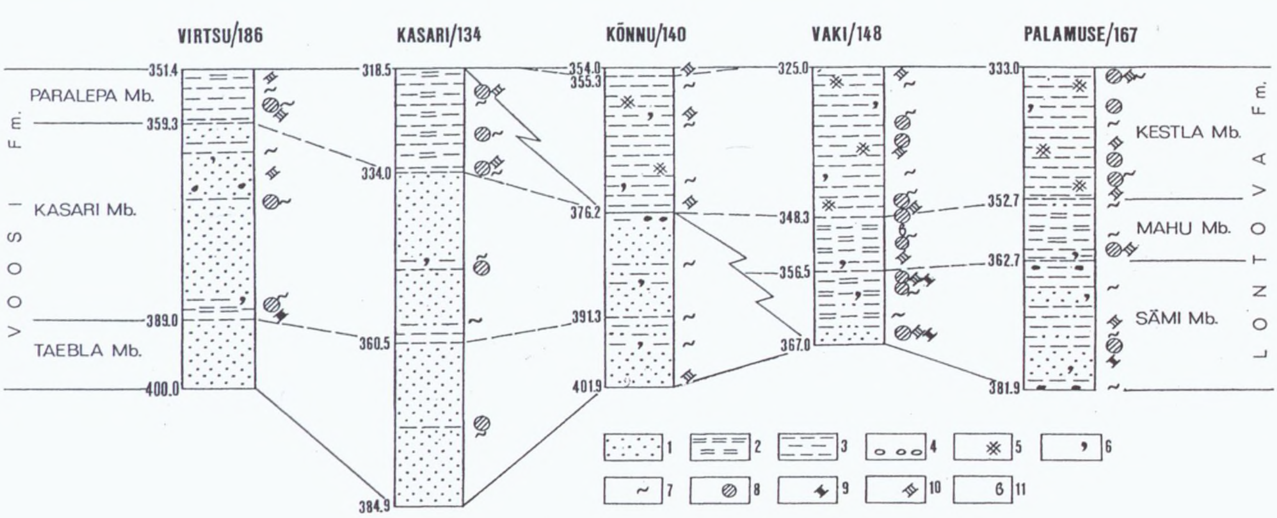


Fig. 15. Correlation of the sections of the Lontova Stage between Virtsu and Palamuse, and the distribution of the most common fossils: 1 - sandstone; 2 - siltstone; 3 - clay; 4 - pebbles; 5 - red-coloured rocks; 6 - glauconite; 7 - ichnites; 8 - Lontova acritarch assemblage; 9 - sabelliditids; 10 - platysolenitids; 11 - *Aldanella kunda*. The numbers of the boreholes correspond to those in Fig. 14.

Lontova Formation. Its lower part extends farther to the east (Fig.16). The thickness of the formation ranges from 72 to 14.6 m.

The formation consists mostly of quartzose sandstones which is the dominant type of rock on the islands of the West-Estonian Archipelago. Claystones are of minor importance and associate mostly with the upper part of the formation in mainland Estonia.

The formation is subdivided in ascending order into the Taebla, Kasari and Paralepa members.

The Taebla Member consists of light fine-grained sandstones with a few thin interbeds of silty claystones. Glauconite is not common.

The Kasari Member is represented by sandstones of various grain-size and with limited claystone content. The sandstones are rather rich in glauconite and in some places flat phosphatized pebbles occur. The sandstones of the Kasari Member are quite similar to those in the Sämi Member of the Lontova Formation.

The Paralepa Member consists of interbedded greenish-grey (with a few purplish spots) argillaceous rocks, mainly silty claystones, and of fine-grained sandstones.

Dominopol' Stage

According to the currently accepted correlation (Решение... 1986), the lowermost part of the trilobite-bearing Cambrian on the East-European Platform, deposited during the Liivi evolutionary stage in the East Baltic area (Менс 1981), is defined as the Dominopol' Stage.

Previously, this unit in the same stratigraphical extent was referred to as the Lükati Stage (Арень и др.1975, Келлер и Розанов 1979а), as the Talsy (= Lükati) Stage (Келлер и Розанов 1979б) or as the Talsy Stage (Биркис и др. 1970, Брангулис и др. 1981). It should be pointed out that the name Lükati in the rank of stage is also used in a more restricted stratigraphical extent (Мардла и др. 1968, Менс 1986).

The Dominopol' Stage was distinguished by Kirjanov (Кирянов 1969) with the stratotype section in the interval of 617.2 to 747 m of the Berezhki 2944 drill core situated in Volyn, the western Ukraine.

In Estonia, the Dominopol' Stage is represented by three succeeding formations: Sõru, Lükati and Tiskre (Table 6) well recognizable on the basis of lithological and palaeontological evidence. The Dominopol' Stage occurs on the islands of the West- Estonian Archipelago (except Ruhnu) and in the western, northern and central parts of mainland Estonia (Fig. 17). Only the upper part of the stage (Lükati and Tiskre formations) is exposed along the North-Estonian Klint. The outcrop extends from the Pakri Peninsula in the west up to the Narva River in the east. The main localities are Türisalu, Rannamõisa, Kakumägi, Lükati (Photo 15), Saviranna, Kunda, and Utria (Fig 17). The maximum thickness of the stage (76.6 m) has been fixed in the Kalana borehole where all three formations occur.

The stage consists of siliciclastic rocks, mainly sandstones. The lower boundary of the stage is lithologically distinct, accompanied by a change in the faunal composition and marked, in some places, by lenses of conglomerate.

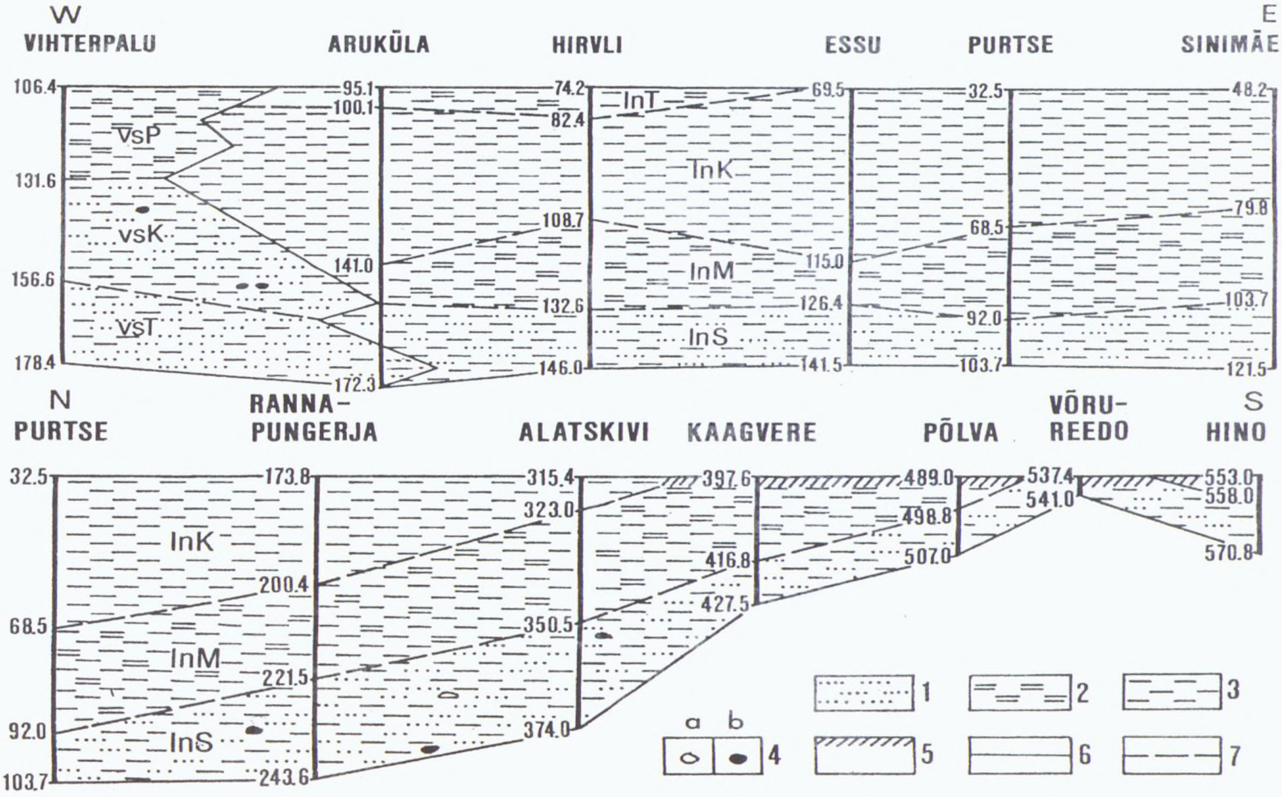


Fig. 16. Latitudinal and meridional cross-sections of the Lontova Stage: 1 - sandstone; 2 - siltstone; 3 - clay; 4 - pebbles: a - clayey, b - phosphatized; 5 - traces of weathering; 6 - boundary of formations; 7 - boundary of members. Indices: vsT - Taebla Member, vsK - Kasari Member, vsP - Paralepa Member, lnS - Sämi Member, lnM - Mahu Member, lnK - Kestla Member, lnT - Tammneeme Member. For location of cross-sections see Fig. 14.

The palaeontological finds in the lower part of the stage (Sõru Formation) are scarce. These are mostly trace fossils, rare shells of agglutinated foraminifers and a few acritarchs. The latter are represented by leiosphaerides and rare *Globosphaeridium cf. cerinum* (Volk.), *Asteridium pallidum* (Volk.), *Loposphaeridium tentativum* Volk. and *Tasmanites bobrowskae* Waz. (Менс 1986). This part of the stage corresponds to the *Rusophycus parallelum* Zone.

The middle part of the stage (Lükati Formation) is palaeontologically well characterized. The most typical species include *Volborthella tenuis* Schm., *V. conica* Schindewolf, *Schmidtiellus mickwitzi* (Schm.), *Mickwitzia monilifera* (Linnars.), torellellids, hyolithids, agglutinated foraminifers ("*Lycatiella*"), and in the west *Platysolenites antiquissimus* Eichw. has been identified in the basal beds. The acritarch assemblage in this part of the stage is abundant and diverse, with baltisphaerids dominating (Менс и Пиррус 1977). The middle and upper parts of the stage correspond to the *Schmidtiellus mickwitzi* Zone.

Of the upper part of the stage (Tiskre Formation), only its lower part (Kakumägi Member) is palaeontologically well characterized. It contains, particularly in the conglomerate lenses, *Mickwitzia monilifera* (Linnars.), *M. formosa* Wiman, *M. concentrica* Gorjansky, *Paterina rara* Gorjansky, *Scenella discinoides* Schm., *S. tuberculata* Schm., *Bradoria? estonica* Melnikova, *Konicekion kundaensis* Melnikova and fragments of trilobites. In the uppermost part (Rannamõisa Member), only occasional indeterminable fragments of brachiopods and trace fossils occur. Acritarchs have been identified in the Kakumägi Member, and only once they were found in the drill core of the Rannamõisa Member (Muraste-2 borehole). Its assemblage is much the same as in the Lükati Formation, except the appearance of *Tasmanites piritaensis* Posti *et* Jank.

Based on the differences in the palaeontological composition and distribution area as well as clear contacts between the three formations of the Dominopol' Stage, these parts of the sequence are considered independent stages (Менс 1986).

The Sõru Formation, resting transgressively either on the Lontova Formation or on the crystalline basement, occurs in the northwestern part of mainland Estonia and on the islands of the West-Estonian Archipelago, except Ruhnu (Fig.18).

Rocks of this formation are known only by core sections. The thickness of the formation ranges from 6.2 m at Vihterpalu to 58.2 m at Eikla (Fig.17-*40*, *178*). The Tahkuna drill core (Fig. 17-*34*) in northern Hiiumaa in the interval of 100.5 to 147 m has been selected as the type section of the Sõru Formation (Решения... 1978, Кала и др. 1984a).

The lower part of the formation consists mostly of massive fine-grained quartzose-feldspatic sandstone with thin clay interbeds and films. The upper part is represented by a complex of interbedded argillaceous rocks and sandstones. In both parts the rocks are light-grey with greenish-grey shade, but red and purplish-red patches also occur.

The Lükati Formation, the most widespread division of the Dominopol' Stage (Figs. 18, 19), lies transgressively on

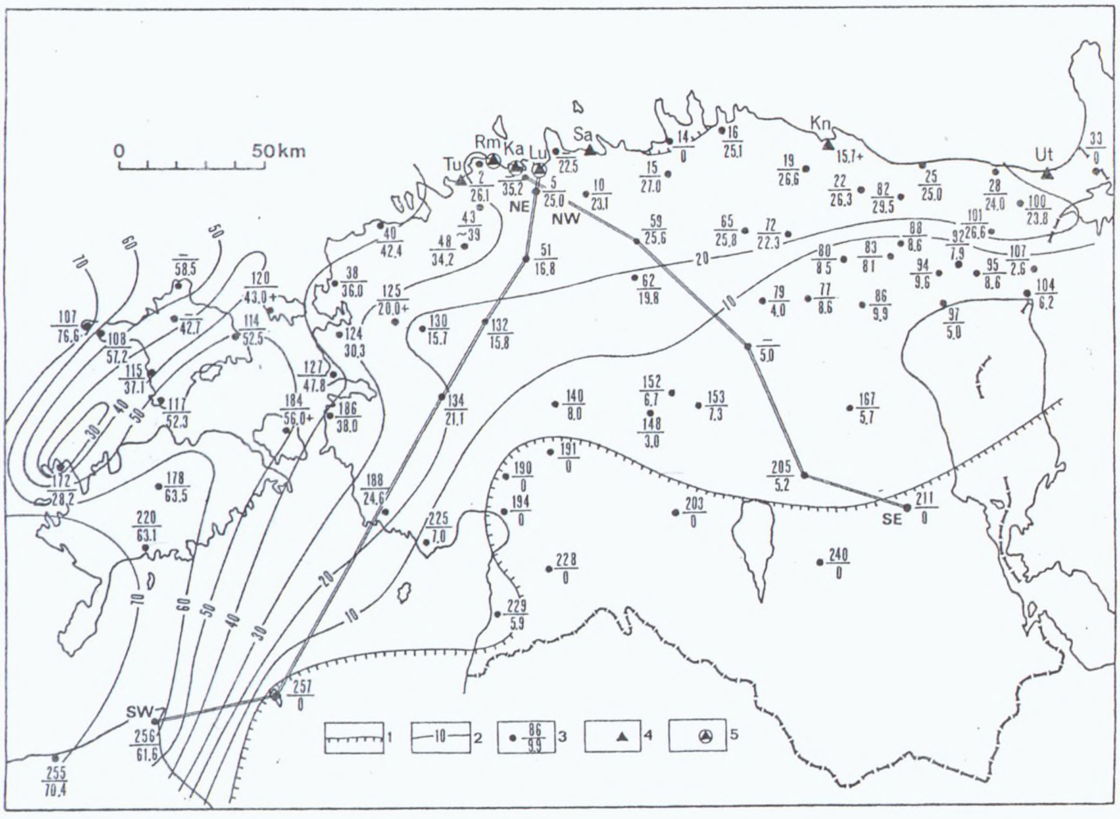


Fig. 17. Distribution and thickness of the Dominopol' Stage: 1 - limits of the present-day distribution of rocks in mainland; 2 - isopachs; 3 - borehole: the numerator marks the number of borehole (see also Fig. 3) and the denominator shows the thickness of the stage; 4 - main localities: Tü - Türisalu, Rm - Rannamõisa, Ka - Kakumägi, Lü - Kose-Lükati, Sa - Saviranna, Ku -Kunda, Ut - Utria; 5 - stratotype section. The NE-SW and NW-SE lines indicate transects of stratigraphical cross-sections presented in Fig. 18.

Photo 15. Type section of the Lükati Formation in the left bank of the Pirita River at Lükati. In the middle of the section, there is the boundary between the Lükati and Tiskre formations. *Photo by A. Rõõmusoks.*

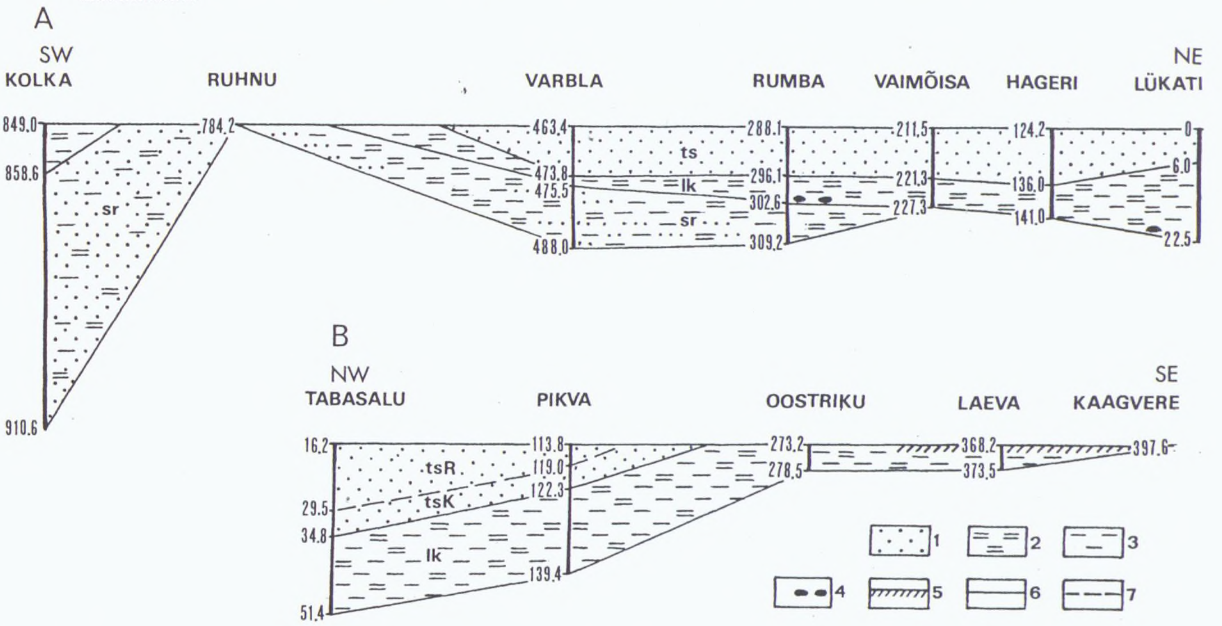


Fig. 18. Cross-section of the Dominopol' Stage along SW-NE (A) and NW-SE (B) lines (Fig. 17): 1 - sandstone; 2 - siltstone; 3 - clay; 4 - pebbles; 5 - traces of weathering; 6 - boundary of formations; 7 - boundary of members. Indices: sr - Sõru Formation, lk - Lükati Formation, ts - Tiskre Formation, tsK- Kakumägi Member, tsR - Rannamõisa Member.

the Sõru Formation in the west and on the Lontova Formation in the east. It is often separated from the underlying units by conglomerate lenses containing pebbles of phosphatized sandstones (Менс и Пиррус 1975). Outside the distribution area of the Tiskre Formation, the upper part of the Lükati Formation is often weathered. The Lükati Formation, formed during the stable phase of the Liivi evolutionary stage, corresponds to the whole stratigraphical extent of the Dominopol' Stage along the southern margin of its distribution area in Estonia.

In the type area in the vicinity of Tallinn, the formation reaches 20 m in thickness. The type section of the formation is an outcrop on the left bank of the Pirita River (Photo 15) in the eastern outskirts of Tallinn (Öpik 1933), complemented today by a drill core from the lower part of the formation (Менс и Пиррус 1977).

The lithologically monotonous formation consists of interbedded greenish-grey argillaceous rocks and very fine-grained sandstones. The latter form mostly 0.1—0.3-m-thick layers, some of which are hard, in places tightly cemented by poikilotopic carbonate. The upper surfaces of the sandstone layers are covered with ripple marks, the lower surfaces with casts of various trace fossils and mud cracks. Pyrite and glauconite are very common. In the lower part, glauconite often forms 1 - 3-cm-thick laminae.

The Tiskre Formation in the current use is interpreted as

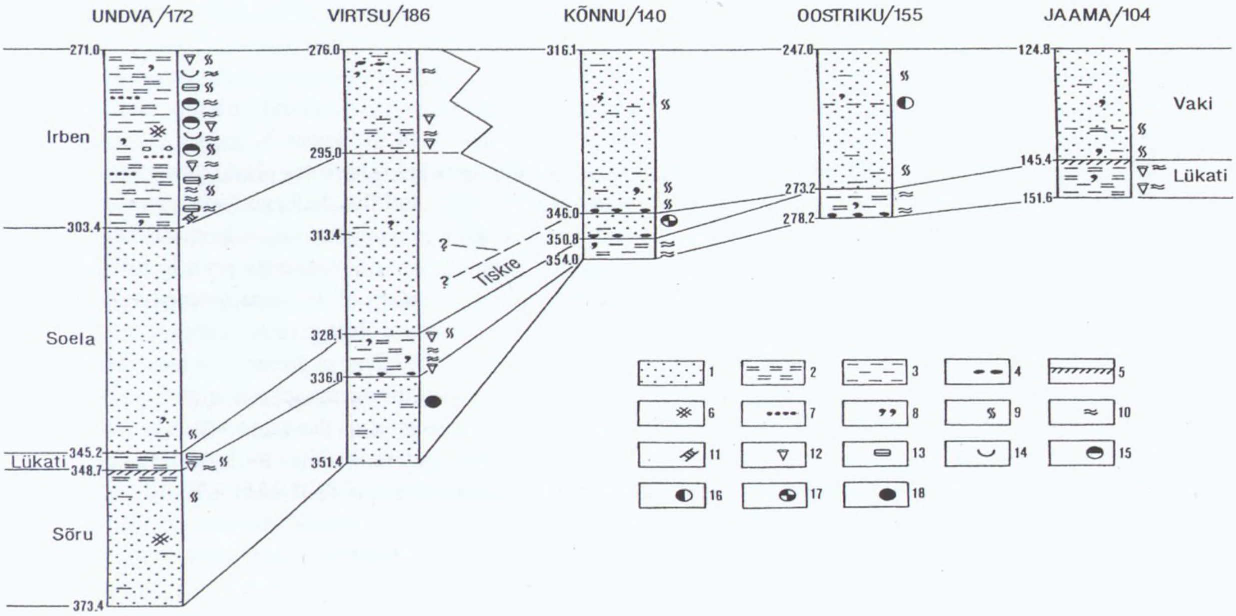


Fig. 19. Correlation of the Dominopol', Ljuboml' and Vērgale sequences between Undva and Jaama, and distribution of the most common fossils: 1 - sandstone; 2 - siltstone; 3 - clay; 4 - pebbles; 5 - traces of weathering; 6 - red-coloured rocks; 7 - ferruginous oolite; 8 - glauconite; 9 - ichnites of the *Skolithos* type; 10 - ichnites of the *Planolites* type; 11 -platysolenitids; 12 - *Volborthella*; 13 - agglutinated foraminifers ("Lycatiella"); 14 - fragments of brachiopods; 15 - Vērgale acritarch assemblage; 16 - mixed Vērgale-Rausvé acritarch assemblage; 17 - pauperized Lükati acritarch assemblage; 18 - Lükati acritarch assemblage. The numbers of the boreholes correspond to those in Figs. 17, 20.

a unit containing the Tiskre beds (= Diplocraterion Sandstone) by Öpik (1933) or the Tiskre Formation by Männil (Мянниль 1958a) together with the Kakumägi beds (*Scenella* Zone) by Öpik (1933) or the Kakumägi Member of the Pirita Formation by Männil (Мянниль 1958).

The Tiskre Formation is distributed in northern and western Estonia (Figs. 18, 19) where it overlies the Lükati Formation. The lower boundary is lithologically abrupt and marked by a change from argillaceous rocks to sandstones. Between Muraste and Aseri, lenses of *Mickwitzia* conglomerate occur at that level. Its type section is an exposure at the southern end of the Rannamõisa Cliff, 14 km west of Tallinn (Менс и Пиррус 1977). The formation is at its thickest (20 m) in the stratotype area.

The Tiskre Formation consists of light-coloured massive or thick-bedded sandstones with thin interbeds of greenish-grey argillaceous rocks (Photo 16). Within the outcrop belt the Tiskre Formation is divided into the Kakumägi (lower) and Rannamõisa (upper) members.

The Kakumägi Member is represented by poorly sorted sandstones with an admixture of clayey material. Conglomerate lenses within the sandy sequence are locally present. In

Photo 16. Klint at Tiskre. The arrow shows the boundary between the Tiskre and Kallavere formations and the Pakerort Stage. *Photo by A. Rõõmusoks.*

the basal part, sandstones are often well cemented with dolomitic cement. Bedding is mostly lenticular, casts of mud cracks, slump-rolls and ripple marks are common.

The Rannamõisa Member consists of horizontally bedded winnowed fine-grained or very fine-grained sandstones with thin interlayers of argillaceous rocks. Glauconite is present. Ripple marks and convolute bedding occur throughout the member (Пиррус 1978).

Ljuboml' Stage

The succeeding Aisčiai evolutionary stage terminates the Early Cambrian sedimentation in Estonia, and embraces deposits of the Ljuboml' and Vērgale stages (Table 6). The former was earlier treated as a part of the Vērgale Stage (Mens *et al.* 1990) or as the Lower Vērgale Substage (Решение... 1986). Kirjanov (Кирянов 1969) distinguished it in the rank of an independent stage and also as a formation with the stratotype section in the interval of 543.5 to 617.2 m of the Berezhki-2944 drill core in Volyn, the western Ukraine.

The Ljuboml' Stage is represented in western Estonia by the Soela Formation and in central and eastern Estonia by the Vaki Formation (Решения... 1978, Менс 1979, Кала и др. 1984a), and is known only from core sections. Since the boundary between the Ljuboml' Stage and the overlying Vērgale Stage is difficult to determine, a map showing the distribution of the late Lower Cambrian rocks in Estonia was compiled jointly for these units (Fig. 20).

The Ljuboml' Stage lies with a stratigraphic unconformity on the rocks of the Dominopol' or Lontova Stage or on the crystalline basement (Ruhnu Island, borehole 257). Accordingly, the lower boundary is expressed variously. It is well recognizable in the areas where the underlying strata consist of crystalline or argillaceous rocks. In the latter case, the topmost part of the Lontova or Lükati Formation is often weathered. Identification of the lower boundary is most complicated in the sections where the Tiskre Formation of the Dominopol' Stage and the Soela or Vaki Formation of the Ljuboml' Stage occur simultaneously, because they are lithologically very similar and extremely poor in fossils. In that case, the lower boundary of the stage is tentatively drawn at the level of the essential change in the composition of detrital minerals (Менс 1979). Due to the general lack of fossils, the boundary between lithostratigraphical units is conditionally taken for the upper limit of the stage which is placed at the base of the Irben Formation. Since the boundaries of the stage have not been firmly fixed, its thickness is difficult to determine, but it is mostly less than 40 m.

Fossils are extraordinarily rare. Occasionally, casts of *Volborthella*, undeterminable fragments of inarticulate brachiopods, valves of agglutinated foraminifers and some ichnofossils, mostly of the genus *Skolithos* are encountered. Acritarchs have been found from the Soela Formation in the Varbla (Fig 20-*188*) drill core (440.1 m) and from the Vaki Formation in the Oostriku-700 (Fig. 20-*155*) drill core (256.2 m). The Soela Formation contains an abundant and diverse acritarch assemblage prevailed by leiosphaerids. The occurrence of *Goniosphaeridium varium* (Volk.) and *Skiagia ciliosa* (Volk.) among acantomorphids suggests late Early Cambrian age (Келлер и Розанов 1979a). The acritarch assemblage from the Vaki Formation is pauperated, beside leiosphaerida it con-

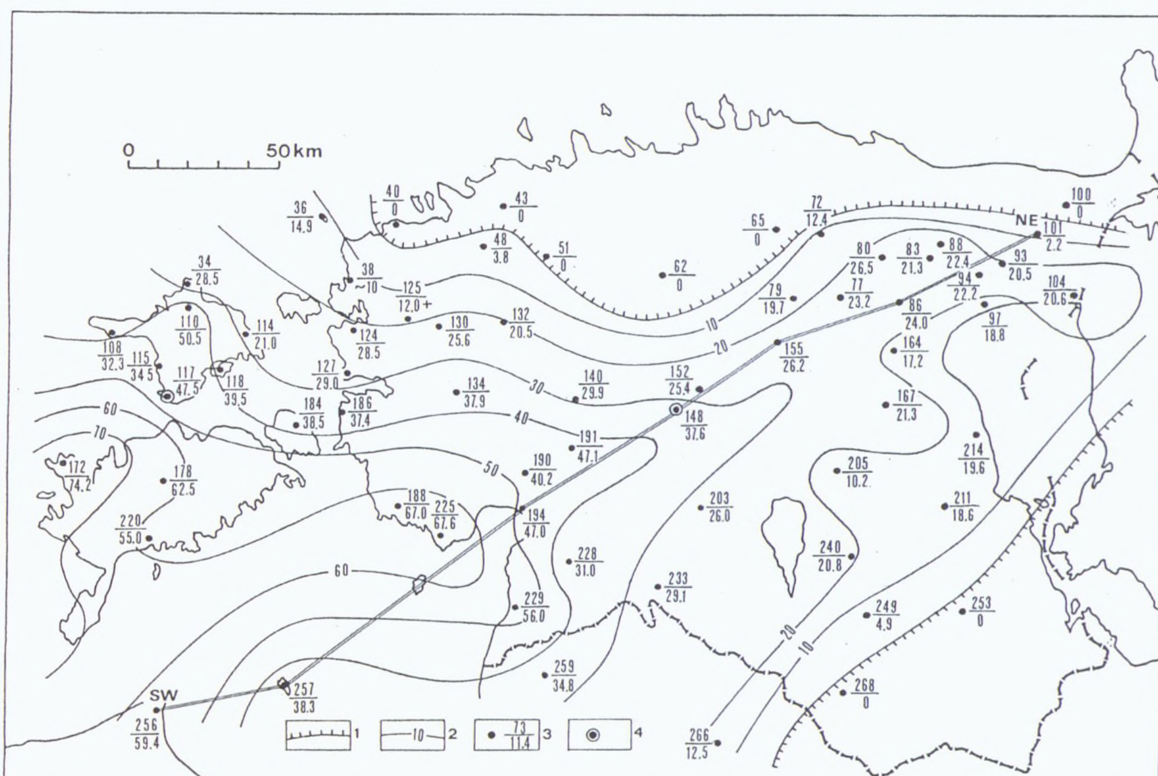


Fig. 20. The distribution and the thickness of the Ljuboml' and Vērgale stages (Aisčiai Group): 1 - limits of present-day distribution; 2 - isopachs; 3 - borehole: the numerator marks the number of borehole (see also Fig. 3) and the denominator of thickness of the stages; 4 - stratotype section. The SW-NE line indicates the transect of the stratigraphical cross-section presented in Fig. 21.

tains a few *Goniosphaeridium volkovae* Hagenfeldt and *Comasphaeridium latviense* (Volk.), and this unit can have a wider biostratigraphic bracketing than the Soela Formation.

The Ljuboml' Stage is regarded as corresponding to the *Holmia inusitata* Zone.

The Soela Formation as an independent stratigraphic unit was distinguished recently (Менс 1979, Кала и др. 1984a). Earlier, this part of the sequence was treated either as the upper part of the Tiskre Formation (Кала 1972, Келлер и Розанов 1979б) or the lower part of the Kurzeme (now Irben) Formation (Менс и Пиррус 1972).

The type section is in the interval of 230.7 to 263.7 m of the Emmaste drill core (Fig 20-*117*), Hiiumaa Island. The formation occurs on the islands of the West - Estonian Archipelago and in the western part of mainland Estonia (Fig. 21). It is known only from core sections, and its lower boundary coincides with the lower boundary of the Ljuboml' Stage (see above). The upper boundary is lithologically clear, only east of the Koluvere - Rumba - Pärnu line it is somewhat debatable. The formation lies transgressively on the Lower Cambrian rocks or the crystalline basement (Fig. 20-*257*). The thickness varies from 7.9 to 41.8 m, increasing from north to south.

The formation consists of weakly cemented light-coloured fine-grained feldspatic (subarkose) sandstones, containing up to 10% of coarse sand and gravel grains. In the lowermost part of the section, mainly beyond the area of distribution of the Tiskre Formation, interbeds of greenish-grey siltstones and argillaceous rocks occur. Pebbles of greenish-grey clayballs and cross-bedding marked by mica flakes and glauconite grains are characteristic of the lower part of the formation.

Vērgale Stage

The Vērgale Stage, embracing the topmost Lower Cambrian in Estonia, was established in the present stratigraphical extent by Birkis *et al.* (Биркис и др. 1970). Previously, it had been used in a wider extent (Келлер и Розанов 1979б, Решение... 1986). The Vērgale-46 drill core in the interval of 1293 to 1318 m has been selected as the stratotype section of the stage (Биркис и др. 1970, Брангулис 1989). Only the lowermost part of the stage occurs in Estonia. The Vērgale Stage is of subsurface distribution on the islands of Hiiumaa and Saaremaa and in the western part of mainland Estonia (Fig. 20). Its thickness decreases northwards being about 40 m at Seliste and less than 1 m at Tahkuna (Fig. 20-*225, 34*).

The lower boundary of the stage is tentatively drawn at the level of the base of the Irben Formation, near the level of the appearance of the Vērgale acritarch assemblage. It is less expressed beyond the distribution area of the Irben Formation where the whole post-Liivi sandy Lower Cambrian succession has been distinguished as a joint Ljuboml'-Vērgale unit or as the Aisčiai Group.

The faunal record of the Vērgale Stage includes *Volborthella*, agglutinated foraminifers, fragments of trilobites and inarticulate brachiopods, also ichnites are very common. The so-called Vērgale acritarch assemblage comprises *Estiastra minima* Volk., *Skiagia ciliosa* (Volk.), *S. insigne* (Fridr.), *S. compressa* (Volk.), *S. orbiculare* (Volk.), *Comasphaeridium strigosum* (Jank.), *Asteridium spinosum* (Volk.), *A. lanatum* (Volk.), *A. tornatum* (Volk.), *Tasmanites volkovae* Kirjanov, *T. bobrowskae* Waz., *T. tenellus* Volk., *Dictyotidium priscum* Kirjanov & Volk., *Leiovalia tenera* Kirjanov, *Pterospermella solida* Volk., *Lophosphaeriduim truncatum* Volk., *Alliumella baltica* Vanderflit, *etc.* with several variations in the species composition from section to section. On the ground of the palaeontological evidence, the stage corresponds to the *Holmia kjerulfi* Zone.

The Irben Formation (Кала и др. 1984a) was earlier termed the Kurzeme Formation (Менс и Пиррус 1972). The stratotype of the formation is the type section of the previous Kurzeme Formation represented by the interval of 1329 to 1387 m of the Pavilosta drill core in Latvia (Лиелдиена и Фридрихсоне 1968).

The Irben Formation is distributed on the islands of the West-Estonian Archipelago and in the western part of mainland Estonia (Figs. 19,20), and is known only from core sections. It rests conformably on the Soela Formation having the maximum thickness (42.4 m) in the Seliste drill core (Fig. 20-*225*). The lower boundary of the formation is marked by the appearance of argillaceous rocks. The upper boundary of the formation is erosional throughout its distribution area in Estonia.

The formation consists of interbedded clay- and siltstones with interlayers of fine-grained sandstones, the number and thickness of which increase eastwards. Brown ferruginous oolith interbeds of goethite oolites are characteristic of the formation; in Estonia, these have been found only in the

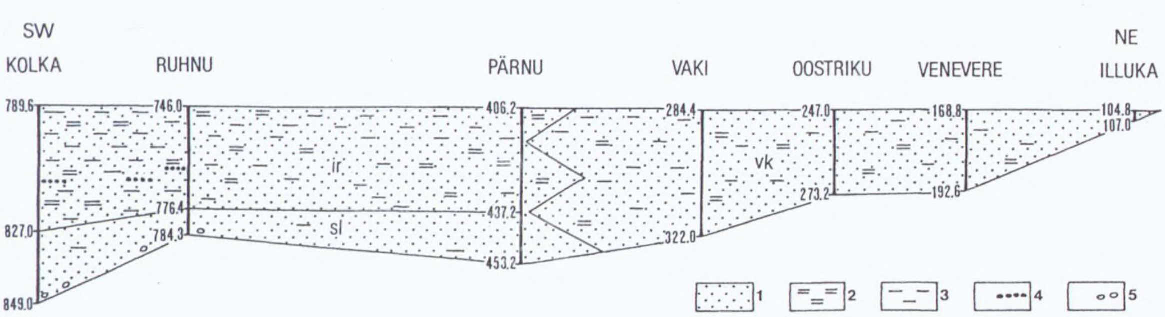


Fig. 21. Cross-section along the SW-NE line of the Aisčiai Group (the Ljuboml' and Vērgale stages): 1 - sandstone; 2 - siltstone; 3 - clay; 4 - ferruginous oolite; 5 - gravel and pebbles. Continuous line indicates the boundary of formations, zigzag line - facial transition. Indices: sl - Soela Formation, ir - Irben Formation, vk- Vaki Formation.

westernmost sections (Figs 19, 21). Owing to the unique lithology and wide distribution of argillaceous rocks, it is an important level for regional lithostratigraphical correlation (Пиррус 1986). Argillaceous rocks are greenish-grey, but in the upper part they are locally dark-grey with a purplish or brownish shade of colour. Besides the wavy horizontal bedding, there are abundant ichnites in clay intervals responsible for a bioturbated structure of the "kråksten" type.

Silt- and sand-size fractions contain quartz (up to 85%), feldspars (up to 20%) and micas (usually 2-10%, rarely up to 25%). Like in the Soela Formation, the content of heavy minerals is low. Clay minerals are dominated by illite.

The Vaki Formation occurs in central and eastern Estonia (Table 6), being probably a shallow-water equivalent of the Soela and Irben formations. It is known only from core sections (Fig. 21). The formation rests with a stratigraphic unconformity on rocks of the Dominopol' or the Lontova Stage. In the latter case, or if the Tiskre Formation is absent, the topmost layers of the underlying units are weathered (Менс и др. 1984).

The Vaki-67 (Fig. 21 - *148*) drill core in the interval of 284.4 to 322.0 m has been selected as a type section for the formation (Решения... 1978). The maximum thickness of the formation (more than 35 m) has been registered in its type area (Fig. 20).

The Vaki Formation consists of weakly cemented glauconite-bearing light-coloured fine-grained or very fine-grained sandstones, or of both, with thin interlayers of greenish-grey and bleached-purplish argillaceous rocks. On some levels the latter are highly micaceous and often contain ichnofossils of *Skolithos* affinites filled with fine sandy material. Besides ichnites, the formation contains scarce fragments of inarticulate brachiopods and in the Oostriku-700 (Fig. 20-*155*) drill core at the depth of 256.2 m also a pauperized acritarch assemblage (further above- the Vẽrgale Stage).

MIDDLE CAMBRIAN

Compared to the Lower Cambrian, the Middle Cambrian is of more limited distribution and known only in the subsurface occurrence (Пиррус 1991).

Due to the lack of fossils and a very low core yield, the Middle Cambrian deposits are stratigraphically and sedimentologically poorly studied. The Middle Cambrian age of rocks was proposed according to their geological setting between the palaeontologically characterized Lower and Upper Cambrian rocks, and justification has been derived from the correlations with adjacent areas where the corresponding rocks contain fossils. Under such conditions, it is not always possible to identify the Middle Cambrian boundaries with a certainty. The lower limit of the Middle Cambrian is taken at the base of a thin band formed of coarse-grained non-glauconitic sandstones with gravel and quartz pebbles overlapping the Lower Cambrian strata or the crystalline basement. Notable is an essential change in the mineral composition of the rocks, marked by the disappearance of glauconite, increase in the degree of maturity, prevalence of allothigenous minerals in the group of heavy minerals, *etc.* In southearn Estonia, the lower boundary is drawn on top of the whethering crust of the Lontova Stage. The identification of the Middle Cambrian is most complicated in the sections situated on the northern margin of the Middle Cambrian distribution area where it is underlain by lithologically similar Lower Cambrian light-coloured quartzose sandstones (Менс и Пиррус 1992).

Throughout the distribution area, the Middle Cambrian consists of siliciclastic rocks dominated by mature light-coloured well-sorted non-glauconitic quartzose sandstones.

The Middle Cambrian succession in Estonia is subdivided into the Ruhnu and Paala formations (Table 6). The former is a local equivalent of the Deimena Formation in western Latvia (Решение... 1986, Менс и др. 1984, Пиррус 1991) and, according to fossil evidence, corresponds to the *Ptychagnostus praecurrens* Zone (Менс и др. 1987, Hagenfeldt 1989b, Mens *et al.* 1990).

The Paala Formation is distributed mainly in the southeast of Estonia and is tentatively interpreted as an equivalent of the Sablinka Formation of the Leningrad and Pskov regions (Mens *et al.* 1990). On the basis of palaeontological data, the Sablinka Formation is referred to the *Paradoxides paradoxissimus* and *P. forchhammeri* zones Judging from the mineral composition, the Middle Cambrian succession in some sections, including Abja and Otepää (Fig. 22-*233, 249*), is not clear.

The Ruhnu Formation is distributed in the southwestern part of Estonia, while its range in the middle of southern Estonia is plausible. It lies transgressively on the Lower Cambrian and is overlain by the Upper Cambrian or Lower Ordovician rocks.

The Ruhnu drill core in the interval of 706.8 to 748 m (Fig. 22-*257*) has been selected as the type section for the Ruhnu Formation (Кала и др. 1984a). The maximum thickness of the formation (41.2 m) has been established in this borehole and it decreases towards the north.

The Ruhnu Formation is represented by light well-sorted, fine- to very fine-grained quartzose sandstones with only a few thin interbeds of dark-grey, in the lower part sometimes variegated, argillaceous rocks. The basal part is often marked by a layer of very coarse-grained sandstone containing over 10% of quartz with gravel or pebble grain-size, or with both. The top of the formation is usually cemented by carbonates and pyrite. Often an admixture of feldspars and micas (muscovite) occurs. Glauconite is lacking. Heavy mineral assemblage is characterized by the prevalence of transparent minerals dominated by zircon and tourmaline. Among opaque minerals, detrital leucoxene and ilmenite occur in almost equal quantities or the former prevails slightly. Clay minerals are dominated by illite, the content of kaolinite often reaches 30-35%. Chlorite is rare reaching occasionally 10% in the lower part of the formation.

The Paala Formation occurs in central and southeastern Estonia transgressively on the crystalline basement or on the Lower Cambrian rocks The type section is the Viljandi drill core (Fig. 22-*203*) in the interval of 409.8 to 434m (Менс и др. 1984). The thickness of the formation varies greatly. Due to the low core yield, the determination of the boundaries is complicated and the thickness of the formation is unclear.

The formation consists of light quartzose non-glauconitic middle- to fine-grained sandstones with pellets of white kaolinitic clay. The grain-size of the rocks in the lower and

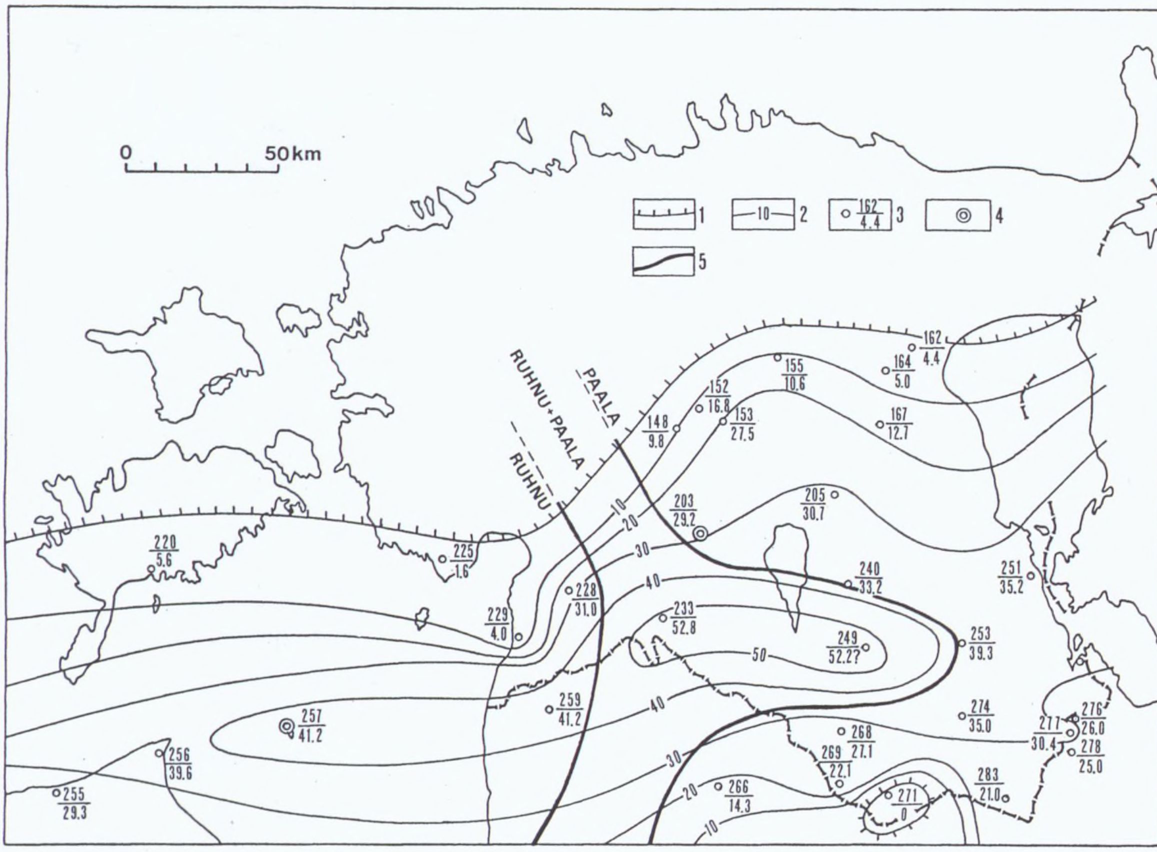


Fig. 22. The distribution and the thickness of the Middle Cambrian: 1 - limits of the present-day distribution of rocks; 2 - isopachs; 3 - borehole: the numerator marks the number of borehole (see also Fig. 3), the denominator shows the thickness of the rocks; 4 - stratotype section; 5 - contour of the transitional area of formations.

upper parts of the formation is coarser than in its middle part where very fine- to fine-grained varieties dominate. Feldspar and muscovite are not common. The heavy mineral assemblage is mainly composed of allothigenic minerals dominated by ilmenite. In the group of heavy transparent minerals, zircon is always the index mineral.

The character of the basal bed depends on the type of underlying rocks. On the crystalline basement, it consists of conglomeratic sandstone, locally with phosphatized pebbles (Laanemetsa, Fig. 22-*269*). Resting upon the argillaceous Lower Cambrian rocks, which are often weathered, the basal bed occurs like a variegated kaolinitic clay comprising coarse grains of quartz redeposited from the weathering crust.

UPPER CAMBRIAN

During the last two decades, the stratigraphic extent of the Estonian Upper Cambrian and its biostratigraphical subdivision has been established more precisely due to the progress in research on acritarchs, conodonts and lingulates (Kaljo *et al.* 1986, Mens *et al.* 1993, Волкова и др. 1981, Попов и др. 1989, *etc.*).

The Upper Cambrian rocks have been palaeontologically documented in two isolated areas - in northern and south-eastern Estonia (Fig. 23). They crop out along the Baltic-Ladoga Klint and in the river valleys crossing it. The main localities in Estonia are Ülgase, Valkla, Turjekelder and Suurjõgi.

In Estonia, the Upper Cambrian strata are distributed sporadically and dominated by sandstones, less than 20 m in thickness. Argillaceous rocks are of limited distribution forming grey to greenish-grey interlayers within light-coloured sandstones in the lower part of the succession and brownish-grey varieties in the upper part.

The Upper Cambrian succession is condensed and interrupted by several minor and major hiatuses. The Upper Cambrian biostratigraphy and interregional correlation are usually based on trilobites. In the Estonian Upper Cambrian sections, where trilobites have not been found, it is based on acritarchs, conodonts and lingulates (Попов и др. 1989, Mens *et al.* 1993). Altogether five acritarch-, four conodont-, and at least three lingulate-based biostratigraphic units are distinguished (see Mens *et al.* 1993).

The oldest Estonian Upper Cambrian acritarch assemblage recorded from the Petseri Formation (SE Estonia) is similar to the acritarch assemblage BK1 б by Volkova (Волкова 1990, Волкова и Кирянов 1995), except the occurrence of *Leiofusa stoumonensis* Vang, which has been found in the base of the acritarch-bearing part of the Petseri Formation. Based on the occurrence of the genera *Stelliferidium, Cymatiogalea, Leiofusa* and *Veryhachium*, which appeared in *Olenus* time (Potter 1974, Downie 1984), and the absence of the *Impluviculus* species, the rocks comprising this acritarch as-

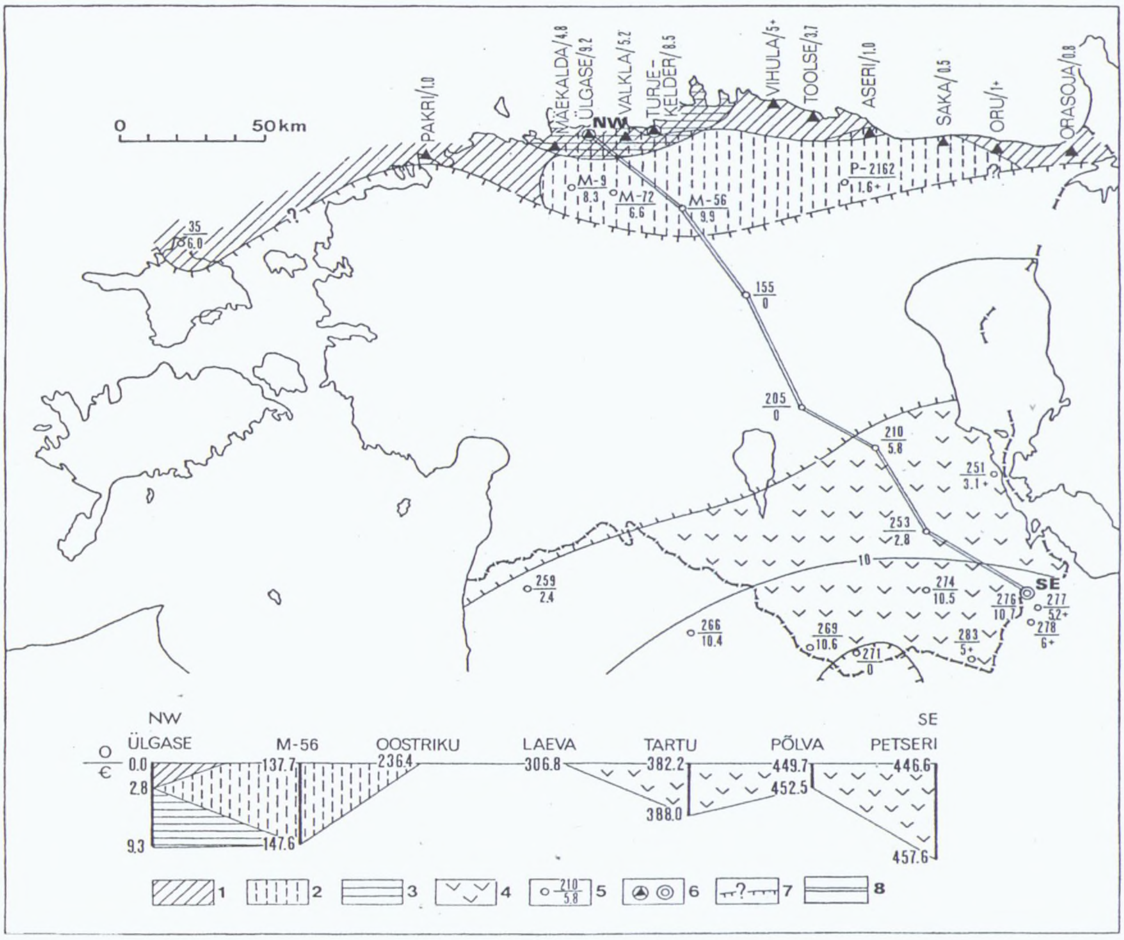


Fig. 23. Geographical extension of the palaeontologically determined Upper Cambrian rocks: 1 - lower part of the Kallavere Formation; 2 - Tsitre Formation; 3 - Ülgase Formation; 4 - Petseri Formation; 5 - borehole: the numerator marks the number of borehole (see also Fig. 3), the denominator shows the thickness of the rocks; 6 - stratotype section: outcrop, borehole; 7 - limits of the present-day distribution of rocks; 8 - line of cross-section.

semblage are considered as a stratigraphic equivalent of the lower or middle, or of the both parts of the *Olenus & Agnostus* Zone.

The next acritarch assemblage in the Ülgase Formation (Table 6) is similar to that described above, but differs in the presence of *Veryhachium dumontii* and representatives of the genus *Impluviculus* and in the lack or restricted distribution of typical Middle Cambrian species. As the appearance of the genus *Impliviculus* has been correlated with the uppermost part of the *Olenus* & *Agnostus* Zone (Downie 1984) and with the lowermost part of the *P. spinulosa* Zone (Martin & Dean 1988), we regard the Ülgase Formation tentatively as a stratigraphic equivalent of the uppermost *Olenus* & *Agnostus* and the lowermost *Parabolina* zones (Table 6).

The third Upper Cambrian acritarch assemblage in the Estonian succession is distinguished by the appearance of *Trunculumarinum revinium* (Vang.) Loeblich *et* Tappan, *Dasydiacrodium caudatum* Vang. corresponds to the microflora A4, *i.e. T. revinium - D. caudatum* assemblage *sensu* Martin and Dean (1988), to the assemblage BK-3 *sensu* Volkova (1990), to the uppermost *P. spinulosa* Zone and to the *Leptoplastus* Zone, as a whole (Table 6).

The next acritarch assemblage, showing a high taxonomic diversity and variation in different sections, contains a significant amount of diacroids and sometimes also endemic forms. On the East-European Platform, this assemblage was first described from the upper part of the Ladoga Formation distributed in the Leningrad Region (Волкова и Голуб 1985) and is referred to as BK-4. In Estonia, the assemblage occurs together with conodonts of the *Proconodontus* Subzone in the Tsitre Formation, but these acritarchs have also been found in the *Cordylodus andresi* Zone in the Kallavere Formation. A relatively similar assemblage together with the trilobites of the *Peltura scarabaeoides* Zone has been determined from the Degerhamn section of southern Öland (Di Milia *et al.* 1989).

The youngest Upper Cambrian acritarch assemblage, which also contains an abundance of diacroids (Волкова 1989, 1990), can be distinguished by the appearance of *Acanthodiacrodium angustum* (Downie) Combaz and *Dicrodiacrodium ramusculosum* (Combaz) Volkova. It occurs together with conodonts of the *C. proavus* Zone, and may be also characteristic of the *C. intermedium* Zone (Волкова и Менс 1988).

Conodonts have been studied from a number of outcrops and drill cores. The number of specimens is small, representing mostly the genera *Phakelodus, Furnishina, Prooneotodus* and *Westergaardodina*. Euconodonts (*Proconodontus, Eoconodontus* and *Cordylodus*) have been found only in the topmost Upper Cambrian. According to the conodont zonation worked out by Sergeyeva and Viira for the Baltic-Ladoga Klint area, the Upper Cambrian sequence is subdivided as follows (from below upwards): the *Westergaardodina* Zone with *W. bicuspidata*, *W. moessebergensis* and *Proconodontus* subzones, and the *Cordylodus andresi* and *C. proavus* zones (Kaljo *et al.* 1986).

Lingulate brachiopods are represented in the Upper Cambrian of Estonia by lingulids and acrotretids. Following the brachiopod zonation suggested by Popov and Khazanovitch (Попов и др. 1989) and adopted by Puura and Holmer (1993), four brachiopod zones can be distinguished. From below upwards these are the *Ungula inornata* Zone, the *Ungula convexa* Zone, the *Ungula ingrica* Zone and the *Obolus apollinis* Zone. The first three zones belong to the Upper Cambrian, while the latter one can belong partly to the Ordovician, depending on the position of the lower boundary of the Ordovician System.

The Estonian Upper Cambrian includes three succeeding lithological units in the rank of formation (Petseri, Ülgase, Tsitre; Table 6), and the lower part of the overlying Kallavere Formation.

The Petseri Formation introduced by Kajak (Каяк 1967) in the rank of beds is known only from core sections in south-eastern Estonia, from where it spreads east- and southwards. In the Petseri borehole in Russia (Fig. 23), serving as the type section, the formation is 10.7 m thick. In complete sections the Petseri Formation can be subdivided into three parts. The lower and upper parts are represented by light-coloured weakly cemented quartzose sandstones, whereas the middle part is predominantly composed of grey argillaceous rocks (Волкова и др. 1981). Sandy parts of the formation contain some glauconite and debris of inarticulate brachiopods. In the argillaceous part, shells and fragments of lingulates of the genera *Ungula* and *Oepikites* and the *Dasydiacrodium setuensis* - *Leiofusa stoumonensis* assemblage of acritarchs occur (Волкова и др. 1981, Волкова 1990, Paalits 1992a). On the ground of this acritarch assemblage, the Petseri Formation, or at least its middle part, is considered as a time equivalent of the lowermost part of the *Olenus* & *Agnostus* Zone (Table 6).

The **Ülgase Formation** was referred by Öpik (1929) to the *Acrotreta* Sandstone and assigned to the local *Acrotreta-Lingulella* Zone. Subsequently, this part of the succession was defined as the Ülgase Member in the limits of the Pakerort Stage (Мююрисепп 1958). In the rank of formation it was first considered by Khazanovich and Missarzhevsky (Хазанович и Миссаржевский 1982). The formation with a thickness of about 10 m is better fixed in the vicinity of Tallinn and within some 50 km east and south of it. The Ülgase Formation consists of light-coloured very fine- to fine-grained sandstones with interbeds and lenses of greenish-grey clay in the lower part and brownish-grey thin films in the upper part. Its upper boundary is transitional and in the earlier papers the lower part of the overlying Tsitre Formation was regarded as belonging also within this formation (Менс 1984, Попов и др. 1989). The formation contains numerous lingulates of the genera *Ungula* (including *U. inornata*), *Oepikites, Angulotreta* and *Cerotreta*. The occasional conodonts belong to the genera *Phakelodus*, *Furnishina* and *Prooneotodus* (Kaljo *et al.* 1986, Mens *et al.*, 1993). *Torellella sulcata* Missarzhevsky abound. In the argillaceous interlayers the acritarchs, forming the *Impluviculus multiangularis* - *Veryhachium dumontii* assemblage are numerous (Волкова 1982, 1990, Волкова и Менс 1988). On the basis of fossil evidence, the Ülgase Formation is assigned to the uppermost part of the *Olenus* & *Agnostus* Zone and to the lower part of the *Parabolina spinulosa* Zone (Table 6).

The Tsitre Formation was introduced by Popov and Khazanovich (Попов и Хазанович 1985) with the stratotype in the Turjekelder section. Earlier, this part of the succession belonged to the Kallavere Formation (Kaljo *et al.* 1986; Решение... 1986). Currently, the Tsitre Formation (Table 6) includes also the underlying beds containing kerogen-bearing argillaceous interlayers and differing in fossil record from the Ülgase Formation (Mens *et al.* 1993).

The Tsitre Formation expands as a narrow belt from Tallinn to Kohtla-Järve (Fig. 23). Its thickness in the outcrop sections is a bit more than 3 m. In the drill core sections its thickness is unclear due to the low core yield, but probably it is less than 10 m.

The formation is typically represented by light-grey weakly cemented fine-grained quartzose sandstones, with a few thin interbeds of variegated, dominantly brownish-grey clayey rocks. These interlayers are often accompanied by bedding planes covered with convex-up lingulate shells. This has been considered in drawing the boundary between the Ülgase and Tsitre formations.

The co-occurrence of *Trunculumarinum revinium* and *Dasydiacrodium caudatum* in the lower part of the Tsitre Formation suggests that these deposits belong to the upper part of the *Parabolina spinulosa* Zone (Martin & Dean 1988, Paalits 1992b). The upper part of the Tsitre Formation contains a rather rich fossil record and its relationship with the trilobite zones is shown in Table 6.

The uppermost part of the Cambrian is considered to be represented by the lower part of the Kallavere Formation. The latter is discussed with the rest of this formation within the Ordovician.

ORDOVICIAN

Introduction

In the Ordovician, epicontinental seas with extensive distribution of carbonate sediments had a greater extent than in any other period. The marine flora and fauna changed markedly in the course of the Ordovician. A number of major taxonomic groups (bryozoans, brachiopods, echinoderms, trilobites, ostracodes, chitinozoans and others) appeared or became common. In this respect, the Ordovician is one of the most interesting periods in the history of marine faunas, and Estonia is among the areas in the world where this fauna is well preserved and studied.

The Ordovician was characterised by an extreme biogeographical differentiation of both planktic and benthic faunas, but with different degree. This makes the worldwide correlation of the Ordovician rocks difficult and has resulted in numerous regional stratigraphic schemes. A series of detailed stratigraphical charts compiled for the East Baltic (see Решения... 1978, 1981, Männil & Meidla 1994, and literature cited in these papers), gave a relatively stable detailed local classification for Ordovician rocks and afterwards obtained the status of a regional standard for most of the East-European Platform (Männil 1990).

The large-scale biogeographical and facies differentiation within the Ordovician Palaeobasin of Baltoscandia is well expressed in the concept of confacies belts (Jaanusson 1976, Fig. 24). The territory of Estonia is divided between the North Estonia and Central Baltoscandian confacies.

An emended version of the correlation chart, presented for the Estonian succession ranging from 70 to 180 m in thickness (Table 7), is based mainly on the above-cited schemes. For practical reasons the present version is simplified and many smaller subdivisions have been omitted. Some subdivisions, defined earlier as formations due to overestimations or difficulties in their specifications, are treated as members. Often the lithounits (Table 7) have diachronous (wavy line) or topical (discontinuous line) boundaries or the unit serves as "topostratigraphic" unit (*sensu* Jaanusson 1976, p. 310) with their boundaries coinciding with the stage boundaries.

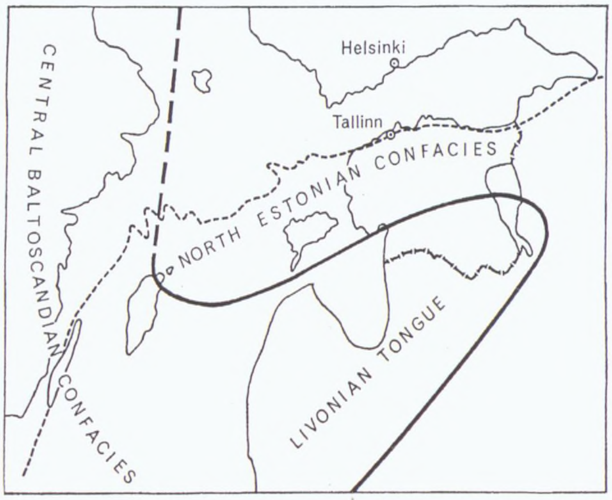


Fig. 24. Approximate boundary of the Ordovician confacies belts. Thin dashed line - the northern limit of the continuous distribution of rocks.

OELAND SERIES

Pakerort Stage

The lowermost Ordovician Pakerort Stage (Table 7) distinguished by Raymond (1916) consists of two different lithotypes - the *Obolus* Sandstone (the Kallavere Formation, Мянниль и Рыымусокс 1984) and *Dictyonema* Shale (Türisalu Formation, Мююрисепп 1958, 1960б). During the last decades, both lithotypes have been studied in particular detail in terms of their industrial use and potential environmental impact. In northern Estonia, the so-called *Obolus*-Conglomerate (brachiopod coquina) occurs usually at the base of the *Obolus* Sandstone. The coquina was used as a good lithological marker in fixing the lower boundary of the Pakerort Stage and the Cambrian/Ordovician boundary in Estonia.

The Cambrian/Ordovician boundary (Photo 17) became an object of special international studies in the 1970s, and since then several different stratigraphical levels have been proposed as the lower boundary of the Ordovician system. According to most of stratigraphers, the definition of the Cambrian/Ordovician boundary should be based on conodonts and the horizon chosen should be close to, but below the lowest planktic (nematophorous) graptolites. Three biostratigraphic horizons of conodonts were considered as possible guides for marking the boundary level. These are the base of *Cordylodus proavus*, of *C. intermedius* and of *C. lindstromi* zones. Currently, the attention is focused on the first appearance of the conodont *Iapetognathus* n. sp. in the lower part of the *C. lindstromi Zone* just above the first appearance of the planktic graptolites *R. praeparabola* and *R. parabola* in the Dayangcha section (China). In Estonia, these conodont zones have been identified in the Kallavere Formation considered preliminarily (Мянниль и Рыымусокс 1984) as the oldest part of the Ordovician sequence. Somewhat later the lower boundary of the system and of the Pakerort Stage was tentatively drawn at the level of the first appearance of *Cordylodus* (base of *C. andresi* Zone, Fig. 25) in the lower part of the Kallavere Formation (Kaljo *et al.* 1986, Решения... 1987, Männil 1990). But if a higher stratigraphical level (*e.g.* the base of the *C. lindstromi* Zone) will be accepted internationally for the boundary between the Cambrian and Ordovician systems, most of the Kallavere Formation must be excluded from the Ordovician (Norford 1991, Miller & Taylor 1995, Fig. 25).

As there are no distinct lithological changes on the boundaries of the conodont zones in most of the sequences, the *Obolus* Sandstone of the Kallavere Formation will be treated below as an entity (Fig. 26). Within that formation the main attention focuses on the distribution of conodonts, zones of which could serve as guides for the boundary between the systems. The following succession of conodont zones (in ascending order) has been established in the *Obolus* Sandstone and *Dictyonema* Shale (Kaljo *et al.* 1986): *Cordylodus andersi*, *C. proavus*, *C. intermedius*, *C. lindstromi* and *C. angulatus*. The *Cordylodus andresi* Zone has been established only in a few sequences in northern Estonia (Kidaste core in Hiiumaa, outcrops at Turjekelder, Vihula and Toolse).

The *C. proavus* Zone occurs in all sections (except Turjekelder) where conodonts have been studied, and in some sections it is rather thick (Fig. 25). The zone is absent in most of the brachiopod coquina (*Obolus*-Conglomerate). The morphological variability of the zonal species of the *C. proavus* Zone suggests that the zone is discontinuous: sometimes the lower, sometimes the upper part is missing. Compared to other conodont zones in the Estonian sequences, the *C. proavus* Zone has the most distinct lower boundary which coincides or occurs close to the lower boundary of the Kallavere Formation.

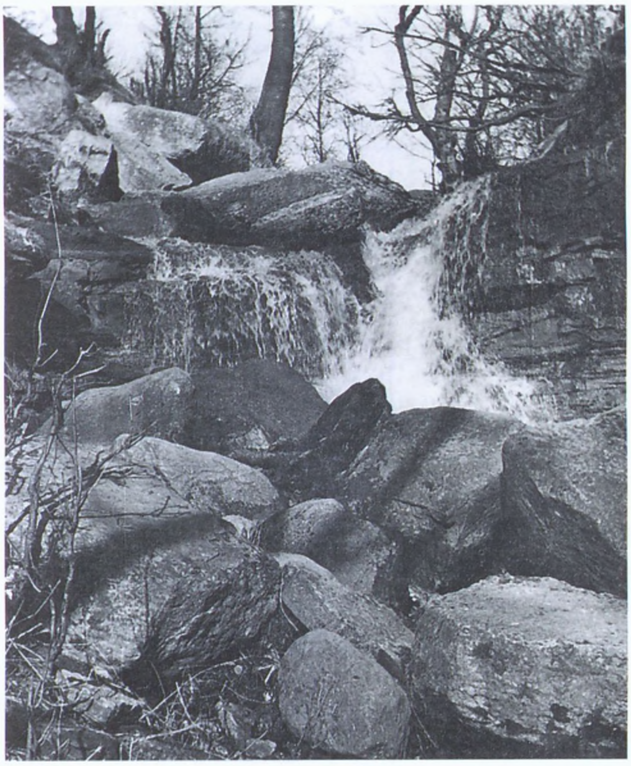


Photo 17. The Cambrian/Ordovician boundary beds around the Hundikuristik Waterfall in Tallinn. *Photo by A. Miidel.*

The *C. intermedius* Zone is of limited distribution in Estonia. The index species has been established only in three sections (Mäekalda in Tallinn, Ülgase, Toolse). The *C. lindstromi* Zone, *vice versa*, is widespread. It is missing only on the Pakri Cape and possibly also in the Vihula section (Fig. 25). The specimens of *C. lindstromi* Druce *et* Jones found in Estonia and Australia (Nicoll 1991), are morphologically very similar (low base, small cavity with one or more pointed secondary tips).

The *C. angulatus* Zone occurs in all northern Estonian sections, except the Pakri Cape. Its lower boundary is well defined by the appearance of numerous specimens of *C. angulatus* Pander apparatus. In the western part of Estonia, it coincides with the lower boundary of the Suurjõgi Member, which is a good lithological marker (Fig. 26).

The co-occurrence of conodonts and graptolites differs considerably from area to area (Kaljo & Viira 1989). The *Rhabdinopora flabelliformis* group makes its first appearance at different levels in relation to the conodont zones. In the vicinity of Tallinn, it occurs in the top of the *C. proavus* Zone or in the *C. intermedius* Zone, east of Tallinn in the *C. lindstromi* Zone, and to the west of it in the *C. angulatus* Zone. This complicates the use of graptolites in the correlation of sections. Nevertheless, the graptolites serve as the most important group of fossils in establishing the upper boundary of the Pakerort Stage which falls into the

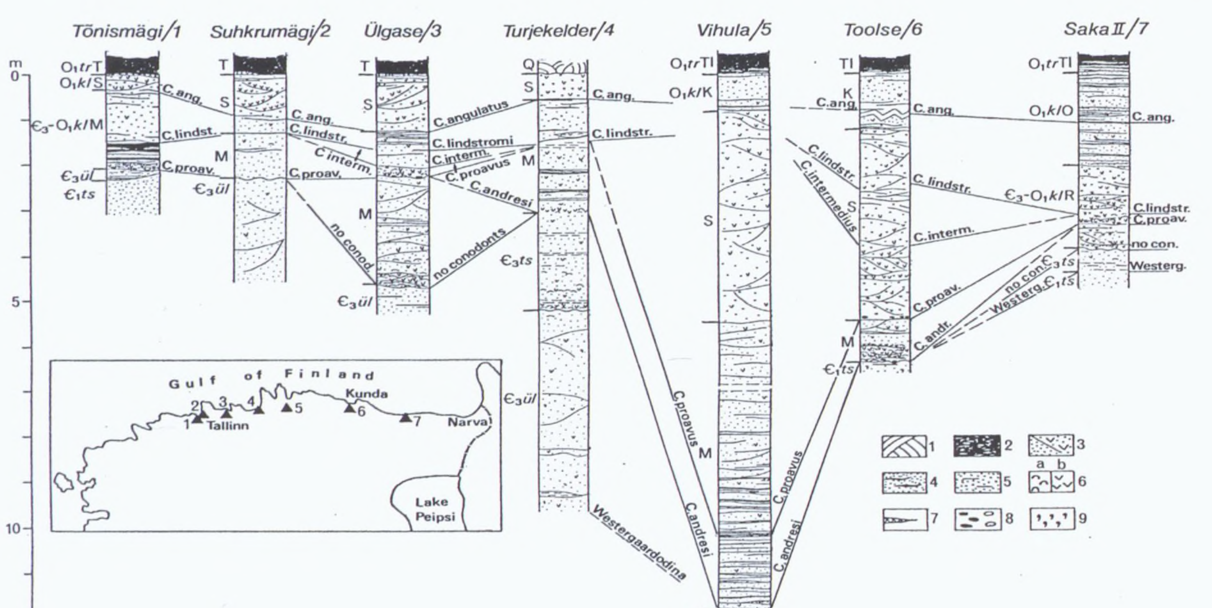
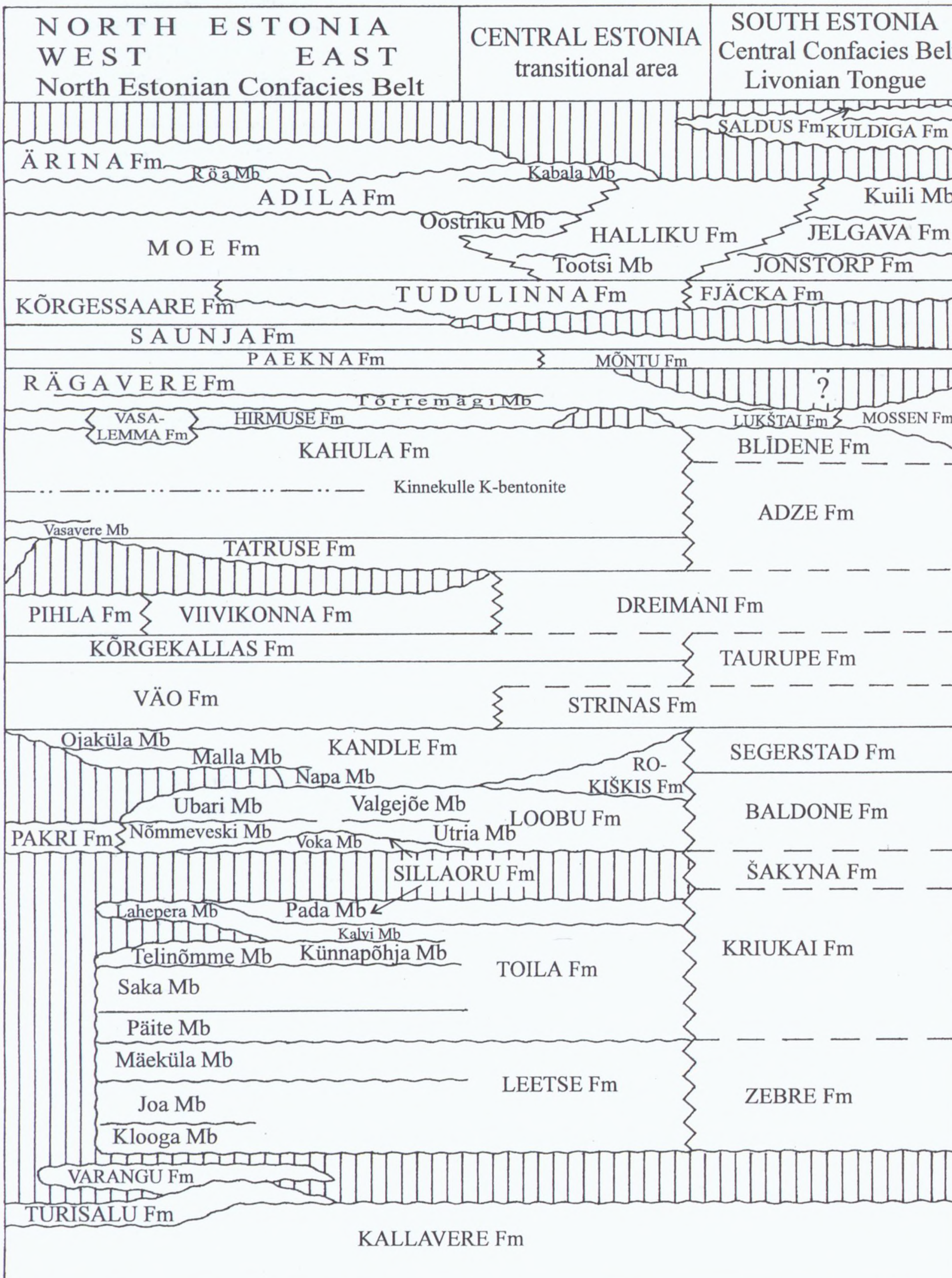


Fig. 25. Conodont zones of Cambrian-Ordovician boundary beds (Kallavere Formation) in North-Estonian outcrops. Lithostratigraphic units: O$_1$*tr*T - Türisalu Formation, Tabasalu Member; O$_1$*tr*Tl - Türisalu Formation, Toolse Member; O$_1$*kl*S - Kallavere Formation, Suurjõgi Member; O$_1$*kl*K - Kallavere Formation, Katela Member; O$_1$*kl*O - Kallavere Formation, Orasoja Member; Є$_3$-O$_1$*kl*M - Kallavere Formation, Maardu Member; Є$_3$-O$_1$*kl*R - Kallavere Formation, Rannu Member; Є$_3$*ts* - Tsitre Formation; Є$_3$*ül* - Ülgase Formation. 1 - Quaternary deposits; 2 - dark argillite *(Dictyonema* Shale); 3 - quartzose sandstone with phosphatic brachiopod fragments, cross-bedded; 4 - quartzose sandstone with dark argillite interbeds; 5 - clay interbeds in sandstone; 6 - phosphatic brachiopods, a) complete valves, b) fragments; 7 - pyrite lenses; 8 - quartzose sandstone pebbles (phosphatic ones marked with dark colour); 9 - glauconite.

Table 7. The Ordovician of Estonia (*compiled by J. Nõlvak*)

British Series	Regional Series	East Baltic stages	substages	Scandinavian graptolite zones	North Atlantic conodont zones and subzones	Baltoscandian chitinozoan zones and subzones
ASHGILL	HARJU	PORKUNI		*persculptus* ?	*ordovicicus*	*scabra* *taugourdeaui*
		PIRGU		*complanatus*		*rugata* *bergstroemi* *barbata*
		VORMSI		*linearis*		
CARADOC		NABALA			*superbus*	*fungiformis*: *reticulifera*, *angusta*
	VIRU	RAKVERE		*clingani*		
		OANDU				
		KEILA				*cervicornis*: *multiplex*
		HALJALA	JÕHVI	*multidens*	*tvaerensis*: *alobatus*	*hirsuta*
			IDAVERE		*tvaerensis*: *gerdae*	*dalbyensis* *curvata* *granulifera*
		KUKRUSE		*gracilis*	*variabilis*	*stentor*: *rhenana*
LLANVIRN		UHAKU		*teretiusculus*	*anserinus*	*tuberculata*
		LASNAMÄGI		*murchisoni*	*serra*: *E. lindstroemi*, *robustus*, *reclinatus*, *foliaceous*	*striata*: *clavaherculi*, *sebyensis*
		ASERI			*suecicus*	
	OELAND	KUNDA	ALUOJA	*artus*	*variabilis*	*regnelli*
			VALASTE			
ARENIG			HUNDERUM	*austrodentatus*		
			LANGEVOJA		*flabellum parva*	*cucumis*
		VOLKHOV	VÄÄNA	*hirundo*	*originalis*	*primitiva*
			SAKA		*navis* *triangularis*	
		BILLINGEN		*elongatus* *densus* *balticus* *phyllograptoides*	*evae* *elegans*	
		HUNNEBERG		*cupiosus* *murrayi*	*proteus*	
TREMADOC		VARANGU		*supremus* *hunnenbergensis*	*deltifer*	no chitinozoans
		PAKERORT		*socialis-flabelliformis* *desmograptoides*	*angulatus* *lindstromi*	

Table 7. The Ordovician of Estonia (continued)



organic-rich argillites of the Türisalu Formation.

Besides the above-mentioned conodonts and graptolites, lingulate brachiopods and acritarchs can be used for the subdivision of the Cambrian-Ordovician boundary beds (Kaljo *et al.* 1986, Puura & Holmer 1993, Mens *et al.* 1993, Paalits 1995).

Kallavere Formation

The quartzose sandstone with interbeds of dark argillite of the Kallavere Formation are distributed almost all over Estonia. It is missing only in a belt running from southwestern to eastern Estonia (Fig. 27). The formation is at its thickest (more than 17 m) in central Estonia. It consists of the Maardu, Rannu, Katela, Orasoja and Suurjõgi members, replacing each other in space or in time (Хейнсалу 1981, 1987, Fig. 26).

Lithologically, the Kallavere Formation is dominated by quartzose sandstone, commonly weakly cemented, with the grain-size of 0.05-0.25 mm. It contains phosphatic brachiopod valves and their fragments, forming a distinct brachiopod coquina layer at the base of the Kallavere Formation (in the Maardu and Rannu members). The thickness of the coquina is only a few cm, except the Maardu (up to 1 m) and Rakvere (up to 4-6 m, sometimes even more) areas. The Kallavere Formation comprises lingulate brachiopods of the *Ungula ingrica* and *Obolus apollinis* zones (Хейнсалу и др. 1987, Mens *et al.* 1993). In the conglomerate bed the most common species is *Ungula ingrica*, accompanied by species of the genera *Schmidtites*, *Keyserlingia* and *Oepikites*.

Dark argillite interbeds, 0.1 mm to 15 cm in thickness, occur generally above the brachiopod coquina. In northwestern Estonia, a 10–30-cm-thick dark argillite (*Dictyonema* Shale) bed with very thin interbeds of sandstone lies immediately on the lower boundary of the formation. In northwestern Estonia, west of the Kunda - Rakvere line, the uppermost part of the formation is represented by the so-called skeletal detritus layer of the Suurjõgi Member which consists of cross-bedded quartzose sandstone comprising brachiopod fragments with a size of 1-3 mm. The thickness of the Suurjõgi Member is about 1 m, except the Toolse - Vihula area where it exceeds 5 m (Fig. 26).

Türisalu Formation

Up to the 1970s, the Türisalu Formation (Мюурисепп 1958а,б, 1960) was considered as the upper part of the Pakerort Stage. The studies of the distribution of graptolites (Кальо и Кивимяги 1970, 1976) allowed to divide the formation between the Pakerort and Varangu stages. The older part is characterized by the occurrence of graptolites of the *Rhabdinopora flabelliformis* Zone and the younger part by the graptolites of the *Kiaerograptus* Zone. West of the

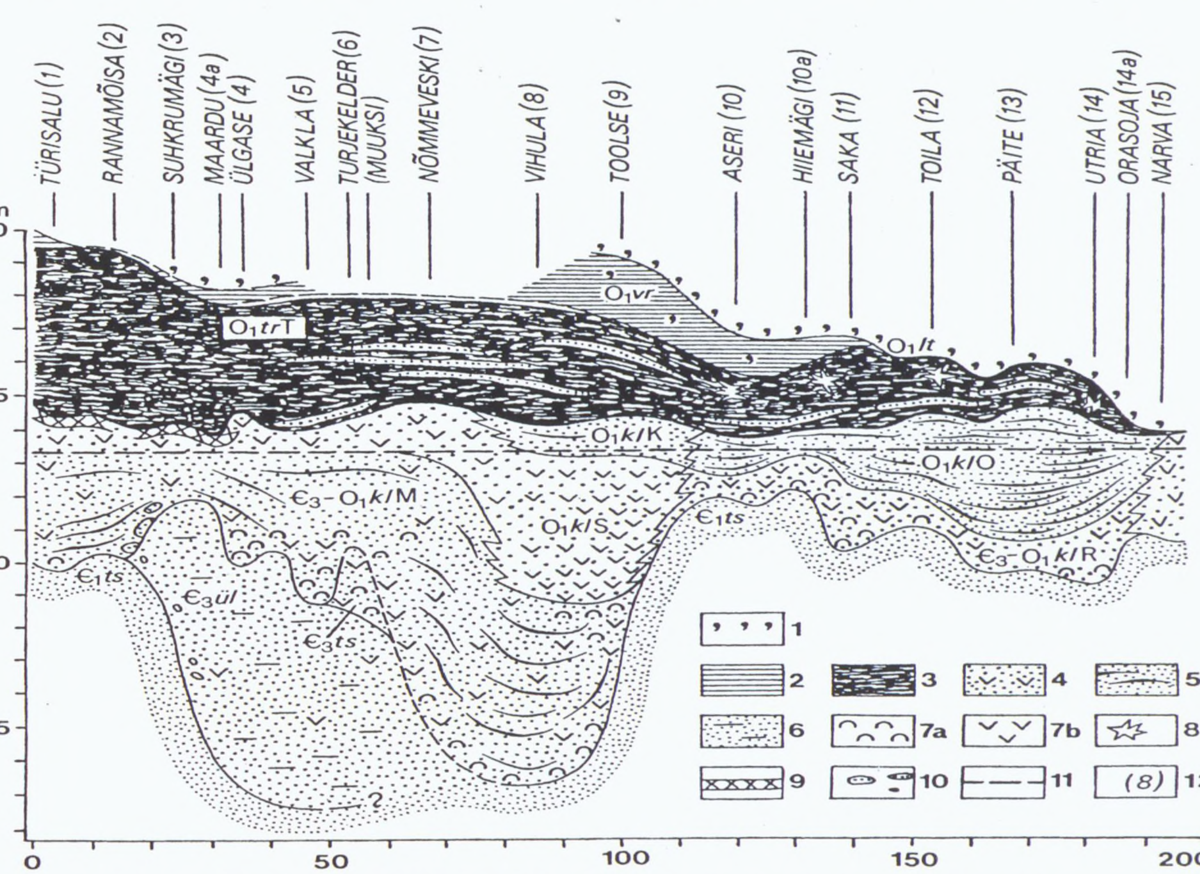


Fig. 26. Cross-section of the Tremadoc in the Baltic-Ladoga Klint outcrops. Formations and members: O$_1$*lt* - Leetse Formation; O$_1$*vr* - Varangu Formation; for explanation of indexes O$_1$*tr*T, O$_1$*tr*Tl, O$_1$*kl*S, Є$_3$-O$_1$*kl*M, O$_1$*kl*K, O$_1$*kl*O, Є$_3$-O$_1$*kl*R, Є$_3$*ts*, Є$_3$*ül*, Є$_1$*ts* see Fig. 25. 1 - glauconitic sandstone; 2 - compact claystone; 3 - argillite (*Dictyonema* Shale); 4 - quartzose sandstone with fragments of phosphatic brachiopods ("*Obolus* Sandstone"); 5 - quartzose sandstone with interbeds of *Dictyonema* Shale; 6 - quartzose sandstone with clayey interbeds; 7 - phosphatic brachiopods: a) complete valves, b) fragments (debris); 8 - anthraconite; 9 - "pyrite layer"; 10 - pebbles, sometimes phosphatized; 11 - lower boundary of the *Cordylodus angulatus* Zone; 12 - number of outcrop (see Figs 27, 28, 29).

Tallinn - Rapla line (Fig. 28) only the older part of the formation is represented. It comprises the *Rhabdinopora f. flabelliformis* (in the lower part) and *R. flabelliformis multithecata* graptolite subzones (Кальо и Кивимяги 1970, 1976), and the upper part of the *C. angulatus* conodont Zone. The most complete stratigraphical sequences of the Türisalu Formation occur between Tallinn and Tapa where the lower part of the formation belongs to the Pakerort and the upper part to the Varangu Stage. All the sections of the Türisalu Formation east of this area are of Varangu Age.

The Pakerort Stage is represented mainly by dark-brown horizontal laminated graptolite argillite. The lamination is caused by the different content of organic matter (intercalation of darker and lighter laminae) or by different grain-size (Хейнсалу 1990, Кивимяги и Лоог 1972). In some cases, wavy or cross-bedded structures or thin (a few cm) interbeds of light, often pyritized, quartzose siltstone occur in the lower part of the formation.

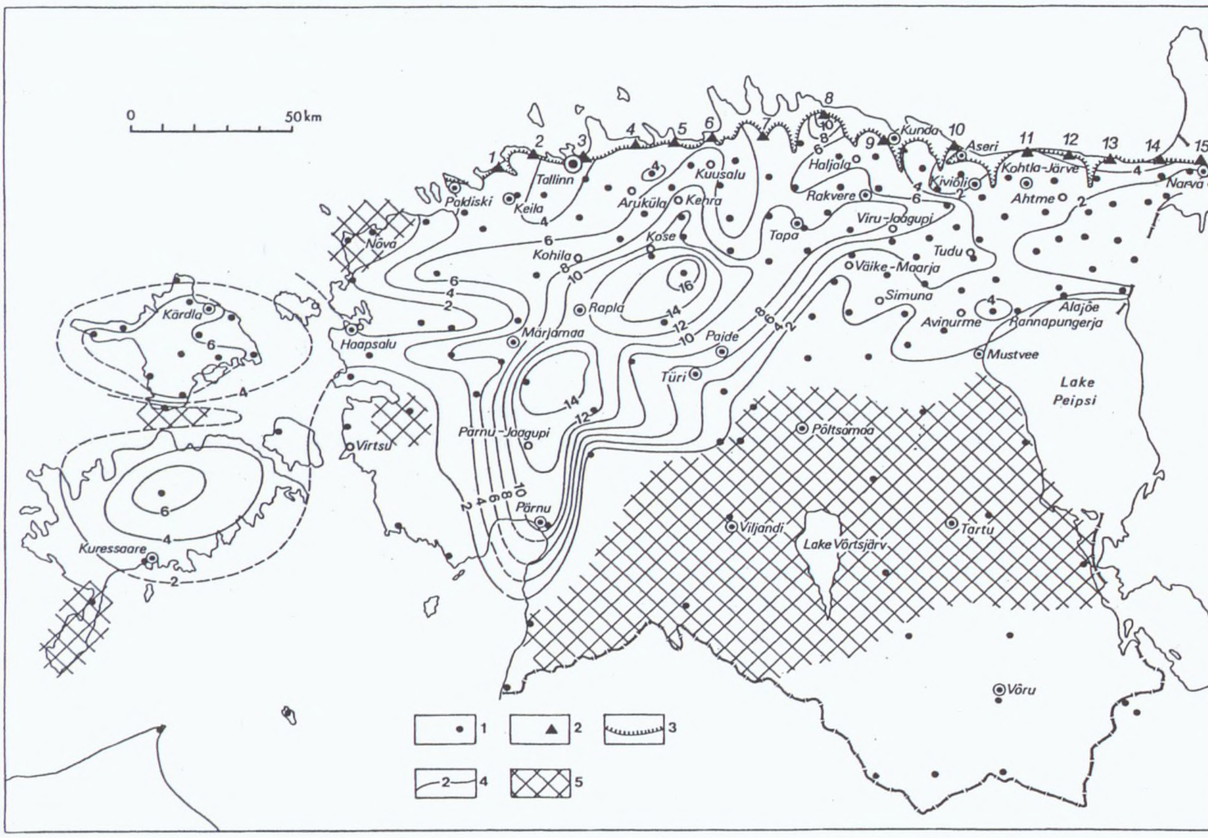


Fig. 27. Sketch-map showing the present distribution and the thickness of the Kallavere Formation: 1 - borehole; 2 - outcrop; 3 - North-Estonian Klint; 4 - isopachyte (m); 5 - area, where the Kallavere Formation is missing. Outcrops: 1- Türisalu; 2 - Rannamõisa; 3 - Hundikuristik (Suhkrumägi); 4 - Ülgase; 5 - Valkla; 6 - Turjekelder; 7 - Nõmmeveski; 8 - Vihula; 9 - Toolse; 10 - Aseri; 11 - Saka; 12 - Toila; 13 - Päite; 14 - Utria; 15 - Narva.

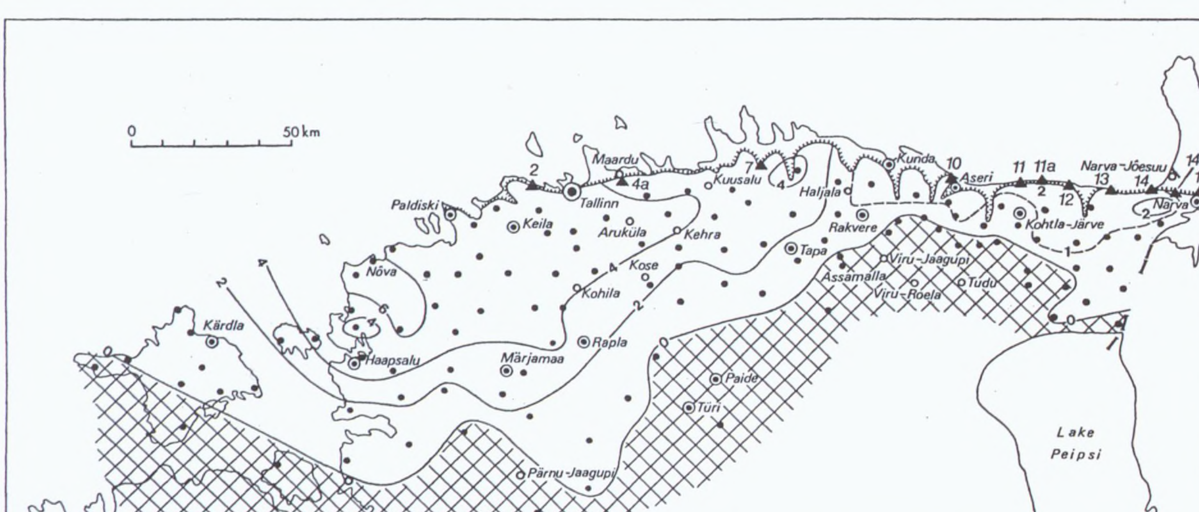


Fig. 28. Sketch-map showing the present distribution and the thickness of the Türisalu Formation. Outcrops: 2 - Rannamõisa; 4a - Maardu quarry; 7 - Nõmmeveski; 10 - Aseri; 11 - Saka; 11a - Valaste; 12 - Toila; 13 - Päite; 14 - Utria; 14a - Orasoja; 15 - Narva. For legend see Fig. 27.

There is no lithological markers for identification of the boundary between the Pakerort and Varangu stages in the limits of the Türisalu Formation and the maps of the distribution of Tremadoc rocks have been compiled by formations (Figs. 27, 28, 29). The thickness of the Türisalu Formation is up to 7 m (Fig. 28).

Varangu Stage

The later Tremadoc rocks, which belong to the Varangu Stage (Männil 1990, = Ceratopyge Stage, Мянниль 1966, Вийра и др. 1970) and have a thickness of 4-5 m extend, as a relatively narrow (20-50 km) belt in northern Estonia (Fig. 29). In the argillites, the lower boundary of the stage can be established by the appearance of graptolites of the *Kiaerograptus* Zone and conodonts of the *Paltodus deltifer pristinus* Subzone. The appearance of adelograptids marks the lower boundary of the Varangu Stage in the lithologically quite uniform Türisalu Formation. The upper part of the formation differs from the lower part, which belongs to the Pakerort Stage, by the occurrence of interbeds of very fine-grained quartzose sands from some mm up to 4-5 cm in thickness. Frequently, these interbeds abound in pyrite concretions.The Toolse area, where the Toolse Member was defined, has been studied in particular detail (Кивимяги и Лоог 1972, Хейнсалу 1980).

Varangu Formation

The Varangu Formation, the youngest part of the Tremadoc, is widely distributed in northwestern Estonia (Fig. 29). It is at its thickest (ca 3 m) between Haljala and Kunda in northeastern Estonia where the Varangu Formation can be subdivided into three lithologically different parts. The lower and upper parts are predominantly clayey, consisting mostly of compact claystone which comprises glauconite and pyrite, scattered or concentrated in small lenses. The middle part is rich in glauconite and very fine-grained quartz, sometimes prevailing over pelitic material. The sand is hardly pyritized. A similar three-part sequence of the Varangu Formation occurs also on the Pakri Cape in northwestern Estonia, but its thickness there is only 0.3-0.4 m.

In most of western Estonia, the Varangu Formation is characterized by the greenish-grey compact silty clay or sandy deposits with glauconite grains. In some sequences the clays of the Varangu Formation are dark in colour which makes them similar to the *Dictyonema* shale.

Hunneberg Stage

The Hunneberg Stage was introduced by Tjernvik (1956) as the Hunneberg Group in Sweden, based mainly on trilobite faunas. During several decades, the stage has been recorded as the lower substage of the Latorp Stage in Estonia (Мянниль 1966, Männil & Meidla 1994 *etc.*, after Jaanusson 1960a). Following Jaanusson (1982), Mägi (Мяги 1984) and Hints *et al.* (1994) considered this unit in the rank of stage.

During the last years, Sweden has served as the key area for biostratigraphical research of the Ontika Subseries comprising the stages from Hunneberg to Kunda. Detailed studies of the earliest post-Tremadoc sequences by Lindström (1955), Tjernvik (1956), Jaanusson (1963) and several other researchers have been supplemented by recent studies of sequences and distribution of graptolites (Lindholm 1991 a.o.) and conodonts (Löfgren 1993a, b; 1994, 1996). In the East Baltic region, the stratigraphy of the Ontika Subseries has been studied by Lamansky (Ламанский 1905), Öpik (1930b) and Männil (Мянниль 1963a, б, 1966). In Estonia modern biostratigraphy of this interval bases mainly on conodonts studied by Viira (Вийра 1966, Мяги и Вийра 1976, Mägi *et al.* 1989).

In northern Estonia, the Hunneberg Stage is represented by poorly lithified glauconitiferous terrigenous sediments: glauconitic siltstones of the Klooga Member (thickness up to 2.9 m) and glauconite silt and sand of the Joa Member (up to 1.2 m) which together form the lower, main part of the Leetse Formation (Figs. 30, 31). The content of glauconite is increasing upwards. The glauconitic siltstones of the Klooga Member are dominated by quartz with a supplement of glauconite (Мяги 1970), while the silt- and sandstones of the Joa Member consist mainly of glauconite (50-70%) and quartz (about 10-20%; Мяги 1970, 1984, Mägi 1990). The lower boundary of the Leetse Formation and the Hunneberg

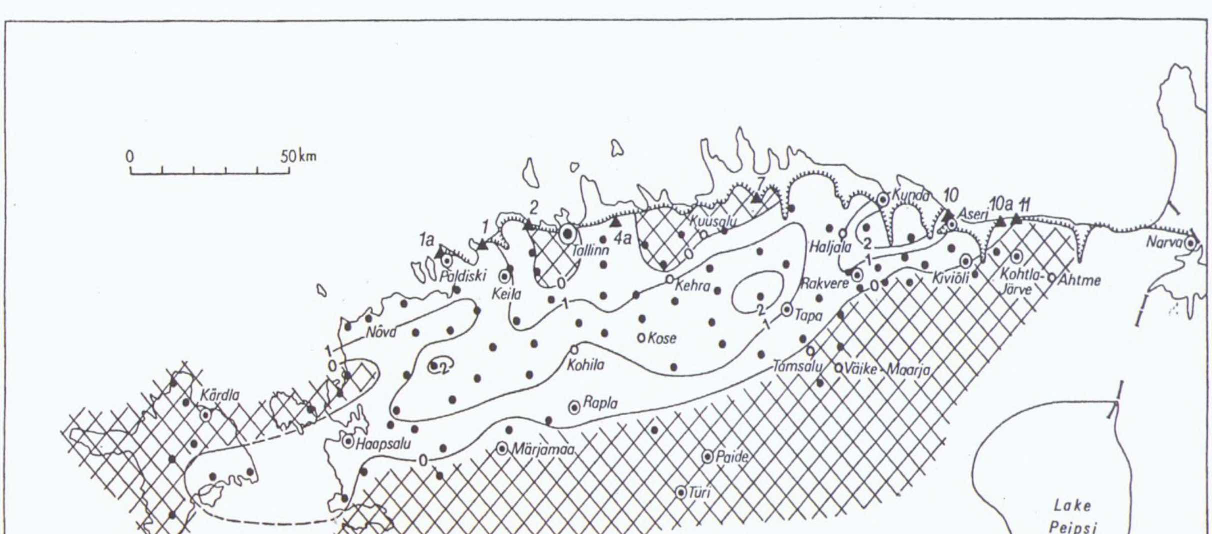


Fig. 29. Sketch-map showing the present distribution and the thickness of the Varangu Formation. Outcrops: 1 - Türisalu; 1a - Pakri Cape; 2 - Rannamõisa; 4a - Maardu quarry; 7 - Nõmmeveski; 10 - Aseri; 10a - Hiiemäe; 11 - Saka. For legend see Fig. 27.

Stage represents a well defined lithological marker level with the glauconitic sandstones and siltstones overlying conformably, sometimes with a discontinuity surface, the dark-brown argillites of the Türisalu Formation or light-grey clays of the Varangu Formation.

The maximum thickness of the stage reaches 4 m in north-western Estonia, but usually it is less than 2 m. As the Hunneberg and Billingen stages have not been differentiated in most sections, the thickness map (Fig. 30) shows only total thicknesses for both stages. In western Estonia, mainly on the islands of the West-Estonian Archipelago, the absence of the Hunneberg Stage has been documented from several sections.

In Sweden, the Hunneberg Stage corresponds to the *Megistaspis armata* and *M. planilimbata* trilobite zones. The base of the stage is close to that of the *Paroistodus proteus* conodont zone (Fig. 32, Löfgren 1993a). In the East Baltic, the lower *M. armata* Zone has been established only with confidence in Latvia (Мяэннилъ 1966, Ульст и др. 1982). In northern Estonia, the lowermost part of the Leetse Formation has generally been assigned to the *Paroistodus proteus* Zone (Männik & Viira 1990). In the Mäekalda section, the thin Klooga Member at the base of the Leetse Formation is referred to the *Paltodus deltifer* Zone (Fig.31, Мяги 1984).

Among macrofossils, a distinctive assemblage of lingulate brachiopods (Горянский 1969) has been recorded. It is characterized by *Thysanotos siluricus* (Eichwald) and *Leptembolon lingulaeformis* (Mickwitz) constituting the *Leptembolon-Thysanotos* assemblage, widely distributed in eastern and central Europe (Popov & Holmer 1994).

The fossil evidence from most of central and southern Estonia, is too fragmentary yet for the limitation of the Hunneberg and the overlying Billingen stages. The Hunneberg age of glauconitic sandstones (up to 0.5 m) has been established in the Karula core (Мяэннилъ 1966), but in most cases the detailed stratigraphy of the undivided Hunneberg - Billingen strata is unclear.

In Latvia, and in some sections in southern Estonia, close to the Estonian - Latvian border, the mudstones of the Zebre Formation, reaching a thickness of 46 m in Latvia, have been considered as equivalents of the Varangu, Hunneberg and Billingen stages (Ульст и др. 1982, see also Fig. 30). The middle part of this formation (Zirni Member) is of Hunneberg age, as it yields a zonal trilobite *Megistaspis planilimbata* (Augelin) (Fig. 32) and several graptolites, including *Tetragraptus phyllograptoides* Strandmark, together with a zonal conodont *Paroistodus proteus* (Lindström) (*ibid.*). The occurrence of the peripheral parts of the Zebre Formation can be assumed in southern Estonia.

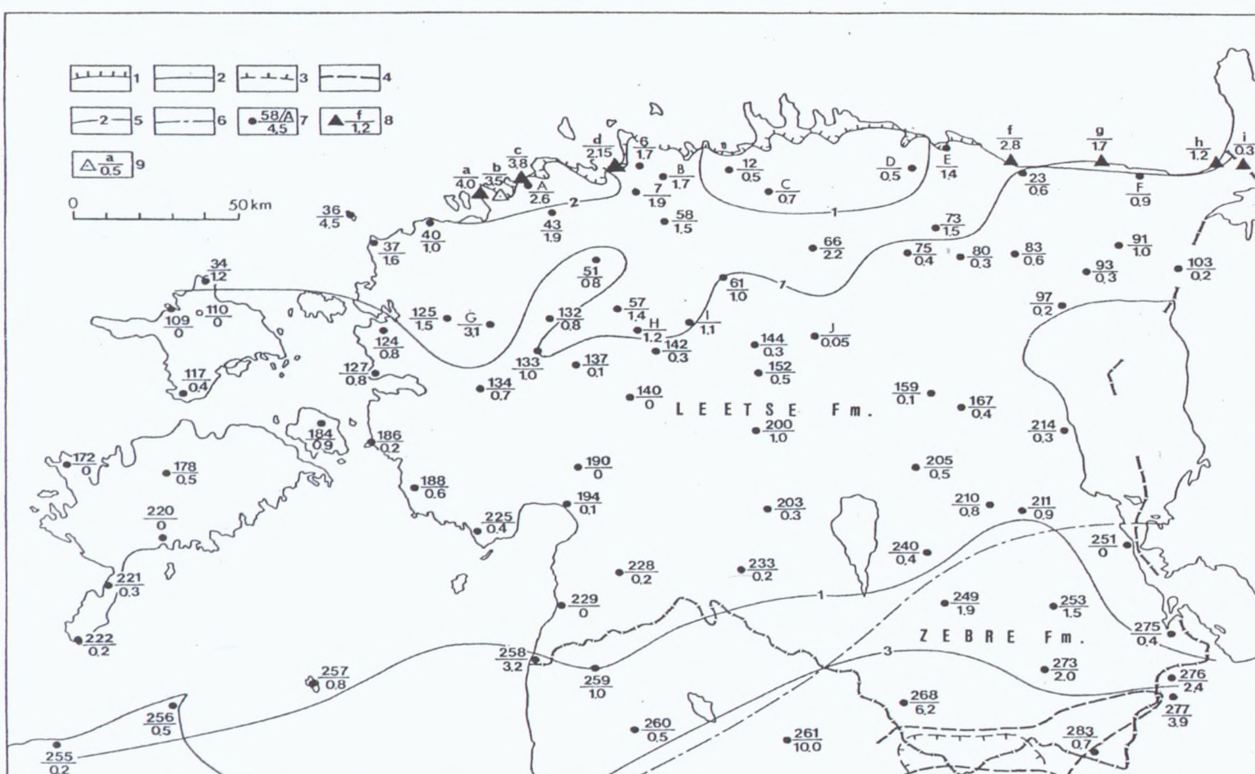


Fig. 30. Sketch-map showing the present distribution and the sum thickness of the Hunneberg and Billingen stages. The thicknesses here and on Figs. 34 and 36 are based on the published data (Рыымусокс 1960, Орвику1960a, Мяэннилъ 1966, Rõõmusoks 1983, a.o.) and unpublished data by L. Põlma. 1 and 2 - outer and inner border of the outcrop area; 3 and 4 - outer and inner border of the area where the unit has incomplete thickness on the Lokno Elevation; 5 - isopachyte (m); 6 - approximate boundary between the distribution areas of the formations; 7 - core section, the number (see Fig. 3) or index (A .., a..) in the numerator and thickness of the stage (or stages) in metres in the denominator; 8 - outcrop, the index in the numerator and thickness of the unit in the denominator; 9 - the same of the type section. Outcrops: a -Väike- Pakri Island, type section of the Pakri Formation; b - Leetse, type section of the Leetse Formation; c - Keila-Joa; d - Mäekalda; f - Aseri; g - Ontika; h - Tõrvajõe; i - Narva. Core sections: A - Keila-Joa; B - Saha-Loo; C - Pillapalu; D - Essu; E - Kunda; F - Voka; G - Maidla; H - Keava; I - Vahastu; J - Kahala II.

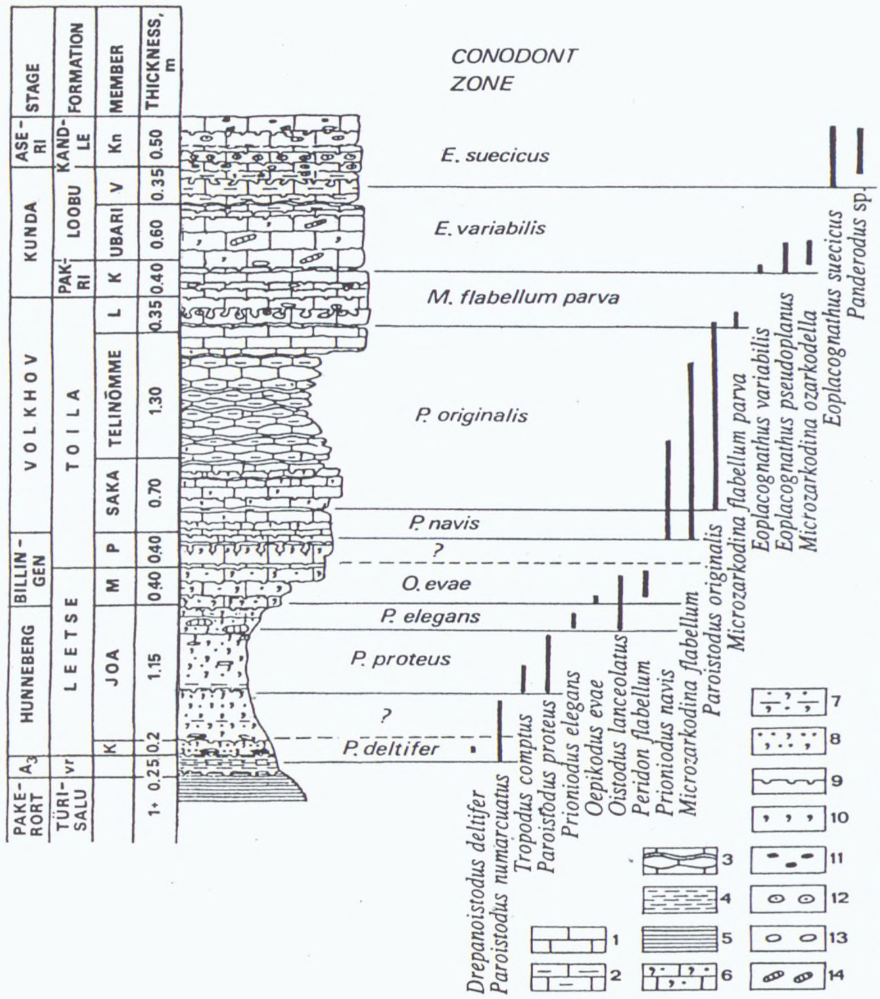


Fig. 31. The lower part of the Mäekalda section (Tallinn) with the conodont zonation and range of selected species (Einasto *et al.* 1966). Indices of stratigraphic units: A_3 - Varangu Stage, vr - Varangu Formation; members: K - Klooga, M - Mäeküla, P - Päite, L - Lahepera, K - Kallaste, V - Valgejõe, Kn - Künnapõhja.
Legend: 1 - limestone; 2 - argillaceous limestone; 3 - nodular limestone with clayey interbeds; 4 - clay; 5 - argillite; 6 - sandy glauconitic limestone; 7 - clayey glauconitic sand; 8 - glauconitic siltstone; 9 - discontinuity surface; 10 - glauconite; 11 - goethitic ooids; 12 - phosphatic ooids; 13 - pebbles; 14 - cephalopods.

Billingen Stage

The Billingen Stage (Tjernvik 1956, Jaanusson 1982), understood here in the sense of the upper Billingen Substage of the Latorp Stage (Jaanusson 1960a, Мянниль 1966, Männil 1990, Männil & Meidla 1994), consists of two distinctive parts in northern Estonia. The lower one is represented by the glauconitic calcareous sandstones and limestones of the upper part of the Leetse Formation (Mäeküla Member, equal to $B_I\beta$ by Lamansky (Ламанский 1905), whereas the glauconitic limestones of the lowermost part of the Toila Formation (Päite Member, equal to $B_I\gamma$ by Jaanusson (1951), form the upper half of the Billingen Stage (Table 7, Fig. 31). In some publications, including the detailed lithological study of the Volkhov and Kunda stages by Orviku (Орвику 1960a), the Päite Member is interpreted as the lowermost unit of the Volkhov Stage. This interpretation is also in use in the Leningrad Region of Russia.

The original concept of the Billingen Stage (Tjernvik 1956) was based on the evidence from the trilobite faunas, but its lower boundary can best be recognized by the distribution of conodonts. In Sweden, it is situated fairly close to the boundary of the *Prioniodus elegans* Zone and this level is recognizable also in several sections of northern Estonia where it nearly coincides with the lower boundary of the Mäeküla Member in the upper part of the Leetse Formation (Fig. 31).

The Mäeküla Member consists of glauconitic sandstones which are replaced upwards by calcareous sandstones and glauconitic limestones. The lower boundary of this member is lithologically fairly distinct in the klint area, and is marked by the change from poorly cemented silts and sands to well cemented sandstones. For the purposes of correlation, the occurrence of the conodonts *Prioniodus elegans* Pander and *Oepikodus evae* (Lindström) is most important (Figs. 31, 32). The thickness of the Mäeküla Member varies from 0 to 0.5 m. Sandy material with the grain-size over 0.1 mm forms up to 80% of the rock, whereas glauconite grains make up some 80% of this fraction. The quartz content varies from 10 to 40% (Мяги 1984, Mägi 1990). The highest content of glauconite has been recorded in central northern Estonia. The macrofauna of the Mäeküla Member has not been described monographically but, according to the available evidence, its main, upper part is comparable with the *Megistaspides dalecarlicus* Zone of Sweden (Fig. 32).

The Päite Member is represented by limestones or dolomites which dominate in the easternmost sequences, with a low content of mainly fine-grained glauconite. The greatest thickness of the member is 1.13 m, and it decreases in the northwest direction. In the Leningrad Region, the presence of several distinctive lithological marker horizons within the equivalents of the Päite Member (roughly equal to the informal Red Dikari Member) has been demonstrated by Dronov *et al.* 1996, part of those can be supposedly distinguished in northeastern Estonia. On the islands of Väike-Pakri (Photo 18) and Osmussaar, the member is sandy and may contain a layer of calcareous, glauconitic sandstone at

British Series	Baltoscandian Series	Baltoscandian Subseries	Stage	Substage	Baltoscandian trilobite zones	North Atlantic conodont zones
Llanvirn	Oeland	Ontika	KUNDA	Aluoja	*Megistaspis gigas*	*Eoplacognathus suecicus*
					Megistaspis obtusicauda	*Eoplacognathus variabilis*
				Valaste	*Asaphus "raniceps"*	
				Hunderum	*Asaphus expansus*	
Arenig			VOLKHOV	Langevoja	*Megistaspis limbata*	*Microzarkodina flabellum parva*
				Vääna	*Megistaspis simon*	*Paroistodus originalis*
				Saka	*Megistaspis polyphemus*	*Baltoniodus navis* / *Baltoniodus triangularis*
			BILLINGEN		*Megistaspis estonica*	*Oepikodus evae*
					Megistaspis dalecarlicus	*Prioniodus elegans*
			HUNNEBERG		*Megistaspis planilimbata*	*Paroistodus proteus*
					Megistaspis armata	
						Paltodus deltifer

Fig. 32. The Ontika trilobite zonation and its correlation with the conodont zones (compiled after Tjernvik 1956, Jaanusson 1982, Männik & Viira 1990, Löfgren, 1996, Nielsen 1995).

the base (Орвику 1960a). The Päite Member is roughly equivalent to the *Megistaspis (Paramegistaspis) estonica* Zone of Sweden (Fig. 32).

In central Estonia, the presence of the Billingen Stage needs further approval. Glauconitic sandstones occuring locally in a restricted thickness in this area, like in the Äiamaa and Võhma cores (Рыымусокс 1960, Rõõmusoks 1983), may belong to this stage but their precise age is not yet clear. Further to the south (at Tartu), the reddish-brown, occasionally glauconitic dolomites may be tentatively assigned to the Billingen Stage. The distribution of the Zebre Formation in Latvia (Ульст и др. 1982) suggests that it extends as far as the southernmost Estonia. In Latvia, the red or mottled clays of the topmost Zebre Formation (Zante Member) contain *Megistaspis (Paramegistaspis) estonica* (Tjernvik) and a zonal conodont *Oepikodus evae* (Lindström) (Гайлите и Ульст 1975, Ульст и др. 1982), characteristic of the Billingen Stage in several sections of northern Estonia (Mägi 1990, Einasto *et al.* 1996, see Fig. 31).

Fossil evidence from southern Estonia is too fragmentary to enable the limitation of the Hunneberg and Billingen stages over the study area. Figure 30 shows only their sum thickness. The absence of the Billingen Stage in several sections of the West-Estonian Archipelago (Saaremaa, Hiiumaa) should be mentioned.

The Mäeküla Member contains the oldest Ordovician representatives of articulate brachiopods of the genera *Plectella, Panderina, Prantlina* and *Angusticardinia* (Рубель 1961) bryozoans, ostracodes (H. Aru, pers. comm) and trilobites. Frequent occurrence of trilobites ("*Megistaspis*") has been recorded in some levels (Орвику 1960a). In the Leningrad Region, the Mäeküla Member yielded the material for original definition of Conodonta by Pander (1830). The yet poorly studied fossil record of the Päite Member contains conodonts, brachiopods, trilobites and ostracodes.

Volkhov Stage

The Volkhov Stage, corresponding roughly to the "Glauconit kalk" by Schmidt (1881, Шмидт 1879), forms a lithologically distinctive unit in the sections of the North-Estonian Klint and nearby river valleys (Photo 19). The term "Volkhov" was introduced by Raymond (1916) as the "Walchow Formation" in a broader meaning (Table 1) cor-

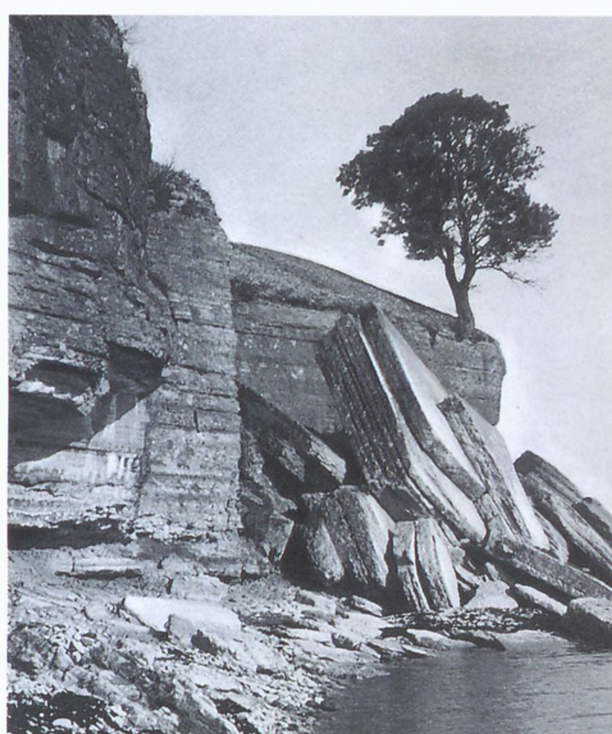


Foto 18. Klint on the island of Väike Pakri. A section from the sandy Billingen Stage (below) to the calcareous Uhaku Stage (top). *Photo by Karl Orviku*

responding to the lower part of the Ontika glauconitic limestones in northern Estonia. Lamansky (Ламанский 1905) was the first to introduce three substages in the Leningrad Region. In Estonia these are in ascending order the Saka, Vääna and Langevoja substages (Männil & Meidla 1994), conceptually largely based on trilobite zonation. The two first units are not accepted in Sweden.

In northern Estonia, the Volkhov Stage is represented by the main part of the Toila Formation, locally succeeded by the lower part of the Sillaoru Formation (Pada Member, Fig. 33) which has tentatively been assigned to the Volkhov Stage (Орвику 1960a, a.o.). In central Estonia, the Toila Formation is assumed to compose the whole Volkhov sequence. Southward the formation grades into the Kriukai Formation (Table 7, Fig. 34).

The Toila Formation is a complicated stratum made up of various, partly dolomitized glauconitic limestones resting on sandstones of the Leetse Formation. The lower boundary of the Volkhov Stage and the Saka Member is marked by a smooth discontinuity surface with "amphore-like borings" (Орвику 1960б) in the uppermost bed of the Päite Member (Орвику 1960a), known as Püstakkiht (Orviku 1961). In northern Estonia, the main part of the Toila Formation, corresponding to the Volkhov Stage, is subdivided into five members (Орвику 1960a) which are partly lateral equivalents. Only the lower, Saka Member consisting of dolomitized glauconitic limestone (up to 1.2 m) forms the base of the Volkhov Stage all over northern Estonia (Fig. 33). It is overlain by two laterally equivalent units: the Telinõmme Member in the west (interbedded greenish-grey limestones and marls, up to 2 m) and the Künnapõhja Member in the east (mottled dolomitic limestone, up to 1 m) (Fig. 33). East of Tallinn, the upper part of the formation is represented by the Kalvi Member (grey argillaceous glauconitic limestones) with a thickness of up to 1.7 m. West of Tallinn, it is composed of the Lahepera Member (glauconitic limestones, partly sandy or conglomeratic, up to 0.5 m) which is assumed to represent the youngest part of the formation, as it locally overlies the Kalvi Member.

According to the trilobite evidence, the lower, Saka Member comprising trilobites *Megistaspis "elongata"* (Schmidt) and *Megistaspis "lata"* (Törnquist), and a zonal conodont *Baltoniodus navis* (Lindström) (as referred to by Мянниль 1966 and Männil & Meidla 1994), represents the lower, Saka Substage in Estonian succession. The Telinõmme and Künnapõhja members comprising *Paroistodus originalis* (Sergeeva) may correspond to the middle, Vääna Substage, while the Kalvi and Lahepera members probably correspond to the upper, Langevoja Substage (Fig. 32). Ostracodes (Fig. 35) and brachiopodes (Рубель 1961) are common in the Toila Formation. Gastropods, cephalopods and cystoids have also been recorded.

The up-to-3.5-m-thick Toila Formation is poorly developed in northwestern Estonia (Fig. 34). In northeastern Estonia where the formation is at its thickest, the rocks have undergone extensive dolomitization.

The Sillaoru Formation, the "untere Linsenschich" by Schmidt (1897), consists of two distinct members of oolitic limestones with noticeably different ages. The lower, Pada Member (up to 0.5 m of limestone with small ferriferous ooids and occasional glauconite grains) comprises *Metaptychopyge truncata* (Nieszkowski) (Решения...

Photo 19. The rocks of Billingen, Volkhov and Kunda stages are exposed in the scarp of the Jägala Waterfall and in the walls of the canyon. *Photo by A. Miidel*

1978), *Ptychopyge angustifrons* (Angelin) (Mägi 1990) and is apparently of Late Volkhov (Early Kunda?) Age. Its age relationship with the Lahepera Member of the Toila Formation remains still open due to the different distribution areas of these members (Fig. 31).

In central Estonia, the sequence of the Toila Formation comprises the same members with similar thicknesses as in northeastern Estonia (Saka, Künnapõhja and Kalvi). In south-

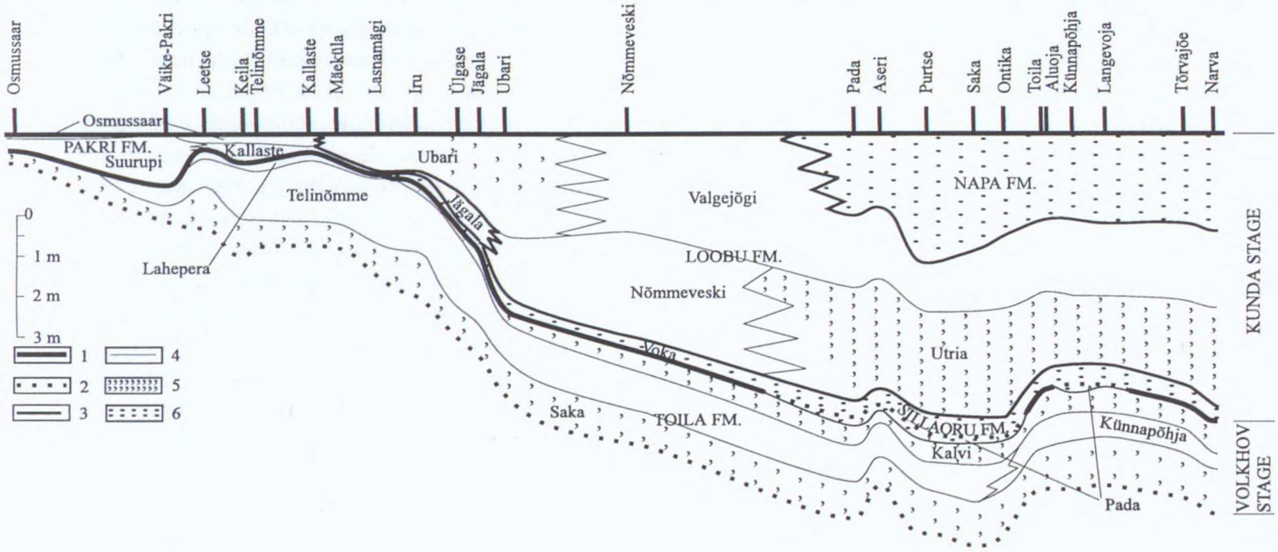


Fig. 33. Stratigraphical profile of the Volkhov and Kunda stages in northern Estonia, from west (Osmussaar, "a" on Fig. 34) to east (Narva, "w"), showing the distribution of formations and members, by Orviku (Орвику 1960a) and Rõõmusoks (1983), modified. Legend: 1 - intraformational stage boundary; 2 - interformational stage boundary; 3 - formation boundary; 4 - member boundary; 5 - glauconite; 6 - goethitic ooids; formation - block letters, member - small letters.

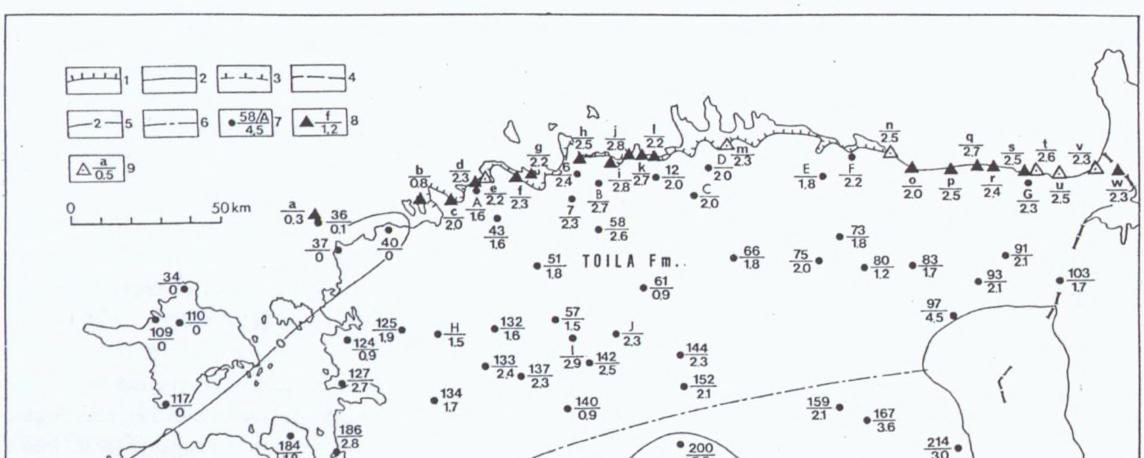
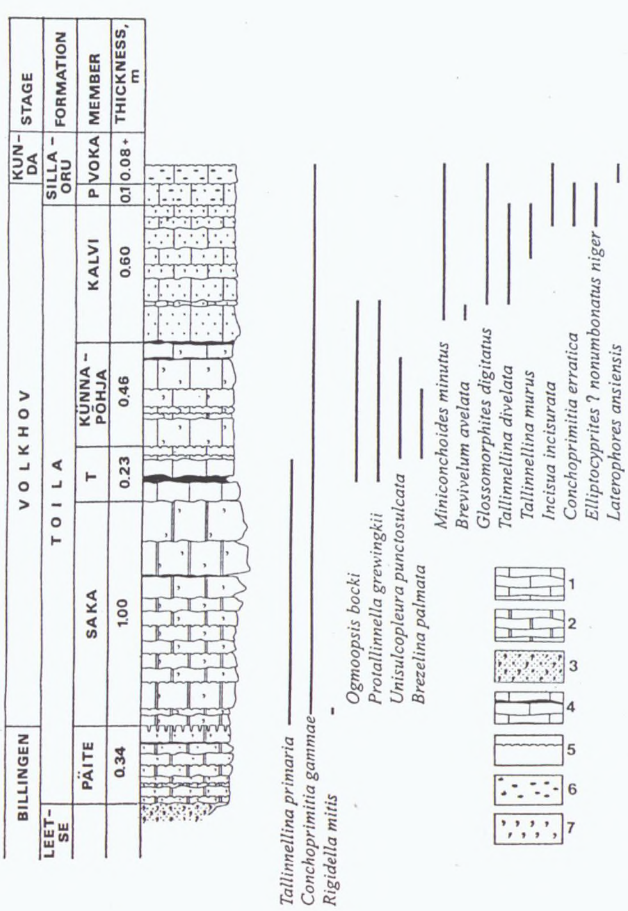


Fig. 34. Sketch-map showing the present distribution and the thickness of the Volkhov Stage. Legend 1- 9 see Fig. 30. Outcrops: a - Osmussaar, b - Väike-Pakri, c - Leetse, d - Keila-Joa, e - Telinõmme, type section of the Telinõmme Member, f - Kallaste, g - Mäeküla, h - Lasnamägi, i - Iru, j - Ülgase, k - Jägala, l - Ubari, m - Nõmmeveski, type section of the Nõmmeveski Member, n - Pada, type section of the Pada Member, o - Aseri, p - Purtse, q - Saka, type section of the Saka Member, r - Ontika, s - Toila and Aluoja, t - Künnapõhja, type section of the Künnapõhja Member, u - Langevoja, type section of the Langevoja Substage, v - Tõrvajõe, w - Narva; core sections: A - Keila-Joa, B - Saha-Loo, C - Pillapalu, D - Kemba, E - Essu, G - Voka, H - Maidla, I - Keava, J - Vahastu.

ern Estonia, the formation grades into the Kriukai Formation (Table 7, Fig. 34) which consists mainly of reddish-brown marls, with limestone and mudstone intercalations. The thickness of these rocks does not exceed 20 m in Estonia, but in Latvia it is much greater, reaching 32.5 m. The age of the formation in southern Estonia has not been established biostratigraphically. However, in Latvia it displays a rich fauna of trilobites, ostracodes, conodonts, more rarely articulate brachiopods and lingulates (*Megistaspis "limbata"* (Boeck), *Ptychopyge angustifrons* Angelin, *Tallinnellina primaria* (Öpik), *Rigidella mitis* (Öpik) (Гайлите и Ульст 1975, Ульст и др. 1982, Männil & Meidla 1994), suggesting the Volkhov Age (Сарв 1959, Meidla & Sarv 1990, Männil & Meidla 1994).

Kunda Stage

The Kunda Stage (Kunda Formation by Raymond 1916) is represented by oolitic, glauconitic (Ламанский 1905) and sandy limestones corresponding to the emended Vaginatum Limestone by Schmidt (1897). A three-part subdivision of the strata, based on the trilobite zonation, was introduced already by Lamansky (Ламанский 1905) in the Leningrad Region. He also assumed the absence of the lower unit - the *Asaphus expansus* Zone in northern Estonia which was afterwards confirmed by several authors (Raymond 1916, Орвику 1960а, Мянниль 1966 a.o.). Orviku (Орвику 1958б) proposed to name the Lamansky's subdivisions in ascending order the Hunderum, Valaste and Aluoja substages, and presented a detailed lithostratigraphical description of the corresponding interval in northern Estonia (Орвику 1958б, 1960а,б; Fig. 33).



In northern Estonia, the Kunda Stage comprises the Valaste (corresponding to the *Asaphus "raniceps"* Zone) and the Aluoja (zones of *Megistaspis obtusicauda* and *Megistaspis gigas*) substages (Fig. 32, Table 7). In most of northern Estonia, the lower boundary of the Valaste Substage is drawn at the base of the oolitic limestone of the Sillaoru Formation, or locally within the unit, being marked by a discontinuity surface on the boundary between the Pada and Voka members, and the disappearance of the glauconite grains, characteristic of the underlying strata (Fig. 33).

In northwestern Estonia, the Kunda Stage is represented by the Pakri Formation, eastwards replaced by the upper part of the Sillaoru Formation and the Loobu Formation (Fig. 33). In northeastern Estonia, the Napa Formation forms the topmost part of the Kunda Stage and grades into the Rokiškis Formation in central Estonia. In southern Estonia, the entire Kundan sequence, including the Hunderum Substage, is represented by the Šakyna and Baldone formations (Fig. 36).

The Pakri Formation (Öpik 1927), up to 4.5 m of yellowish-grey sandy limestones and calcareous sandstones, sometimes with conglomerate beds, occurs in northwestern Estonia in the area west of Tallinn. In the westernmost part of its distribution area, the main, lower part of the Pakri Formation consists of up-to-4m-thick nodular kerogenous calcareous sandstone of the Suurupi Member, overlain by the thin (0.5 m) sandy limestone of the Osmussaar Member. In the surroundings of Tallinn, the formation is represented by limestones with quartz and glauconite grains, locally with a basal conglomerate (Kallaste and Jägala members, up to 0.8 m). On the Island of Osmussaar and, to a lesser extent, in the neighbouring mainland areas, a system of sedimentary dikes cuts through the Pakri Formation and the underlying Volkhov-Billingen strata (Пуура и Туулинг 1988). The time of the formation of the dikes is dated as middle-late Kunda.

In most of northern Estonia (except the distribution area of the Pakri Formation) and in central Estonia, the basal part of the Kunda Stage consists of the oolitic limestone of the Sillaoru Formation (Решения... 1978, Мянниль и Рыымусокс 1984). The main, Valaste time part of the formation (Voka Member, up to 0.6 m) consists of clayey limestones with abundant ferriferous ooids, developed around skeletal particles or glauconite grains (Мяги 1984). Among the skeletal particles, fragments of trilobites and ostracodes dominate (50-70%, Мяги 1984). The Voka Member generally serves as a good marker level in the North-Estonian sequence, although in restricted areas of northern and northeastern Estonia it overlies the thin oolitic Pada Member which differs from the main part of the formation by the presence of glauconite grains and has been included in the Volkhov Stage by Orviku (Орвику 1960а) and subsequent authors. In northern and northeastern Estonia, the Loobu Formation constitutes the main part of the Kunda Stage. Detailed study of the formation (Орвику 1958б, 1960а) has

Fig. 35. The ranges of ostracodes in the Saka section. Legend: 1 - limestone; 2 - dolomite; 3 - glauconite sandstone; 4 - clayey interbeds in limestone; 5 - discontinuity surface; 6 - goethitic ooids; 7 - glauconite; T - Telinõmme Member, P - Pada Member.

revealed its two-part subdivision; both the lower and the upper parts consist of two laterally equivalent units. In central northern Estonia, east of Tallinn, the formation is represented by clayey limestone of the Nõmmeveski Member (up to 2 m) and glauconitic limestone of the Ubari Member (up to 2 m, Fig. 33). In northeastern Estonia, the lower part of the formation consists of glauconitic limestone of the Utria Member (up to 3 m), overlain by clayey limestones of the Valgejõgi Member (up to 4.7 m, Мянниль 1987). In the outcrop area, large nautiloids *Cyclendoceras vaginatum* (Schlotheim), *Estonioceras ariense* (Schmidt), *Paracyclendoceras cancellatum* Eichwald *etc.*, (Рыымусокс 1960) are characteristic of most of the Loobu Formation. In northeastern Estonia, the rocks have undergone extensive dolomitization resulting in a mottled red colour and cavernous structure. The Loobu Formation reaches its maximum thickness (7 m) in the central part of northern Estonia (Fig. 36), in central Estonia it is less than 3 m thick (Решения... 1987). In that area the formation consists of grey, partly clayey glauconitic limestones, overlying the oolitic limestones and marls of the Sillaoru Formation (0.5 m).

The Napa Formation, an oolitic marl and limestone body (up to 4 m), is supposed to replace the upper part of the Loobu Formation in northeastern and central Estonia (Fig. 33).

The relation of the formations and members forming the Kunda Stage in northern Estonia is well demonstrated by Orviku (Орвику 1960a). The correlation is largely based on the trilobite evidence. *Asaphus "raniceps"* Dalman has been identified in the Suurupi Member of the Pakri Formation and in the lower part of the Loobu Formation (Nõmmeveski Member). The Osmussaar Member comprises *Pseudoasaphus globifrons* (Eichwald), which is known from the upper part of the Loobu Formation (Ubari and Valgejõgi members). The Napa Formation is characterized by *Megistaspis gigas* Angelin (Решения... 1978, Mägi 1990). In terms of conodont zonation, the Valaste and Aluoja substages roughly correspond to the *Eoplacognathus variabilis* Zone (Fig. 32). The shelly fauna is represented by brachiopods, ostracodes, gastropods and cephalopods (Öpik 1927, Сарв 1959, Рубель 1961, Mägi 1990).

In central Estonia, the Napa Formation grades into the Rokiškis Formation (Fig. 36), which is represented by red mottled oolithic limestone (up to 15 m). The fauna of this unit is poorly known in Estonia. Based on *Panderodus cf. sulcatus* (Fåraeus) and *Pinnatulites procera* (Kummerow) recorded by Männil (Решения... 1987), the Kunda-Aseri age has been suggested. In southern Estonia, the sequence of the Loobu and Rokiškis formations grades into the Šakyna and Baldone formations, represented by grey glauconitic limestone and clayey red limestone, respectively. Palaeontologically, these units are poorly characterized in Estonia and the age relationship to the northern and central Estonian sequences is obscure. The fauna of the Šakyna Formation in Latvia contains trilobites, more rarely brachiopods and graptolites, the Baldone Formation is more fossiliferous (Гайлите и Ульст 1975, Ульст и др. 1982). In this area the succession of the above-named formations comprises the entire Kunda Stage and stratigraphically the section of southern Estonia is the completest in this interval.

The thickness of the Kunda Stage demonstrates an obvious decreasing trend towards northwestern Estonia. In most of northern and central Estonia, it does not exceed 10 m, but in southeastern Estonia may locally reach 20 m (Fig. 36).

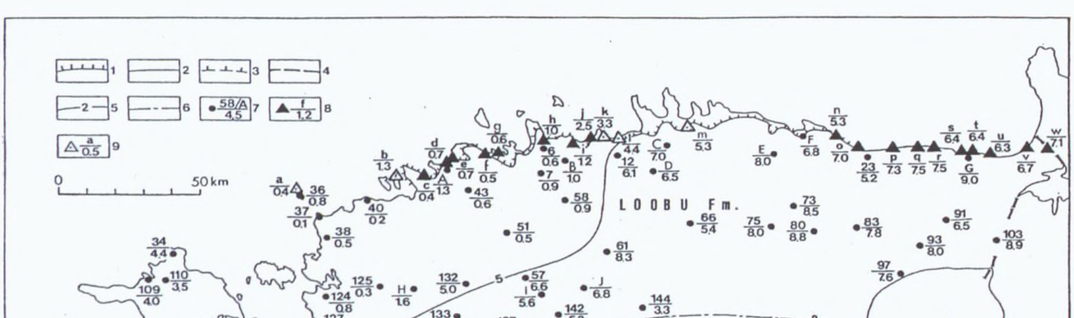


Fig. 36. Sketch-map showing the present distribution and the thickness of the Kunda Stage. Legend 1 - 9 see Fig. 30. Outcrops (a - w) see Fig. 33; among them a - Osmussaar, type section of the Osmussaar Member, b - Väike -Pakri, type section of the Pakri Member, k - Jägala, type section of the Jägala Member, l - Ubari, type section of the Ubari Member, m - Nõmmeveski, type section of the Nõmmeveski Member; core sections: A - Keila-Joa, B - Saha-Loo, C - Kemba, D - Pillapalu, E - Essu, F - Kunda, G - Voka, H - Maidla, I - Keava, J -Vahastu. The signs "L. p." and "U.p" show whether the unit forms the lower or upper part of the stage.

VIRU SERIES

Aseri Stage

The term Aseri Stage was used first by Bekker (1922, 1923) for the Schmidt´s (1897) Upper Oolitic Limestone (Obere Linsenschicht). In nowadays understanding the Aseri Stage bases in a great deal on the studies carried out by Orviku (Jaansoon-Orviku 1927, Orviku 1929, 1930a, 1940), Rõõmusoks (Рыымусокс 1960, 1970) and Männil (Мянниль 1966).

In northern and central Estonia, the Aseri Stage is 0.1 - 5 m thick (Fig. 37) and consists of bioclastic limestones with unevenly distributed ooids, predominantly brown ferriferous (goethitic) ooids (Orviku 1940, Орвику 1960б). In places, the ooids are frequent in the lower and upper parts of the stage, but in the dolomitic limestones of northeastern Estonia they occur only in the upper part (Fig. 38). White phosphatic ooids are distributed mainly in the westernmost sequences. These, early Middle Ordovician oolitic limestones have been treated as the Kandle Formation (*sensu stricto*; Мянниль и Рыымусокс 1984). Afterwards, Männil (1990, Männil & Meidla 1994) proposed the name Aseri Formation for the oolitic limestones of Aseri Age. Here preference is given to the term Kandle Formation, because in many cases the upper boundary of the Aseri Stage is difficult to determine; it may coincide with the upper boundary of the oolitic limestones or fall into the upper part of it.

The Kandle Formation is subdivided into the Malla (Мянниль и Рыымусокс 1984) and Ojaküla (Orviku in Аалоэ и др. 1958) members. The lower, Malla Member (*Asaphus* and *Echinosphaerites* limestones by Jaansoon-Orviku 1927) is lithologically the most variable part of the Kandle Formation and differs from the predominantly oolitic limestones of the Ojaküla Member (*Cephalopod* Limestone) in the occurrence of glauconite, *e.g.* in the surroundings of Jägala, or in the absence of ooids in some places or parts of the sequence. The thickness of the Malla Member decreases from 2.5 m in the eastern to 0.30 m in the central part of the klint area. West of Jägala, the Malla Member is missing and the Aseri Stage is represented by the 10–20-cm-thick sandy oolitic (mainly with phosphatic ooids) limestones of the upper, Ojaküla Member.

The Kandle Formation extends to central Estonia (Fig. 38) with the dominantly grey-coloured limestones of northern sections turning southwards brownish-grey or yellowish-grey. In southern Estonia, the stage is represented by the up-to-9m-thick reddish-brown limestones of the Segerstad Formation (Мянниль 1966, Männil & Meidla 1994). In the transitional area between the Kandle and Segerstad formations, reddish-brown and mottled limestones with occasional goethitic ooids (Männil 1990, Männil & Meidla 1994) have been distinguished. They belong presumably to the upper part of the Rokiškis Formation (Лашков и др. 1984). In practice, identification of the latter unit in sections seems in a great deal subjective, and its distribution area is difficult to determine.

The lower boundary of the Aseri Stage in recent use was defined by Orviku (Jaansoon-Orviku 1927, Orviku 1929). In contrast to Bekker (1922), he excluded from that stage the lowermost part of the oolitic limestones which comprises several early Ordovician (Oelandian) taxa (*Ahtiella baltica* Öpik, *Antigonambonites* sp., *Megistaspis* sp.,

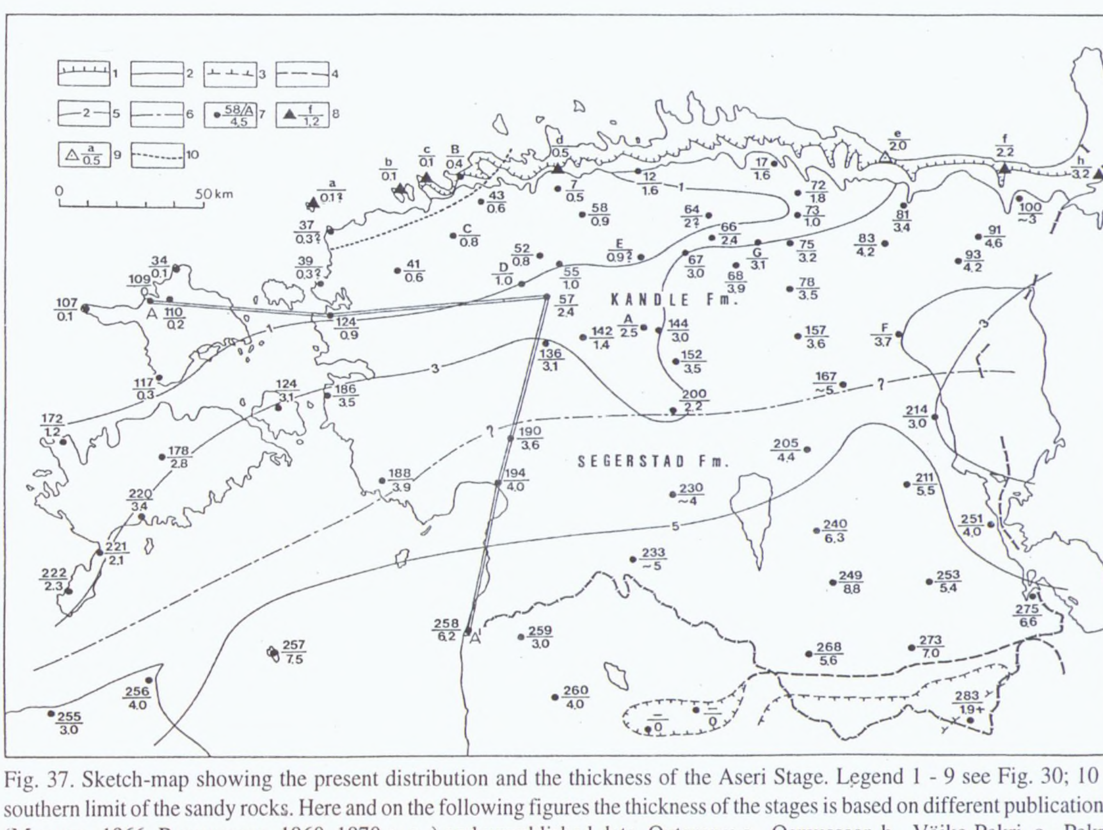


Fig. 37. Sketch-map showing the present distribution and the thickness of the Aseri Stage. Legend 1 - 9 see Fig. 30; 10 - southern limit of the sandy rocks. Here and on the following figures the thickness of the stages is based on different publications (Мянниль 1966, Рыымусокс 1960, 1970 a. o.) and unpublished data. Outcrops: a - Osmussaar, b - Väike-Pakri, c - Pakri Cape, d - Lasnamägi. e - Aseri, type section of the Aseri Stage, f - Voka, h - Narva; core sections: A - Väätsa, B - Keila Joa, C - Munalaskme, D - Oela, E - Atla, F - Mustvee, G - Tamsalu. Profile line A - A‘, see Fig. 40.

Рыымусокс 1970 p. 30). The boundary is marked by essential changes in the faunal composition, especially in trilobites and cephalopods (Рыымусокс 1970, table 3, see also Jaanusson 1960a, 1963). Notable is the disappearance of the trilobite genus *Megistaspis* and appearance of *Asaphus (Neoasaphus)*, represented at least by six species in northern Estonia (Рыымусокс 1970; table 3). *Asaphus platyurus*, a characteristic species in the Segerstad Limestone in Sweden and Latvia (Jaanusson 1960a, Мянниль 1963б, 1966) occurs also in southern Estonia (Karula core, Мянниль 1966, fig. 12). Of new faunal elements, *Echinosphaerites* as a quite easily notable fossil is also worth of mentioning (Jaansoon-Orviku 1927, p. 15, 16; Orviku 1929, p. 9-11).

The data published on the distribution of ostracodes in the Aseri Stage in Estonia is scanty (Сарв 1959, Мянниль 1966). The ostracode *Pinnatulites procera* Zone of the Kunda Stage is replaced by the *Piretella tridactyla* Zone in the Aseri Stage (Meidla & Sarv 1990). In several core sections (Мянниль 1966, figs. 12-14), *Euprimites effusus* Jaanusson appears close to the lower boundary of the Aseri Stage.

The chitinozoans are known only in the grey-coloured rocks of the Kandle Formation which comprises the *Cyatochitina regnelly* and *C. striata* zones (Table 7). The boundary between these zones coincides with the boundary between the Malla and Ojaküla members.

The Aseri Stage corresponds to the lower part of the *Didymograptus murchisoni* graptolite zone and roughly to the *Eoplacognathus suecicus* conodont zone (Männil 1990, Männik & Viira 1990, Einasto *et al.* 1996).

Lasnamägi Stage

In northern Estonia, the fairly uniform Early Viru sequence of comparatively thick-bedded, hard bioclastic limestones abounding in discontinuity surfaces (Saadre 1992, 1993), was first distinguished as a separate unit - the Building Limestone (Baukalkstein), by Orviku (Jaansoon-Orviku 1927). Subsequently, this unit, determined mainly by the lithological criteria, was termed (Orviku 1940) the Lasnamägi Stage after the sections in the Lasnamägi quarry in the northeastern part of Tallinn. The Lasnamägi Stage is well-exposed also in some other sections, including Suhkrumägi (Photo 20) and Mäekalda (see Einasto *et al.* 1996, fig. A16) in the vicinity of the type section. In general lines, Orviku's interpretation of the Lasnamägi Stage kept valid until the 1970s (Jaanusson 1945, Рыымусокс 1960, 1970, Мянниль 1963a). In 1966, Männil (Мянниль 1966) stated that the Building Limestone comprises two distinct successive faunal associations of which the upper one with several characteristic trilobites, such as *Xenasaphus devexus devexus* (Eichwald), *Asaphus (Neoasaphus) lepidus* Törnquist, and graptolites including *Gymnograptus linnarssoni* (Moberg), is closely related to the fauna of the overlying argillaceous limestones of the Uhaku Stage. The beds containing the "upper" fauna were included (Мянниль 1966, 1976, Решения... 1978) to the Uhaku Stage, while the term Lasnamägi Stage was restricted to the lower half of the Väo Formation (Photo 21) in recent use (Мянниль и Рыымусокс 1984), corresponding roughly to the Kallaste Substage by Rõõmusoks (Рыымусокс 1970), and to the lower part of the former Building Limestone.

The 4–10-m-thick Väo Formation (Fig. 39) is subdivided

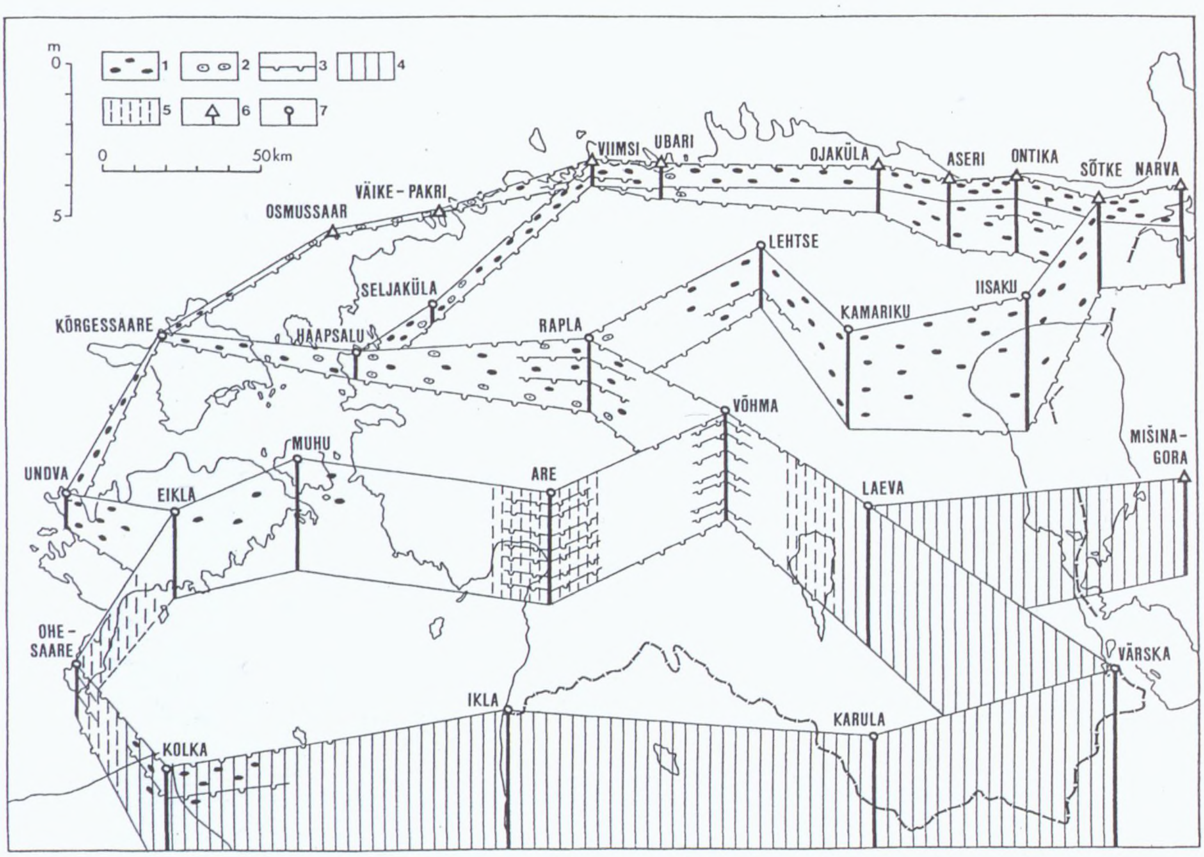


Fig. 38. Sketch-map showing the sections of the Aseri Stage: 1 - goethitic ooids; 2 - phosphatic ooids; 3 - discontinuity surface; 4 - red-coloured and 5 - mottled limestones; 6 - outcrop; 7 - core section.

into three units; in ascending order these are the Rebala (the relatively argillaceous part, thickness up to 3 m), Pae (dolomites, up to 1.5 m) and Kostivere (hard limestones, up to 6 m) members. Besides, in the stratotype area where the formation has a detailed bed-by-bed stratification, each layer has a name of its own given by quarry-workers (Mägi 1990, Einasto *et al.* 1996). Männil (Мяннилъ 1976) included to the Lasnamägi Stage the Rebala and Pae members and the lower part of the Kostivere Member, up to the discontinuity surface, above which *Gymnograptus linnarssoni* appears.

In northern Estonia, the lower, Lasnamägi part of the Väo Formation is up to 4.5 m and in the stratotype area at Lasnamägi up to 4 m thick (Мяннилъ 1976, Мяннилъ и Саадре 1987, Mägi 1990). Still in many sections the exact level of the upper boundary of the Lasnamägi Stage is not established and in Fig. 39 the total thickness of the Väo Formation is given. In southern Estonia, the Lasnamägi Stage is represented by red (lower part) to grey (upper part) bedded, mostly micritic limestones of the Stirna Formation (Ульст и Гайлите 1976), equivalent to the Seby and Folkeslunda limestones of Öland Island and mainland Sweden (Männil & Meidla 1994). The Stirna Formation is up to 15 m thick (Fig. 39) which corresponds to the maximum thickness of the formation in Estonia and northwestern Latvia (Ульст и др. 1982, fig. 45). In the transitional area between the Väo and Stirnas formations in central Estonia, the oolitic lithofacies is developed (Пылма 1982, fig. 7).

In northern Estonia, the lower boundary of the Lasnamägi Stage falls into the upper part of the oolitic limestones, predominantly with goethitic ooids, of the Kandle Formation. It is marked by the discontinuity surface above which there appear brachiopods (*Equirostra, Noetlingia*), trilobites (*Illaenus schroeteri* (Schlotheim), *Illaenus schmidti* Nieszkowski, cephalopods (*Lituites* sp.) and others (Jaanusson 1945, Рыымусокс 1970). The level of the appearance of phosphatic ooids in the top of oolitic limestones is used as the lower boundary of the Lasnamägi Stage, if the boundary discontinuity surface is absent or the palaeontological data are insufficient.

The lists of fossils published earlier for the Lasnamägi Stage (Рыымусокс 1970, table 4) can be used with consideration that only the data from the Kallaste Substage by Rõõmusoks (Рыымусокс 1970) characterize the Lasnamägi Stage in recent meaning. In northern Estonia, the macrofauna is quantitatively dominated by sedentary forms, particularly articulate brachiopods and bryozoans (Jaanusson 1984). Cephalopods occur mostly in lower quantities, except northeastern Estonia where the lowermost beds of the stage abound in orthocones and where lituitids are also fairly common. The same groups of fossils seem to be relatively abundant in core sections as well (Хинтс и Пылма 1981).

Important information on the range of North Atlantic conodont zones (see Bergström, 1971for the reference), graptolites and chitinozoans of the Lasnamägi Stage in northern Estonia is provided by Männil (Мяннилъ 1976, fig. 2; 1986, fig. 2.1.1). According to him, the stage is comparable to the *Eoplacognathus foliaceus* Subzone and the main lower part of *E. reclinatus* Subzone. Although the chitinozoans are not very dignostic for the distinction of the Lasnamägi Stage, the *Cyathochitina sebyensis* Zone is a good marker for the Aseri-Lasnamägi boundary beds (Table 7).

Uhaku Stage

The Uhaku Stage comprises, in the revised and amended form (Jaanusson 1960a, Мяннилъ 1966, 1976, Männil 1990), the *Caryocystites* Zone (Jaansoon-Orviku 1927, = Uhaku Stage by Orviku 1940) and the upper part of the Building Limestone (Väo Formation). These two parts of the Uhaku Stage are considered also as substages (Мяннилъ 1976, Männil 1990).

The thickness of the Uhaku Stage varies from 5-10 m in western to about 20-25 m in eastern Estonia (Fig. 39). In northern Estonia, the hard bioclastic limestones of the Väo

Photo 20. The Suhkrumägi outcrop in Tallinn displays carbonate rocks from the Volkhov to Lasnamägi stages which have been described by many scholars. *Photo by Ago Aaloe.*

Photo 21. The steps of the Treppoja Waterfall west of Tallinn have been eroded into the resistant rocks of the Lasnamägi Stage. *Photo by A. Miidel.*

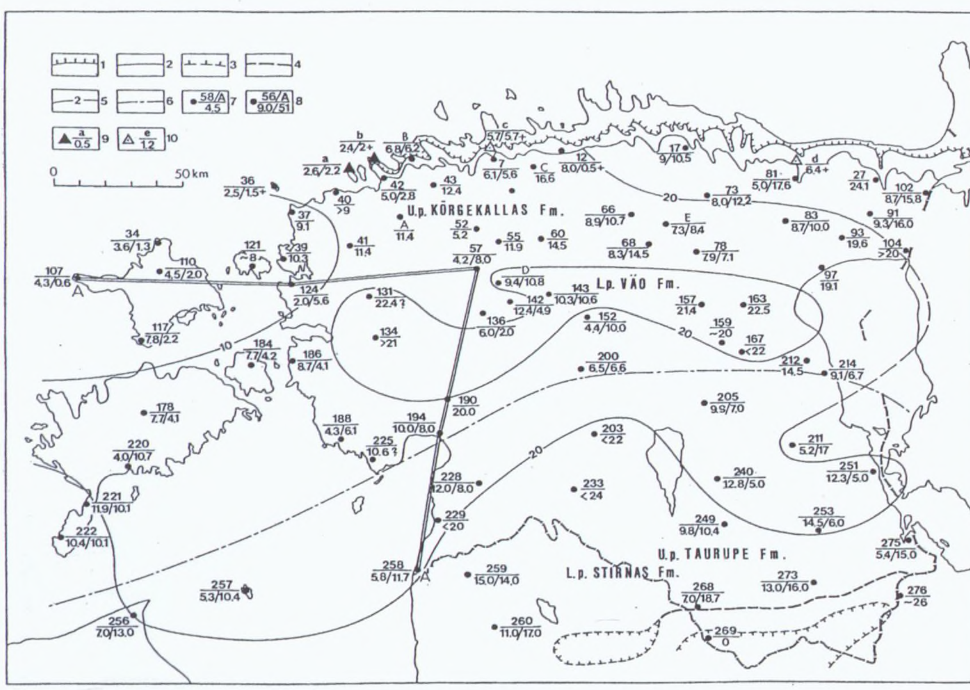


Fig. 39. Sketch-map showing the present distribution and the sum thickness of the Lasnamägi and Uhaku stages. Legend 1 - 6, 9 and 10 see Fig. 30; 7 - borehole, the number (see Fig.) or index (A...F) in the numerator and the sum thickness of the Lasnamägi and Uhaku stages in the denominator, 8 - the same, but the denominator shows the thickness of the Väo (first number) and Kõrgekallas formations (second number). Outcrops: a - Väike-Pakri, b - Pakri Cape, c - Lasnamägi, type section of the Lasnamägi Stage, d - Uhaku, type section of the Uhaku Stage; core sections: A - Munalaskme, B - Keila-Joa, C - Raasiku, D - Keava, E - Tamsalu. Sections on the profile line A - A' see Fig. 40. The signs "L. p." and "U. p" show whether the unit forms the lower or upper part of the described interval.

Formation, forming the lower part of the Uhaku Stage, are of a rather stable thickness (4-5 m). The upper part of the Uhaku Stage, made up of relatively thin-bedded argillaceous limestones of the Kõrgekallas Formation, is subdivided (Table 7) into the Koljala, Pärtlioru and Erra members (Мяннилъ и Рыымусокс 1984). The thickness of the formation decreases from about 18 m in northeastern to 1-2 m in northwestern Estonia (Figs. 39, 40). The lower boundary of the Kõrgekallas Formation and the Koljala Member, formed of argillaceous limestones with marly intercalations, supposedly coincides with the lower boundary of the *Conochitina tuberculata* Zone (Мянниль и Бауерт 1986, p. 17; Table 7). In the Pärtlioru and Erra members, the argillaceous intercalations in the bioclastic limestones are partly kerogeneous. In the Oil Shale Basin in northeastern Estonia (Пуура 1986), thin kukersite beds (up to 2 cm) occur, or they form together with limestones and marls distinct intervals (up to 1.6 m) between relatively pure limestones (Мянниль и Бауерт 1986). These are the oldest kukersite beds in the Middle Ordovician sequence in northern Estonia. In central Estonia, the kukersite beds appear in the Kukruse Stage (*cf.* Мянниль 1966, 1986).

In southern Estonia and also in Latvia, the Uhaku Stage is represented mainly by micritic limestones with intercalation of bioclastic limestones and marls of the Taurupe Formation (= Furudal Formation in Мянниль 1966) with a thickness of 6 - 19 m. Only in a few sequences in western Latvia, the Taurupe Formation is over 20 m thick (Ульст и др. 1982, fig. 46). The limestones of the transitional belt between northern and southern Estonia are characterised by an interfingering pattern which resembles that of northern Öland in Sweden (Мянниль 1966). Both goethitic and phosphatic ooids occur in some places in the basal part of the Uhaku Stage testifying to continuous shift of the oolitic lithofacies in time (Пылма 1982, fig. 7, Pärnu and Ikla cores, Fig. 40).

In northern Estonia, the lower boundary of the Uhaku Stage falls into the lithologically rather uniform Väo Formation. In practice, a prominent discontinuity surface is used as a boundary marker above which several new taxa appear, some of them widespread and frequent throughout the Baltic Basin. Mass occurrence of the trilobite *Xenasaphus d. devexus* (Eichwald) is recorded from the Island of Osmussaar in the west as far as Ingria (L. Popov and R. Einasto, pers. comm., Алихова 1960, 1969) in the east. Of the graptolites of the *Hustedograptus teretiusculus* Zone, *Gymnograptus linnarssoni* (Moberg) is identified from the Oslo Region up to the Moscow Syneclise (Мянниль 1976). According to Männil (Мянниль 1986), the lower part of the Uhaku Stage corresponds to the *Eoplacognathus robustus* and *E. lindstroemi* subzones of the *Pygodus serra* Zone (Table 7). The upper part of the stage corresponds to the *Pygodus anserinus* Zone. The latter zonal species appears close to the base of the Kõrgekallas Formation. Nevertheless, on the basis of the distribution of conodonts, the boundary between the Lasnamägi and Uhaku stages is unclear, at least on the subzones level (Table 7). The chitinozoan *Conochitina clavaherculi* Subzone comprises the most part of the Väo Formation, including the strata with the first finds of *G. linnarssoni* (Мянниль 1986, fig. 2.1.1).

The Uhaku Stage comprises a varied sedentary benthic fauna, particularly articulate brachiopods, bryozoans and cystoids. Since there is no generally acknowledged interpretation of the Uhaku Stage, the lists of fossils presented by different researchers comprise taxa from different stratigraphical intervals (Рыымусокс1960, 1970; Мянниль 1963a, 1966).

Macrofossils are poorly known in the subsurface area of the Uhaku Stage (Fig. 40). The Taurupe Formation, which is distributed in southern Estonia, includes many elements, such as *Nileus* and *Upplandiops* (=*Estoniops* sp. n. in Мянниль 1966, fig. 12) among trilobites, and both *Alwynella*? and *Christiania* among articulated brachiopods, which are widely distributed in the Furudal limestones in Sweden (Jaanusson 1960a, 1963, Jaanusson & Ramsköld 1993).

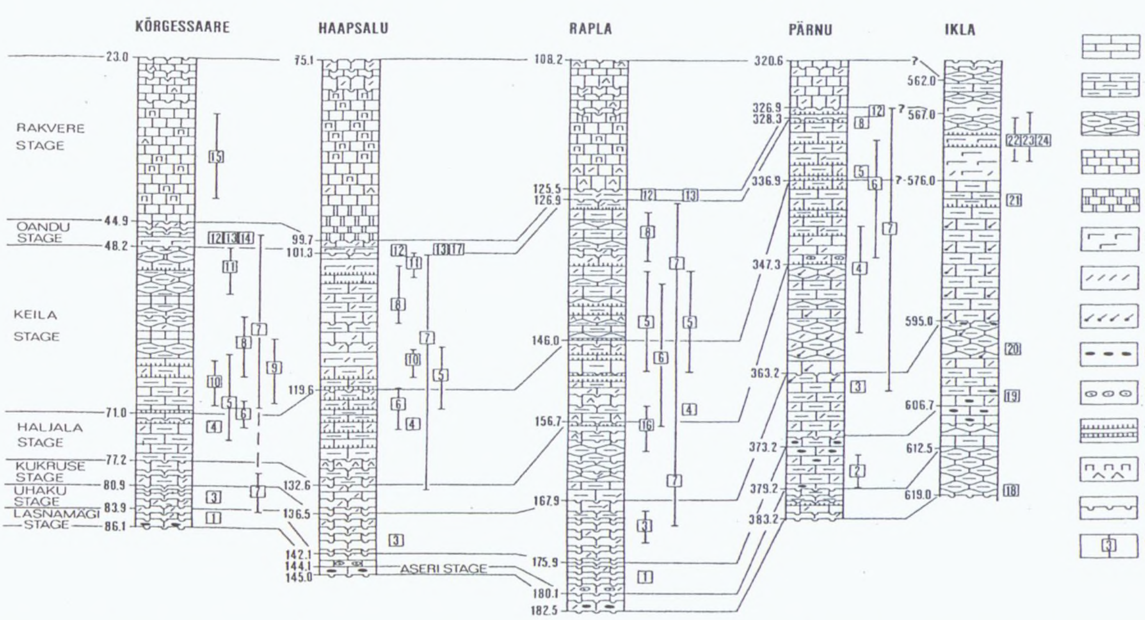


Fig. 40. Correlation of the Middle Ordovician sequences on the line A-A‘ (see Fig. 39). Legend: 1 - limestone; 2 - argillaceous limestone; 3 - nodular argillaceous limestone; 4 - micritic (aphanitic) limestone; 5 - dolomite; 6 - calcareous marl; 7 - skeletal sand; 8 - pyritized skeletal sand; 9 - goethitic ooids; 10 - phosphatic ooids; 11 - K -bentonite; 12 - pyritized spots (upper sign), kukersite kerogen (lower sign); 13 - discontinuity surface; 14 - number and stratigraphical interval of selected fossils: *1 - Christiania oblonga, 2 - Lituites lituus, 3 - Clitambonites squamatus, 4 - Bilobia?* sp., *5 - Mastopora concava, 6 - Pyritonema subulare, 7 - Sowerbyella* sp. sp., *8 - Keilamena*? sp. 9 - *Saukrodictya* sp., *10 - Baltocrinus hrevicaensis, 11 - Horderleyella? kegelensis, 12 - Howellites* sp., *13 - Vellamo oanduensis, 14 - Kjaerina orvikui, 15 - Holtedahlina* sp., *16 - Bekkerina*? sp., *17 - Ilmarinia* sp., *18 - Asaphus (Neoasaphus) platyurus, 19 - Illaenus chiron, 20 - Christiania* sp., *21 - Bimuria* aff. *peregirina*, 22 - *Platystrophia rava, 23 - Skenidioides* sp., *24 - Haplosphaeronis* sp.

Kukruse Stage

The Kukruse Stage (Kuckerssche Schicht by Шмидт 1879, Schmidt 1881) as a stratigraphical unit comprises the commercially exploited oil shale (kukersite) seams (Chapter X) and the richest and most diverse faunal assemblage in the Ordovician of Estonia represented by more than 330 species and subspecies (Рыымусокс 1970, table 10).

The stratigraphy of the Kukruse Stage has been dealt with in several papers (Рыымусокс 1957, Мяннилъ 1984, Мяннилъ и Бауерт 1984, 1986; Bauert 1993, Saadre & Suuroja 1993b). The bed-by-bed stratification of the kukersite complex with special sets of indices for the individual kukersite seams form the base for the correlation of the sequences within the kukersite basin (Мяннилъ 1984, fig. 2; Bauert & Puura 1990).

The thickness of the Kukruse Stage (Fig. 41) ranges from about 3 m in western to more than 20 m in eastern Estonia (Saadre & Suuroja 1993a). The stage consists of three lithologically distinct formations. The argillaceous bioclastic limestones with intercalations of kukersite (oil shale) and marls of the Viivikonna Formation (Мяннилъ и Рыымусокс 1984) are distributed northeast of the line Osmussaar Island (southwestern Estonia) - Mehikoorma (south coast of Lake Peipsi) (Fig. 41). Based on the frequency of kukersite seams or the content of the kerogenous component, the Viivikonna Formation is subdivided into the Kiviõli, Peetri and Maidla members (Fig. 42). The Kiviõli (lower) and Peetri (upper) members differ from the Maidla (middle) Member by the occurrence of 10—14-cm-thick kukersite seams, while the middle part of the formation consists of kerogenous and variously argillaceous limestones (Мяннилъ и др. 1986, Bauert 1993, figs 3, 4). Due to the facies shift of the kukersite beds (Мяннилъ и др. 1986), the boundaries of the Viivikonna Formation are diachronous. As a result, the upper part of the Viivikonna Formation (Peetri Member) is missing in northeastern Estonia, but it is exposed in the vicinity of Tallinn (Fig. 42, Nõlvak & Hints 1996) and is well-known by core sections south of the outcrop area (Мяннилъ 1984, Мяннилъ и Бауерт 1984, Мяннилъ и Саадре 1987).

Westwards, the Viivikonna Formation grades into the bioclastic limestones of the Pihla Formation with a thickness of about 3 - 6 m (Saadre & Suuroja 1993b) and southwards into the limestones with dark pyritized skeletal detritus and nodular intercalations of argillaceous marls of the Dreimani Formation (Fig. 41, Спрингис 1974). The thickness of the latter varies from 7 to 14 m and only in southeastern Estonia it is about 20 m, which is nearly the same as in eastern Latvia (Ульст и др. 1982, fig. 47).

For the lower boundary of the Kukruse Stage, Bekker (1923, 1924b) proposed the base of the lowermost commercially important kukersite seam "A" at the base of the Viivikonna Formation. The renovation of faunal association begins with the appearance of new bryozoans in seam "A". Somewhat higher, in seam "C" several new species, includ-

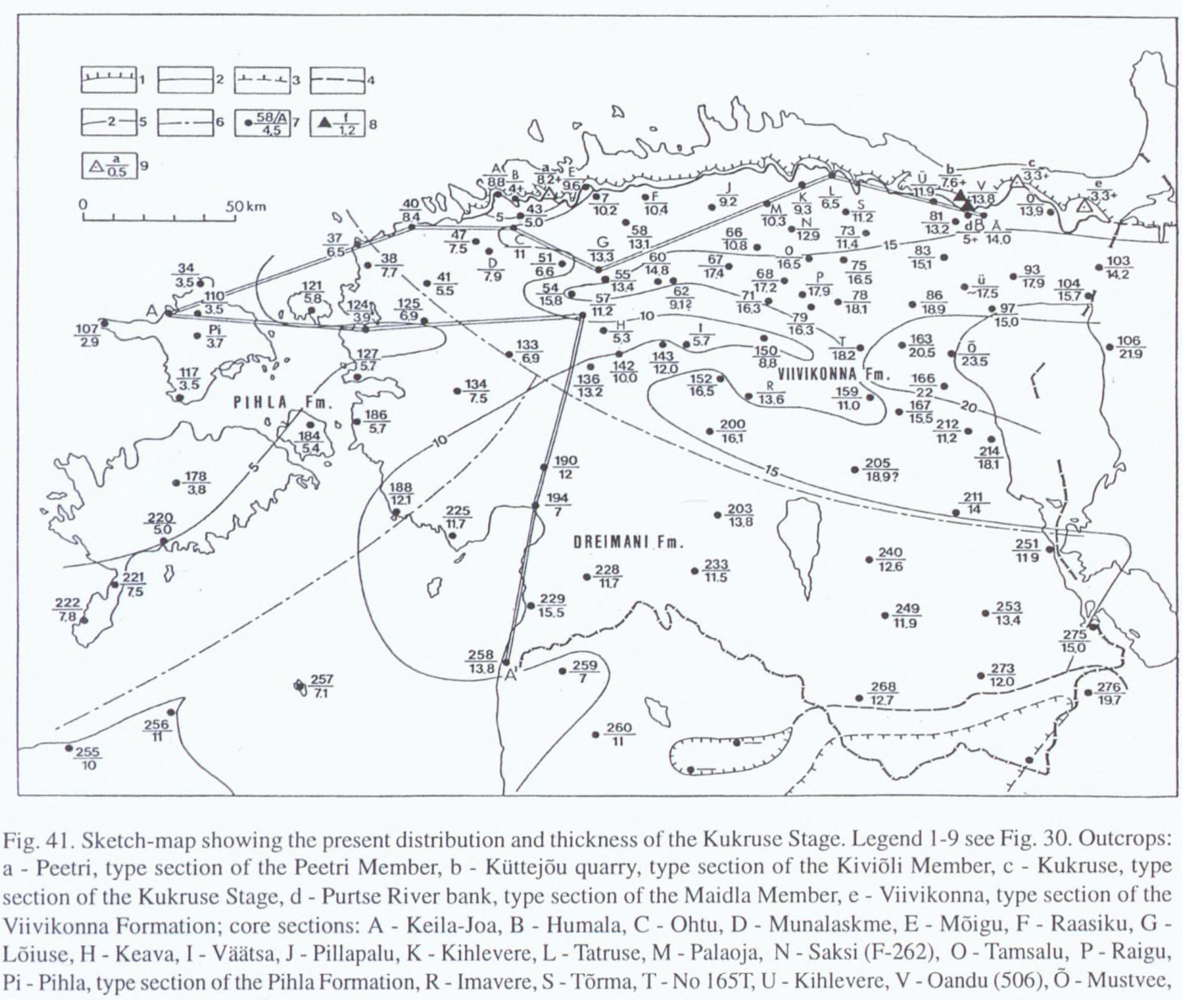


Fig. 41. Sketch-map showing the present distribution and thickness of the Kukruse Stage. Legend 1-9 see Fig. 30. Outcrops: a - Peetri, type section of the Peetri Member, b - Küttejõu quarry, type section of the Kiviõli Member, c - Kukruse, type section of the Kukruse Stage, d - Purtse River bank, type section of the Maidla Member, e - Viivikonna, type section of the Viivikonna Formation; core sections: A - Keila-Joa, B - Humala, C - Ohtu, D - Munalaskme, E - Mõigu, F - Raasiku, G - Lõiuse, H - Keava, I - Väätsa, J - Pillapalu, K - Kihlevere, L - Tatruse, M - Palaoja, N - Saksi (F-262), O - Tamsalu, P - Raigu, Pi - Pihla, type section of the Pihla Formation, R - Imavere, S - Tõrma, T - No 165T, U - Kihlevere, V - Oandu (506), Õ - Mustvee, Ä - Kohtla, Ö - Ruskavere, Ü - Tagajõe. Sections on the profile line A-A'see Fig. 40; on the line A - B – Fig. 42.

ing the brachiopods *Bilobia musca* (Öpik), *Sowerbyella (S.) liliifera* (Öpik), *Estonomena estonensis* (Bekker), and the trilobites *Asaphus (Neoasaphus) nieszkowskii* Schmidt, *Estoniops exilis* (Eichwald), *Paraceraurus aculeatus* (Eichwald) appear (Рыымусокс 1970, table 9). In western and southern Estonia, the base of the Pihla or Dreimani Formation is used as the lower boundary of the Kukruse Stage. This level is marked by the appearance of indicator ostracodes *Baltonotella kuckersiana* (Bonnema), *Conchoprimitia leperditioides* Thorslund, *Euprimites locknensis* Thorslund and others, several of which are common with the lower part of the Dalby Limestone in Sweden (Мянниль 1966, Jaanusson 1976). At the same time, several early Viru taxa, such as *Chasmops odini odini* Eichwald, *Sowerbyella (Viruella) uhakuana* (Rõõmusoks), *Platystrophia biforata* (Schlotheim), *Dianulites fastigiatus* (Eichwald) and others, disappear close to the lower boundary of the Kukruse Stage (Рыымусокс 1970, p.156, 157). The graptolite *Orthograptus uplandicus* whose range zone corresponds to the Kukruse Stage (Мянниль 1984) and the chitinozoa *Cyathochitina savalaensis* appear roughly on the lower boundary of the Kukruse Stage. In all likelihood, also the boundary between the North Atlantic conodont *anserinus* and *tvaerensis* zones (Мянниль и Бауерт 1986) falls into the lower part of the Kukruse Stage.

The diverse assemblage of Kukruse macrofossils is represented first of all by bryozoans (more than 60 species), brachiopods (about 90 species) and trilobites (about 50 species: Рыымусокс 1970, table 10) which form about two thirds of the species identified. The most abundant and diverse association occurs in the Kiviõli Member in the lower part of the stage. Still some species, such as *Hesperorthis inostrantzefi inostrantzefi* (Wysogorski), *Echinosphaerites aurantium suprum* Hecker, are notable due to their mass occurrence in the upper part of the stage (Рыымусокс 1970, p. 169). The character of the distribution of some brachiopods and trilobites, such as *Estlandia marginata magna* Öpik, *Otarion planifrons* (Eichwald), *Pharostoma nieszkowskii* (Schmidt) and others, shows a facies shift from the lower part of the Kukruse Stage (Kiviõli Member) in northeastern to the upper part (Peetri Member) in northwestern Estonia.

In the core sections, macrofossils are of secondary importance due to their scarcity, especially in western Estonia (Fig. 40). Still, the occurrence of some species should be noticed. In some northernmost core sections, the brachiopod *Kullervo panderi* (Öpik) marks the lowermost part of the Kukruse Stage (Рыымусокс 1970). In the outcrops, this species appears presumably in the kukersite seam "G", which lies 1-4 m above the lower boundary of the stage. In the southern periphery of the Viivikonna Formation and in the Dreimani Formation, *Asaphus (Neoasaphus) ludibundus* Törnquist and *Bilobia musca* (Öpik) appear in the Kukruse Stage and in some areas *Echinosphaerites* becomes frequent.

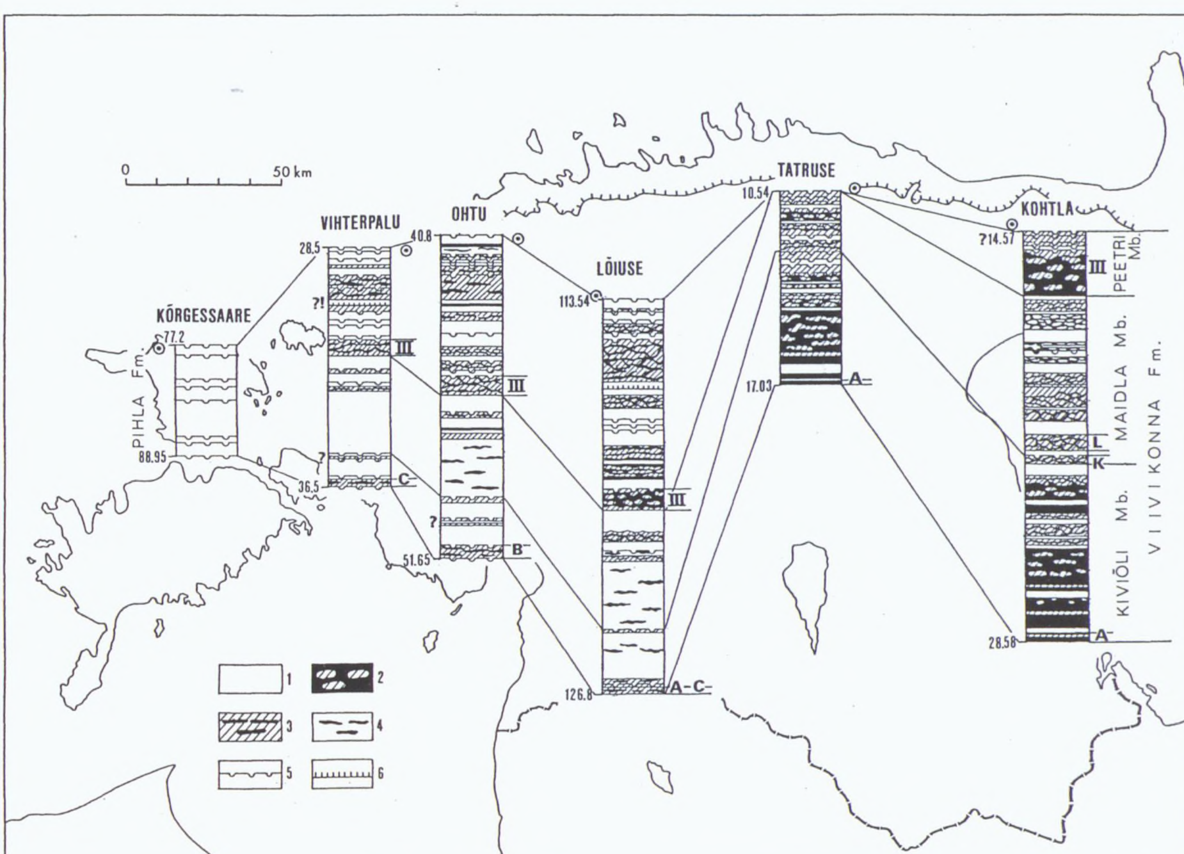


Fig. 42. The composition and correlation of the Kukruse Stage on the line A-B (see Fig. 41): 1 - pure limestone; 2 - kukersite with carbonate noduls; 3 - kerogeneous limestone with kukersite noduls; 4 - kukersite noduls; 5 - discontinuity surface; 6 - K-bentonite in the Vihterpalu and Lõiuse cores. Compiled after Männil and Bauert (Мянниль и Бауерт 1986). A, B, C, D, K, L, III - indices of kukersite seams.

Haljala Stage

Jaanusson (1995) proposed the term Haljala Stage for the unit which comprises the Idavere and Jõhvi chronostratigraphical subdivisions, previously regarded as separate stages. These two subdivisions, now classified as the Idavere and Jõhvi substages (Table 7), comprise most of K-bentonite beds which lie below the thickest bed ("d" by Юргенсон 1958a) established in eastern Baltic. The substages are difficult to differentiate in southern Estonia, in areas where K-bentonite beds are uncertain or absent. Also the faunal distinction between the substages is rather inconsiderable (Пылма и др. 1988, figs. 9-11). In Estonia, the thickness of the Haljala Stage varies mostly from 10 to 20 m (Fig. 43).

The Idavere Substage comprises the regularly bedded hard bioclastic limestones of the lower, Tatruse Formation and argillaceous limestones with intercalations of marls and some thin K-bentonites of the upper, Vasavere Formation (emended by Männil & Meidla 1994). This substage has the most reduced sequence in northern Estonia and in some places in the surroundings of Tallinn it is entirely missing (Jaanusson 1945). The Tatruse Formation (Пылма и др. 1988) corresponds roughly to Schmidt's (1881) original concept of "Itfersche Schicht" and the fauna recorded by him and his contemporaries from the "Itfer" belongs only to this formation. The Vasavere Formation contains usually two, but in the west sometimes up to 18 K-bentonite beds, which belong to the Grefsen complex of bentonites (Bergström *et al.* 1995). In the areas where only beds "a" and "b" (Юргенсон 1958a) are recognizable, the upper bed is regarded as the top of both the Idavere Substage (Мяннилъ 1963a) and the Vasavere Formation. In areas farther south where bentonite beds of the Grefsen complex disappear or in the west where they are numerous, a distinction between the Idavere and Jõhvi substages is difficult.

The Jõhvi Substage, which is at its thickest (more than 10 m, Fig. 44) in northwestern Estonia, comprises argillaceous bedded to nodular limestones with argillaceous intercalations in the middle part (Мяннилъ и Рыымусокс 1984, Пылма и др. 1988). These limestones form the lower part of the Kahula Formation (Männil & Meidla 1994). A fairly persistent K-bentonite bed (bed "c" by Юргенсон 1958a, "Sinsen K-bentonites" by Bergström *et al.* 1995) occurs close to the boundary between the middle and upper parts of the Jõhvi Substage.

In southern Estonia, the Haljala Stage, 8 - 18 m in thickness, is represented by argillaceous limestones with thin K-bentonite beds and in places with phosphatic ooids of the Adze Formation (Ульст и др. 1970).

In the outcrop area, the lower boundary of the Haljala Stage and the Tatruse Formation is formed by a conspicuous discontinuity surface – a hardground which in places is penetrated by cavities, some 5 cm or even more in diameter at the surface and extending sometimes some 40 cm downwards (Пылма и др 1988). The limestones above the basal discontinuity (Kisuvere Member) comprise up to 16% of quartz sand. The most detailed stratification of the lowermost beds of the Haljala Stage is based on chitinozoans.

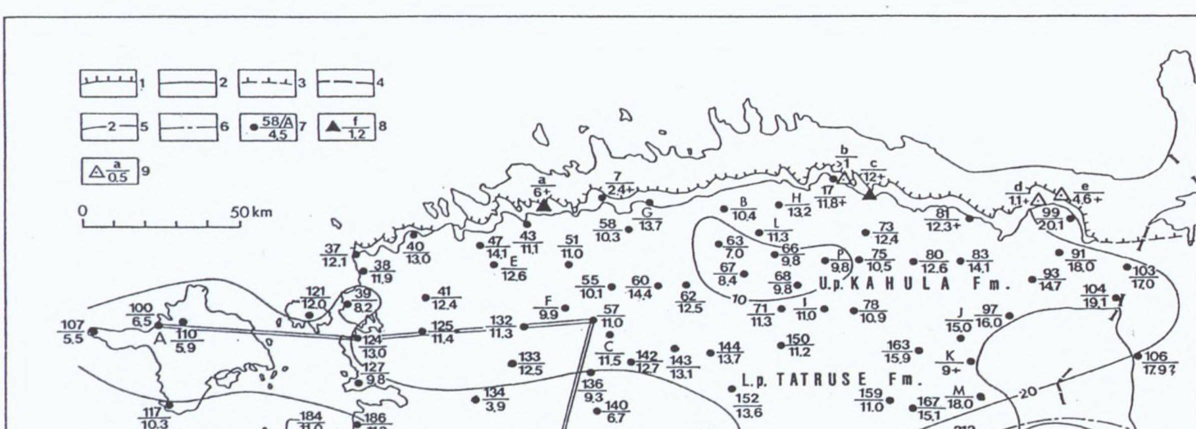


Fig. 43. Sketch-map showing the present distribution and the thickness of the Haljala Stage. Legend 1 - 9 see Fig. 30. Outcrops: a - Peetri, b - Idavere, type section of the Idavere Substage, c - Aluvere, d - Kahula, type section of the Kahula Formation, e - Kämbemägi, type section of the Jõhvi Substage; core sections: A - I see Fig. 44, J - Piilsi, K - Mustvee, L - Lehtse, M - Kose. Sections on the profile line A - A'see Fig. 40. The signs "L. p." and "U. p" show whether the unit forms the lower or upper part of the stage.

The oldest part of the stage, the *Armoricochitina granulifera* and *Angochitina curvata* zones (Мянниль 1986, fig. 5.1.1, Nõlvak & Grahn 1993) occurs in the Laeva area in eastern central Estonia (Fig. 45) where the stage has the maximum thickness (ca. 25 m, Fig. 44, core No. 285). These two zones do not occur in the northenmost sequences, where the *Lagenochitina dalbyensis* Zone forms the basal part of the substage.

The gap on the boundary between the Kukruse and Haljala stages is rather well expressed by differences between faunas, especially of brachiopods and trilobites in northern Estonia. Both these groups are represented in the Idavere Substage with about 40 species, a few of which occur also in the underlying Kukruse Stage (Рыымусокс 1970, table 12). The occurrence of several Kukruse bryozoans in the upper part of the Idavere Substage (in the Vasavere Formation) is seemingly of facies origin. The lower boundary of the Haljala Stage is also marked by rather sharp changes in the composition of ostracodes (Пылма и др. 1988, figs. 7, 9-11), though some typical Idavere - Jõhvi species, *e.g. Braderupia asymmetrica* (Neckaja) appear in the top of the Kukruse Stage. The frequent occurrence of *Leiosphaeridia* above it, is a rather good marker for the lower boundary of the Haljala Stage. The base of the Haljala Stage is close to both the graptolite *Diplograptus multidens* Zone and the conodont *Baltoniodus gerdae* Subzone (Männil 1990, Jaanusson 1995).

The changes in the faunal composition on the transition between the Idavere and Jõhvi substages are continuous which is clearly revealed in core sections, especially by ostracodes (Пылма и др. 1988). Several new macrofossil taxa, such as *Toxochasmops maximus* (Schmidt), *Clinambon anomalus* (Schlotheim), presumably appear somewhat higher (1.5 - 2 m, Мянниль 1963а, б) of the boundary K-bentonite bed between the Idavere and Jõhvi substages.

In the core sections located far away from the outcrop area, the alga *Mastopora concava* (Eichwald), spicules of *Pyritonema subulare* (Roemer) and also some brachiopods (*Bilobia*) occur (Fig. 40), but there is no characteristic species among macrofossils for determination of the lower boundary of the Haljala Stage.

Keila Stage

In most of northern Estonia, the Keila Stage (Kegelsche Schicht, Schmidt 1881) comprises the argillaceous bioclastic limestones, with intercalations or occasionally thicker (up to 4 m) intervals of relatively pure limestones of the Kahula Formation (Table 7). Only in a restricted area in northwestern Estonia, the upper part of this formation is replaced by the Vasalemma Formation where the greatest thickness of the Keila Stage (more than 30 m) has been recorded (Fig. 46).

Initially, due to the unclear relationship between the fossilifereous argillaceous limestones of the Keila Stage and the Schmidt's "Wassalem'sche Schicht", the term Keila-Vasalemma Stage was introduced by Bekker (1922, see also

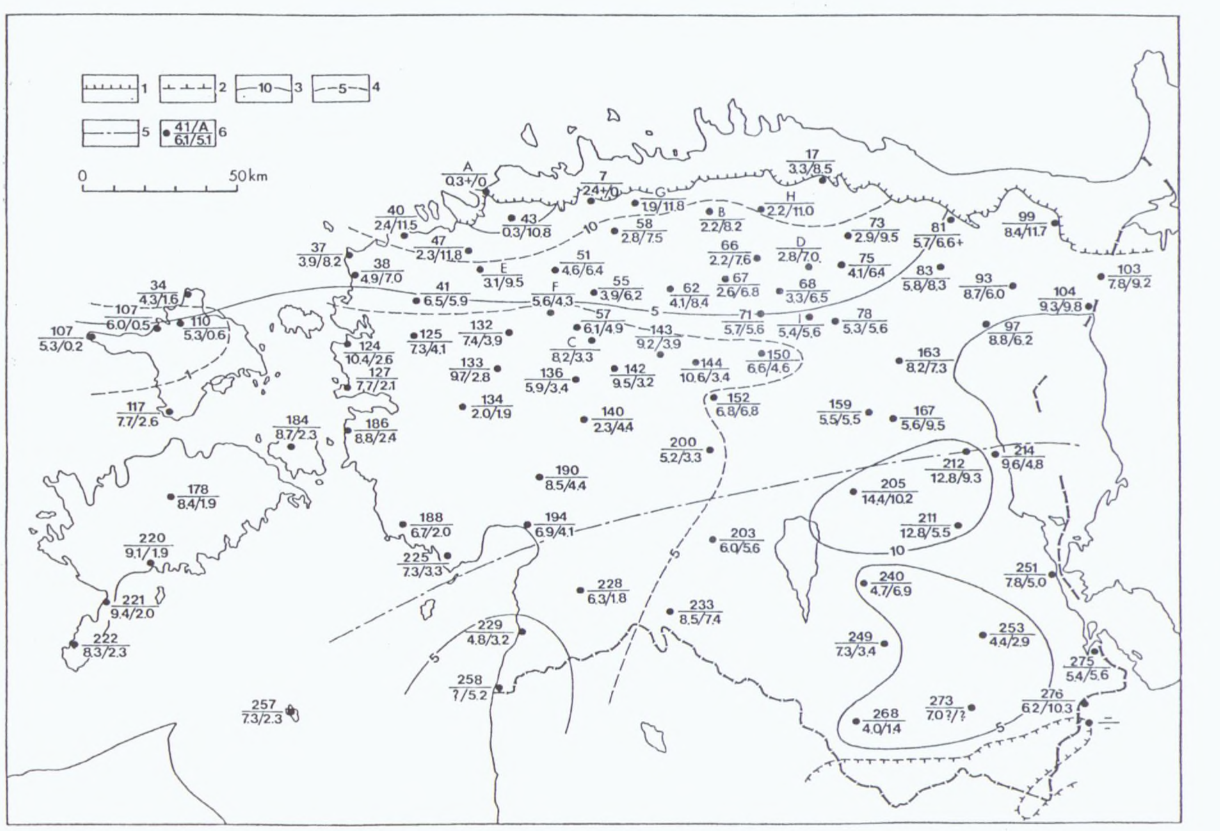


Fig. 44. Sketch-map showing the present distribution of the Haljala Stage and the thickness of the Idavere and Jõhvi substages: 1 and 2 - limits of the distribution of the Haljala Stage in the outcrop and subsurface area; 3 - isopachyte of the Idavere Substage; 4 - isopachyte of the Jõhvi Substage; 5 - approximate northern boundary of the Adze Formation; 6 - core section, number or index in the numerator, thickness of the Idavere (first number in the denominator) and Jõhvi (second number in the denominator) substages.

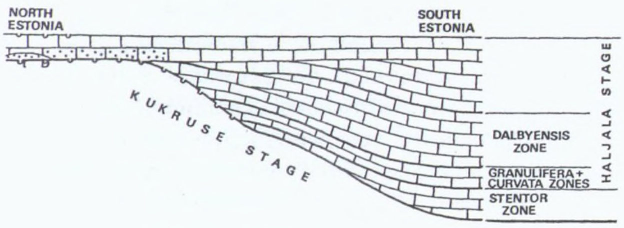


Fig. 45. Meridional cross-section of the lower Idavere Substage and distribution of the chitinozoan zones (Hints *et al.* 1994).

Öpik 1930b). Later, Jaanusson (1945) and Männil (Мянниль 1958в, 1963б, 1966) subdivided the Keila Stage into several members and defined the lower boundary of the stage on the level of the thickest K-bentonite (bed "d" by Юргенсон 1958а, see also Jaanusson & Martna 1948, Вингисаар 1972). The composition of the Kahula Formation and the distribution of members overlying the boundary K-bentonite is shown in Figure 47.

The lowermost part of the Keila Stage (Kurtna Member) is represented by argillaceous limestones. The Kurtna Member is overlain by relatively pure limestones, in places with argillaceous intercalations of the Pääsküla Member. This unit, although differently understood by stratigraphers (Nõlvak 1996), can be identified in the core sections of northwestern Estonia as a complex of biomicritic limestones, up to ca. 7 m in thickness (Пылма и др. 1988, Fig. 47). It may be replaced by intercalating argillaceous bioclastic and biomicritic limestones with a thickness of up to 20 m (Ainsaar 1991), seemingly corresponding to a longer time interval than the Pääsküla Member in the sense of Põlma *et al.* (Пылма и др. 1988).

The younger part of the Keila Stage comprises the Saue and Lehtmetsa members, the fossiliferous argillaceous limestones and detrital marls with thin layers of argillaceous limestones, respectively. Contemporaneously, the formation of carbonate buildups (interpreted as reefs, Raymond 1916, or bioherms, Мянниль 1960, or mud mounds, Põlma & Hints 1984) has been developed in northwestern Estonia. They belong to the Vasalemma Formation, nearly constituting the upper half of the Keila Stage in the surroundings of Keila - Vasalemma, whereas a distinct eastward shift of the corresponding facies is recorded during late Keila time (Fig. 48). The Vasalemma Formation, in thickness up to 15 m, consists of several principal lithotypes (Мянниль 1960, Пылма 1967, Hints 1996). The most characteristic type of rock is the bedded grainstone (cystoid limestone), which is intercalated with clayey limestones in the lower part of the formation. Cystoid limestone consists mainly of skeletal sand particles aggregated with pure calcite cement (content of terrigenous material less than 3%). The grainstones contain irregular buildups, measuring up to 10 m vertically and up

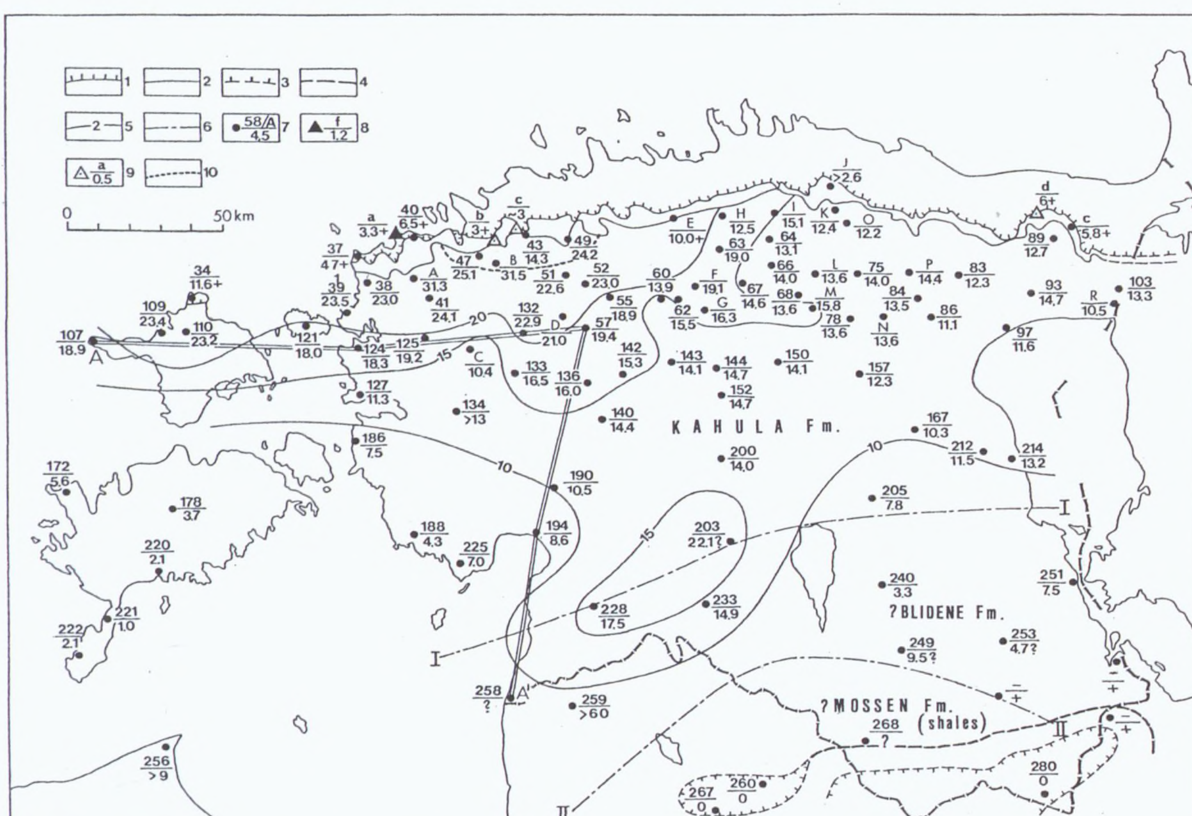


Fig. 46. Sketch-map showing the present distribution and the thickness of the Keila Stage. Legend 1- 9 see Fig. 30; 10 - southern border of the Vasalemma Formation. Outcrops: a - Ristna b - Vasalemma, type section of the Vasalemma Formation, c - Keila, type section of the Keila Stage, d - Kahula, type section of the Kahula Formation, e - Rausvere; core sections: A - Kaldamäe, B - Munalaskme, C - Maidla, D - Oela, E - Kehra, F - Atla, G - Mustla, H - Pillapalu, I - Palaoja, J - No 269, K - No 150, L - Tamsalu, M - Raigu, N - No 542, O - Tõrma, P - Viru-Roela, R - Permisküla. Sections on the line A-A' see Fig. 40. Line I - I - approximate southern boundary of the Kahula Formation, II - II - northern border of the distribution of black shales of the Mossen Formation, which supposedly may have in places the Keila Age (see Meidla 1996).

to 300 m horizontally and consisting of pure limestones with a low content of skeletal sand (less than 10%) and terrigenous material (up to 6%), occasionally with inclusions of fossiliferous marls. The buildups mostly lack the reef-like framework and are considered as carbonate mounds. Still, in some "mounds" the edrioasteroid *Cyathocystis rhizophora* Schmidt is frequent and may form frame-like structures. The lower and middle parts of the Vasalemma Formation contain a number of species common with the Kahula Formation - *Estlandia pyron silicificata* Öpik, *Clinambon anomalus* (Schlotheim), *Horderleyella? kegelensis* (Alichova), *Sowerbyella (S.)* cf. *forumi* Rõõmusoks a.o., which indicate the Keila Age of corresponding rocks.

In southern Estonia, the Keila Stage comprises the upper part of the bioclastic limestones of the Adze Formation and clayey limestones and marls, which contain some species common with the Blidene Formation in Latvia (Ульст и др. 1982). The lowermost part of the Mossen Formation may also be of Keila Age (Table 7, Meidla 1996). Still, the correlation of the Lukštai and Blidene formations with a unit of siltstone and silty limestone identified in southern Estonia (Ainsaar 1995) needs to be adjusted. Due to this uncertainty, the identification of the Keila Stage is complicated in the transition between the distribution areas of the Kahula and Adze formations.

The total thickness of the Kahula Formation may exceed 30 m, and in northwestern Estonia its main part corresponds to the Keila Stage. In general, the thickness of the Keila Stage part of the formation (mostly 10-15 m) decreases in the southeast direction. In the same direction, the formation becomes lithologically more uniform and argillaceous. In southern Estonia, the thickness of the equivalents of the Keila Stage presumably does not exceed 10(?) m.

A rich and diverse fauna of bryozoans, brachiopods, trilobites, echinoderms and other sedentary and vagile groups (see Рыымусокс 1970) is distributed in the Kahula Formation. In the upper part of the formation, corresponding to the Keila Stage, several macrofossil taxa are common with the Haljala Stage, but a specific component in this particular association comprises last representatives of several brachiopod genera (*Clinambon, Cyrtonotella*), trilobites (*Asaphus (Neoasaphus) nieszkowskii* Schmidt and *Toxochasmops maximus* (Schmidt)), crinoids (*Ristanacrinus marinus* Öpik and different baltocrinids) or species characteristic of the Keila Stage only *Keilamena occidens* (Männil), *Longvillia asmusi* (Verneuil), *Horderleyella? kegelensis* (Alikhova). The data on macrofauna come mostly from northern Estonia. In southern Estonia, the Keila Stage is characterised by a brachiopod - trilobite association, which comprises several taxa (*Skenidioides, "Sampo", Eoplectodonta*), appearing on a higher stratigraphical level in northern Estonia or being related to the Scandinavian faunas.

The Keila Stage presumably corresponds to the uppermost part of the *Diplograptus multidens* and the lowermost part of the *Dicranograptus clingani* graptolite zones (Männil 1990). The lower boundary of the stage, the level of the K-bentonite bed "d" corresponds to the lower boundary of the chitinozoa *Angochitina multiplex* Subzone (Table 7) and is close to the Northern Atlantic conodont *superbus* Zone (Männik & Viira 1990).

Oandu Stage

In northern Estonia, the Oandu Stage comprises rocks of two different lithofacies forming the Vasalemma and Hirmuse formations. The Vasalemma Formation, distributed in northwestern Estonia, consists of bedded fine- to coarse-grained bioclastic limestones with irregular bodies of aphanitic massive limestones (carbonate buildups). These rocks were identified first as the *Hemicosmites* Limestone (Eichwald 1854) or Wasalemm'sche Schicht (Schmidt 1881). Vasalemma, as the name of a chronostratigraphic unit was applied also to the rocks of another lithofacies – the argillaceous limestones and marls, named Oandu beds by Öpik (1933) and the Hirmuse Formation by Männil and Rõõmusoks (Мянниль и Рыымусокс 1984), which are exposed on the banks of the Oandu River in northeastern Estonia (Рыымусокс 1953, Аалоэ 1958). Later studies

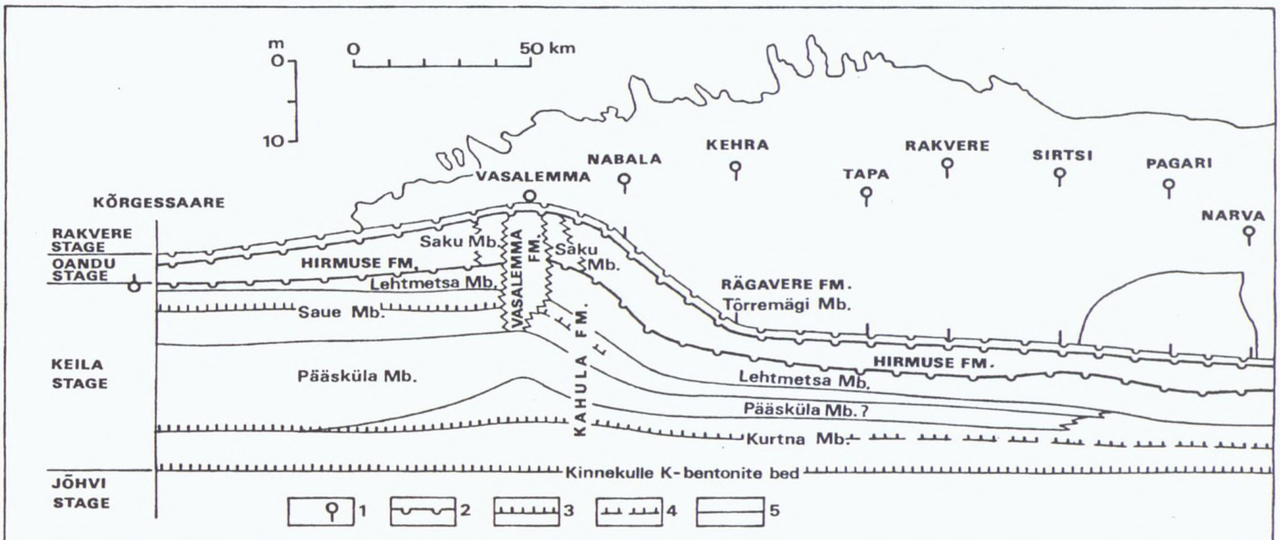


Fig. 47. Profile of the Keila and Oandu stages in northern Estonia from west (Kõrgessaare) to east (Narva), compiled by Põlma *et al.* Пылма и др. 1988). Legend: 1 - core section; 2 - discontinuity surface; 3 - K-bentonite; 4 - discontinous K-bentonite; 5 - boundary between units.

(Мяннилъ 1958д, 1960) showed that the lower and middle parts of the Vasalemma Formation are of Keila Age (Fig. 48) and the name Oandu was proposed for chronostratigraphic unit of post-Keila Age. The Oandu Age of the uppermost Vasalemma Formation is presumed by the appearance of the corals *Lyopora tulaensis* (Sokolov), *Eofletcheria orvikui* Sokolov and the brachiopods *Rhynchotrema*? *parva* Oraspõld, *Rostricellula nobilis* (Oraspõld), *Dactylogonia luhai* (Sokolskaya) (Мяннилъ 1960, Рыымусокс 1970), or it is supposed by the disappearance of *Leiosphaeridia* and brachiopods of the Keila Stage (Rummu core, Пылма и др. 1988). In northern Estonia, the Oandu Stage is restricted in thickness (1-4 m, Fig. 49); only in the limits of the Vasalemma Formation it is up to 6 m thick.

In the stratotype area in northeastern Estonia, the lower boundary of the Oandu Stage and the Hirmuse Formation, is known only by the core sections where it is marked by a sharp discontinuity surface with up-to-35-cm-deep pockets, on the upper boundary of the Kahula Formation (= Kahula Group, Männil & Meidla 1994). Below this level, a great number of Middle Ordovician species and even genera common with several older stages, including the brachiopods *Cyrtonotella, Estlandia,* trilobites *Asaphus (Neoasaphus) nieszkowskii* Schmidt, *Pseudobasilicus*, ostracodes *Tetrada (Tetrada) harpa* (Krause), *Polyceratella spinosa* Sarv (Fig. 50) disappear. Notable is the disappearance or sharp decrease in the frequency of the acritarch *Leiosphaeridia* which is abundant in the Keila Stage (Fig. 51). This fossil seemingly can be used for the preliminary establishing of the abovementioned boundary in core sections, especially when the uppermost part of the Kahula Formation is more argillaceous and possibly belongs to the Lehtmetsa Member of late Keila Age (Fig. 47, Пылма и др. 1988, fig. 32). A new complex of fossils with the ostracodes *Bolbina rakverensis* (Sarv), *Klimphores minimus* (Sarv), *Disulcina perita perita* (Sarv), brachiopods *Howellites wesenbergensis* (Alichova), *Equirostrata wesenbergensis* (Teichert) appears near the lower boundary of the Oandu Stage (Fig. 50, see also Пылма и др. 1988). In some cases, these species are found even below the boundary discontinuity surface, seemingly they occur in the deep pockets filled with deposits of Oandu Age. Due to the essential changes in the faunal composition (Мяннилъ и др. 1966, Hints *et al.* 1989), several authors have suggested to use the lower boundary of the Oandu Stage as the regional subseries or series boundary (Jaanusson 1945, Рыымусокс 1956а).

The Hirmuse Formation thins out within a rather short distance in the southern direction, and in many places in central Estonia the Oandu Stage is represented only by the Tõrremägi Member of the Rägavere Formation with a thickness less than one metre (Fig. 49). This area separates the rich and diverse fauna of bryozoans, brachiopods, echinoderms and trilobites of the Hirmuse Formation in northern Estonia (Пылма и др. 1988) from the relatively rich brachiopod and trilobite fauna in the marls and argillaceous limestone in southern Estonia, corresponding presumably to the Lukštai Formation. Beside some species common for northern and southern Estonia (*Howellites wesenbergensis* Alichova, *Rhactorthis kaagverensis* Hints a.o.), several brachiopods (*Reushella magna* Hints, *Laticrura* sp. *Skenidioides* sp., *Leptellina*? sp.) have been identified only in the latter region and are also known in the Lukštai Formation in Lithuania or in the Moldå Limestone in Sweden (Jaanusson 1982, fig. 7).

Identification of the Oandu Stage is most complicated in southeastern Estonia where the black shales and the overlying marls of the Mossen Formation are distributed (Karula

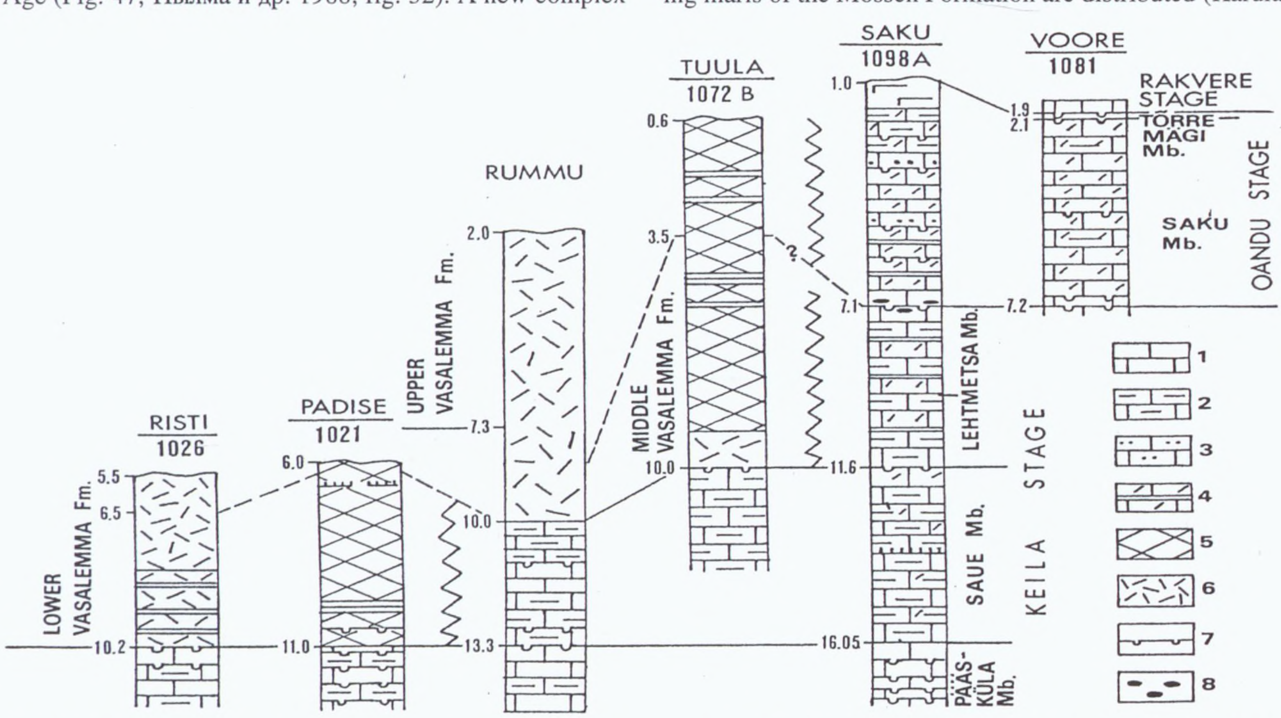


Fig. 48. Correlation of the Keila and Oandu stages in the distribution area of the Vasalemma Formation (Hints 1996). Legend: 1 - limestone; 2 - argillaceous limestone; 3 - sandy limestone; 4 - organodetrital limestone with argillaceous interlayer; 5 - massive limestone of the mud mound; 6 - massive organodetrital limestone (cystoid limestone); 7 - discontinuity surface; 8 - pebbles. Other signs in the log see Fig. 40.

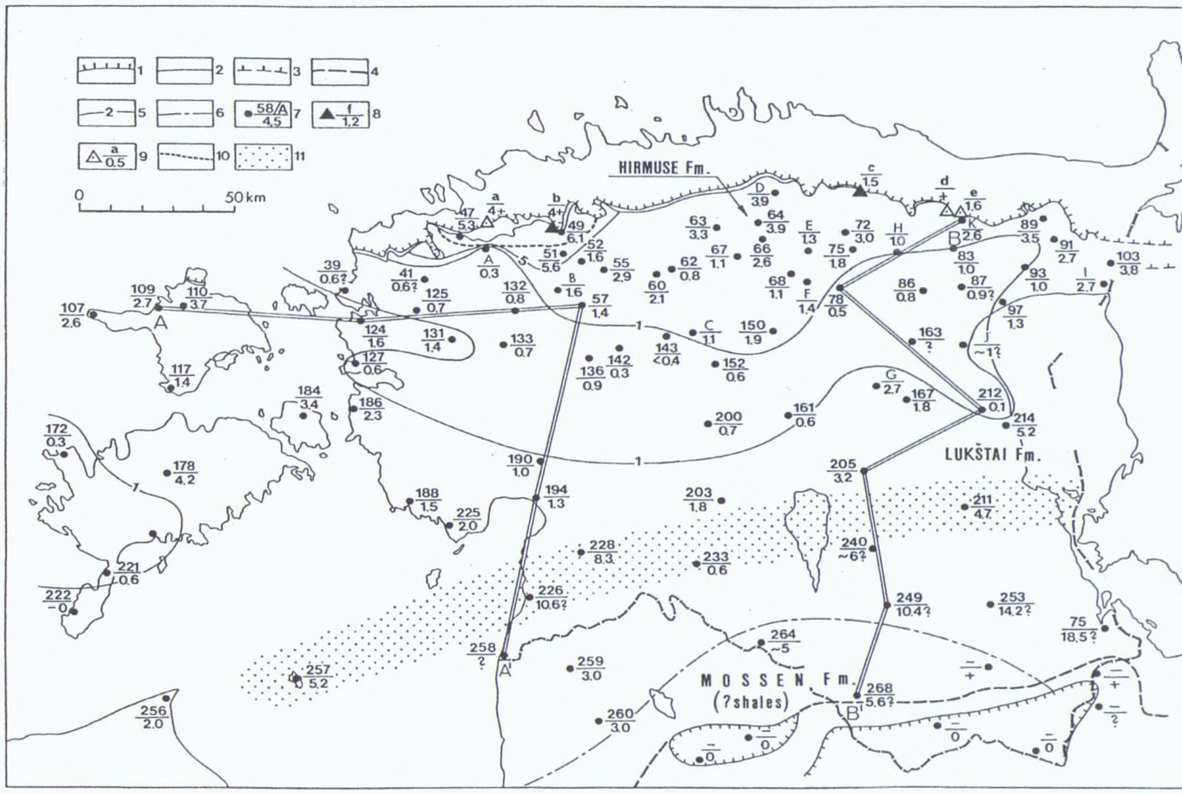


Fig. 49. Sketch-map showing the present distribution and the thickness of the Oandu Stage. Legend 1- 9 see Fig. 30; 10 - boundary of the distribution area of the Vasalemma Formation; 11 - distribution area of the sandy limestones in the boundary beds of the Keila and Oandu stages (Ainsaar 1995). Outcrops: a - Vasalemma, type section of the Vasalemma Formation, b - Saku, c - Rakvere (Ussimägi), d - Hirmuse, type section of the Hirmuse Formation, e - Oandu, type section of the Oandu Stage; core sections: A - Munalaskme, B - Oela, C - Paide, D - Palaoja, E - Tamsalu, F - Raigu, G - Ellamaa, H - Viru-Roela, J - Permisküla. Sections on the profil line A-A‘ see Fig. 40, on line B-B‘ - Fig. 51.

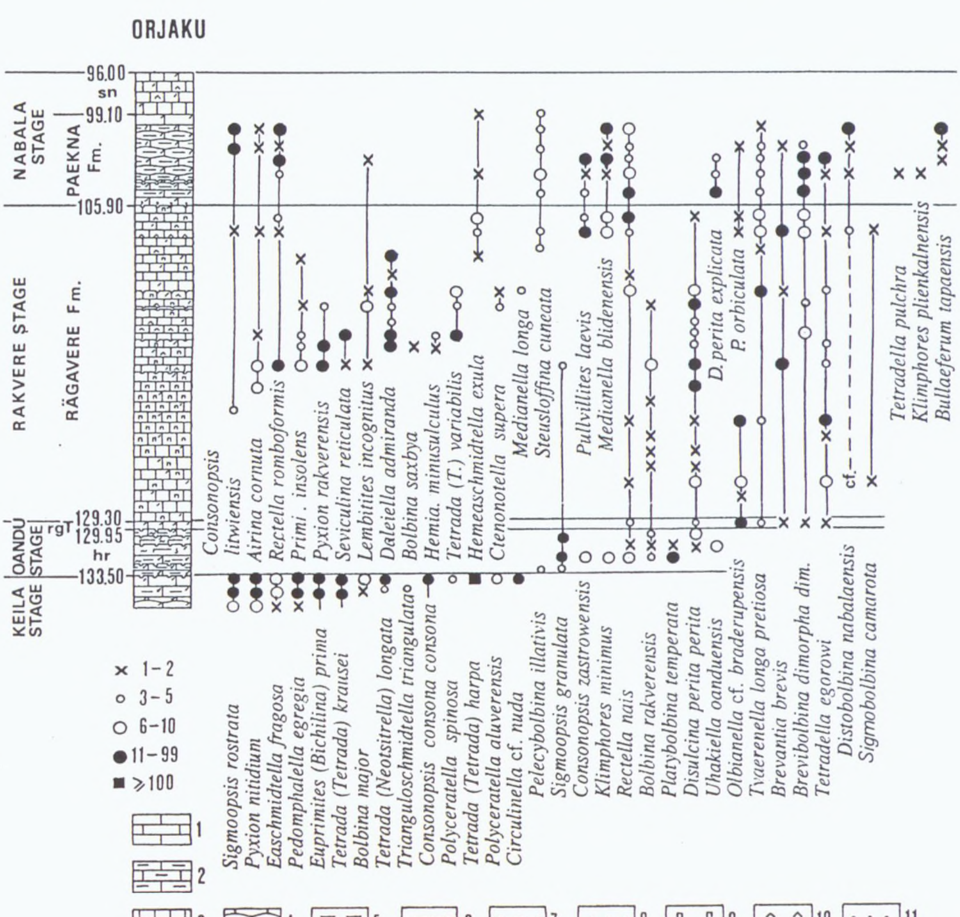


Fig. 50. The distribution of selected ostracodes in the Orjaku core section (by Meidla 1996). Legend: 1 - limestone; 2 - argillaceous limestone; 3 - micritic (aphanitic) limestone; 4 - argillaceous nodular limestone; 5 - carbonate marl; 6 - discontinuity surface; 7 - skeletal detritus; 8 -pyritized skeletal detritus; 9 - pyrite mottles; 10 - kukersite kerogen; 11 - glauconite. The symbols showing the approximate number of specimens are given in the lower left corner. Indices of stratigraphical units: rgT - Tõrremägi Member of the Rägavere Formation, hr - Hirmuse Formation, sn - Saunja Formation.

core in Fig. 51). The shales, encountered in various sections have been included to the Keila (Meidla 1996), Oandu (Хинтс 1975) or Rakvere (Мяннилъ 1966) Stage. The overlying marls of the Priekule Member are correlated with the uppermost Oandu and/or the Rakvere Stage. In some sections, the marls below the black shale (about 4 m in the Karula core, Fig. 51) comprise brachiopods known in the other sections (Otepää, Laeva) mainly in the beds presumably of Oandu Age. In favour of this age testifies also the disappearance of *Leiosphaeridia* at a depth of about 3 m below the shales. At the same time, the ostracode record allows to suppose Keila Age of the lowermost part of the Mossen Formation (Meidla 1996). The contradiction in interpreting the age by macrofossils and ostracodes may be caused by insufficient data available, or it may indicate the patchy distribution of deposits in the Keila - Oandu boundary interval in southeastern Estonia.

Still, in most of southern Estonia, the Oandu Stage can be identified most realiably on the basis of ostracodes. The lower boundary of the Oandu Stage is marked by the appearance of *Sigmoopsis granulata* Sarv, *Bolbina rakverensis* Sarv, *Pelecybolbina illativis* Neckaja and *Klimphores minimus* (Sarv), and a general rapid faunal change which occurred throughout the Estonian part of the palaeobasin (Meidla 1996).

Rakvere Stage

The "Wesenbergsche Schicht" by Schmidt (1881) corresponds roughly to the Rakvere Stage in nowadays understanding (Мяннилъ 1958б, 1963а, Кырвел 1962, Пылма и др. 1988). In northern Estonia, the Rakvere Stage forms the lowermost, relatively thick part of Late Viru and Harju pure micritic (aphanitic) limestones which intercalate with more or less argillaceous varieties. The cycles of different lithotypes generally constitute distinct lithostratigraphical units (Пылма 1982, Hints *et al.* 1989), whereas the clayey parts of the cycles are characterized by the appearance of abundant new taxa .

The Rakvere Stage consists of the Piilse and Tudu members (Кырвел 1962) which form the main part of the Rägavere Formation. The stage is at its thickest (28 m) in western Estonia (Fig. 52) and it thins notably in the southeastern direction. The lower, Piilse Member with a thickness of up to 27 m (Рыымусокс 1983) consists of pure, in places dolomitized micritic limestones with a low content of terrigenous material (3 - 9%) and skeletal sand (less than 5%, Кырвел 1962, Пылма и др. 1988). The member is characterised by a distinct pyritic pattern, following abundant burrows in the former sediments. The upper, Tudu Member is up to 10 m thick and differs from the Piilse Member in the higher content of skeletal sand (commonly 15 %, Пылма и др. 1988), in the occurrence of thin, up to 3 cm thick kukersite layers and rare and weakly developed pyritic patterns.

Southwards the limestones of the Rägavere Formation become more argillaceous and in southern Estonia they are supposedly replaced by the carbonate marls of the Priekule Member in the upper part of the Mossen Formation (Männil & Meidla 1994, Meidla 1996). On the basis of chitinozoan distribution (Nõlvak & Grahn 1993) it is supposed that in some places (Ruhnu and Ohesaare cores) the Rakvere Stage is missing.

The data on the distribution and composition of macrofossils in the Rakvere Stage, particularly in the Tudu Member is scanty due to relatively few outcrops. The Rakvere Stage with the relatively sparse macrofauna of bryozoans, brachiopods and trilobites is characterized by frequent and diverse association of calcareous algae (*Cyclocrinites, Rhabdoporella etc.*, Kõrts *et al.* 1990). From the Rakvere Stage up to the end of Ordovician, calcareous algae and their fragments dominate in the composition of skeletal particles (Пылма 1972, 1982) where they may account even for 97.6%.

Unlike macrofossils, the ostracode record of the Rakvere Stage is rich, comprising more than 80 species (Meidla 1996). Several distinct associations have been recorded in the lower part of the stage corresponding approximately to the Piilse Member of the Rägavere Formation (Meidla 1996, fig. 47). Valuable is the ostracode record of the upper part of the stage which contains only sparse macrofauna. This interval, nearly equal to the Tudu Member, corresponds to the *Daleiella admiranda* Subzone (subzone of *Daleiella* sp. n. in Meidla & Sarv 1990, Table 10), a range zone prominent in the sections of northern and central Estonia. The appearance of several long-ranging taxa, such as *Steusloffina cuneata* (Steusloff), *Medianella blidenensis* (Gailite),

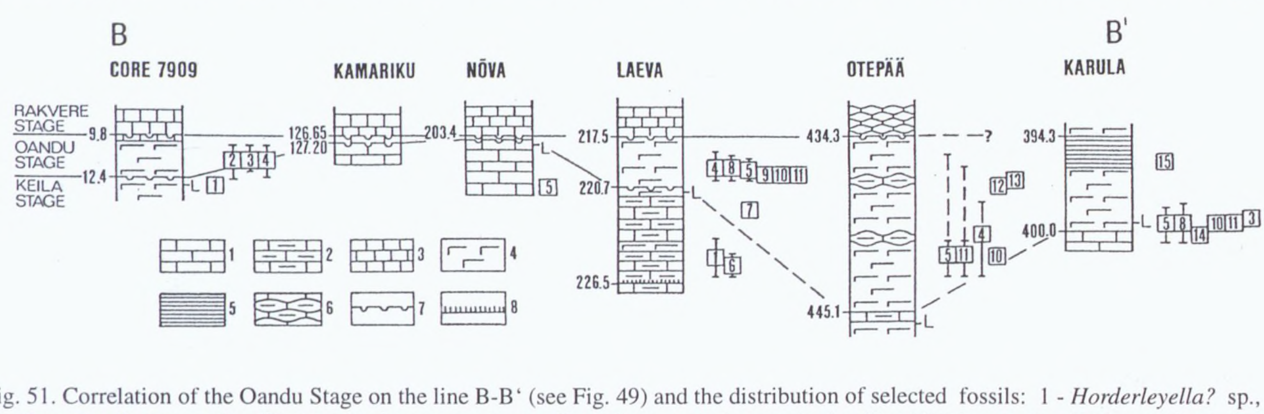


Fig. 51. Correlation of the Oandu Stage on the line B-B' (see Fig. 49) and the distribution of selected fossils: 1 - *Horderleyella?* sp., 2 - *Sowerbyella tenera*, 3 - *Rhactorthis kaagverensis*, 4 - *Howellites wesenbergensis*, 5 - *Skenidioides* sp., 6 - *Keilamena*? sp., 7 - *"Leptaena"* sp., 8 - *Leptellina*? sp., 9 - *Camerella* sp., 10 - *Platystrophia rava*, 11 - *Onniella longa*, 12 - *Reushella magna*, 13- *"Leptaena"* sp., 14 - *Oanduporella* sp., 15 - *Onniella* sp. (cf. *bamcrofti*); l - uppermost finds of *Leiosphaeridia*. Legend: 1 - limestone; 2 - argillaceous limestone; 3 - micritic (aphanitic) limestone; 4 - marl; 5 - black shale; 6 - argillaceous nodular limestone; 7 - discontinuity surface; 8 - K-bentonite.

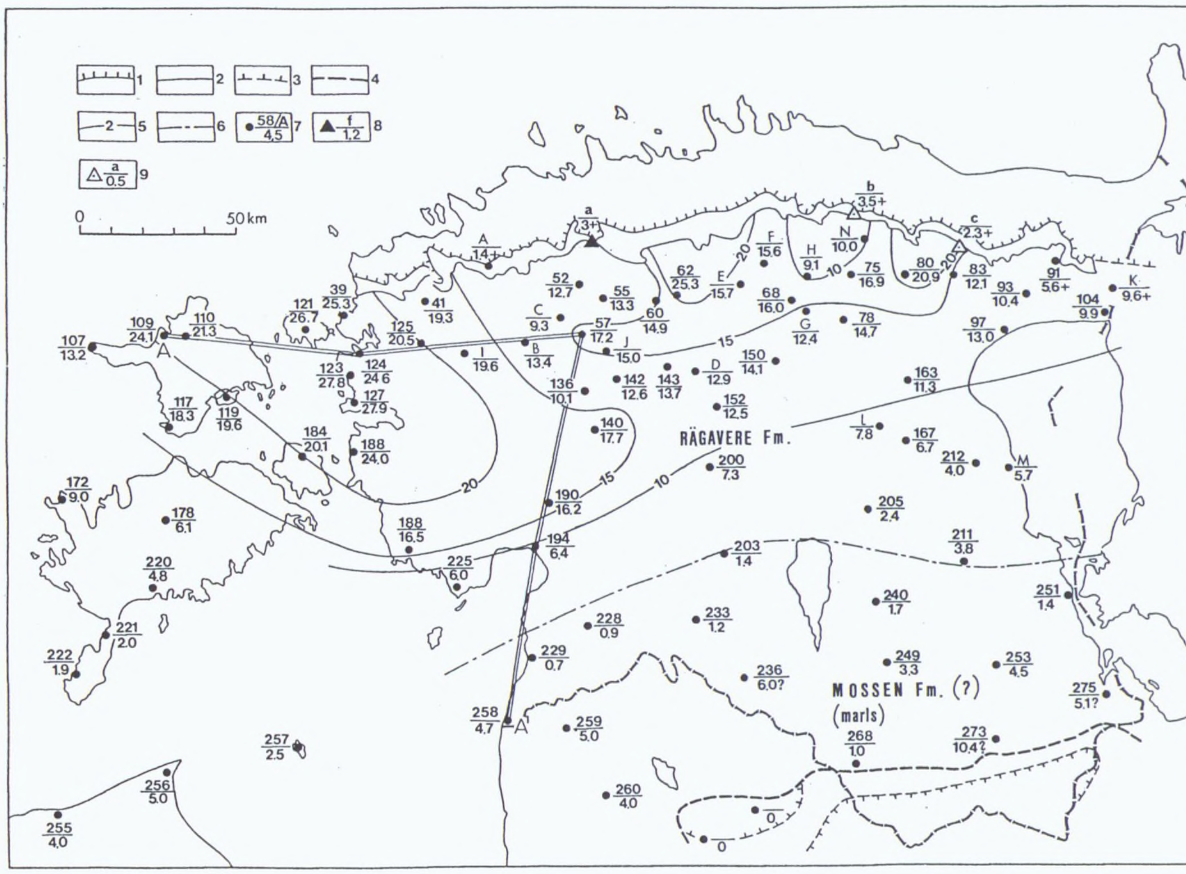


Fig. 52. Sketch-map showing the present distribution and the thickness of the Rakvere Stage. Legend 1- 9 see Fig. 30. Outcrops: a - Nabala, b - Rägavere, type section of the Rakvere Stage and Rägavere Formation, c - Piilse on the Oandu River bank, type section of the Piilse Formation; core sections: A - Munalaskme, B - Vaimõisa, C - Oela, D - Väätsa, E - Järva-Madise, F - Ambla, G - Raigu, H - Tamsalu, I - Risti-Maidla, J - Keava, K - Kuningaküla, L - Ellavere, M - Alatskivi, N - Koeravere. Sections on the profile line A - A' see Fig. 40.

Pullvillites laevis (Abushik & Sarv), *etc*., has been recorded within this interval (Fig. 50).

The lower boundary of the Rakvere Stage is lithologically more or less distinct in northern Estonia where it coincides with the pyritized discontinuity surface on the top of the Tõrremägi Member in the lower part of the Rägavere Formation. The appearance of several new brachiopods, including *Microtrypa estonica* Rõõmusoks, *Platystrophia lutkevichi satura* Oraspõld, *P. quadriplicata* Alichova, *Sowerbyella (Sowerbyella) raegaverensis* Rõõmusoks, *Vellamo wesenbergensis* (Pahlen) and trilobites *Chasmops wesenbergensis* (Schmidt), *Encrinuroides seebachi* (Schmidt), *Pharostoma pediloba* (Roemer) and others, above this boundary shows the renovation of faunal associations. However, macrofossils are very scarce, particularly in core sections, and cannot be used for the purposes of detailed biostratigraphy. The same applies to ostracodes, because most of the species characteristic of the Late Ordovician ostracode fauna appear below this boundary, in the Tõrremägi Member of the Oandu Stage which represents a facies similar to the Rakvere Stage. Also the zonal chitinozoa *Fungochitina fungiformis*, characteristic of the Rakvere and Nabala stages, appears in the Tõrremägi Member. The suggestion to include the Tõrremägi Member to the Rakvere Stage (Meidla 1996) follows partly the earlier wider interpretation of that stage, according to which the Oandu beds by Öpik were included to the Rakvere Stage (Jaanusson 1945, Алихова 1960).

The Rakvere Stage corresponds roughly to the lower part of the graptolite *Pleurogratus linearis* Zone and the chitinozoan *Cyathochitina angusta* Subzone of the *Fungochitina fungiformis* Zone (Nõlvak & Grahn 1993, Table 7).

HARJU SERIES

Nabala Stage

The Nabala Stage was distinguished by Männil (Мяннилъ 1958б) as the lower part of the Schmidt's (1858) "Lyckholm'sche Schicht". He subdivided the stage into the lower, Paekna and the upper, Saunja substages (see also Öpik & Laasi 1937, Jaanusson 1994). Nowadays they are used as lithostratigraphical units (formations) which represent the Nabala Stage with a total thickness of 10 to 35 m in northern and partly in central Estonia (Fig. 53). The Paekna Formation is up to 16 m thick and comprises predominantly argillaceous bioclastic limestones intercalating with micritic limestones. The thickness of micritic interlayers is usually about 0.1 m, but occasionally it may reach 1-3 m. The up-to-28-m-thick micritic limestones of the Saunja Formation are lithologically uniform and occur all over Estonia. Their thickness decreases towards the south until it is only 0.3 m (Meidla 1996). South of the Muhu - Mustvee line, the lower Paekna Formation is replaced by the Mõntu Formation. This 3–7-m-thick complex consists of argillaceous bioclastic limestones with rare thin (5-30 cm) layers of micritic limestones containing glauconite (Oraspõld 1995).

On the transition from the Rakvere to the Nabala Stage the shelly fauna undergoes notable renovation. Of about 150 species and subspecies occurring in the Nabala Stage, only one third is common with the Rakvere Stage (Мяннилъ и др. 1966). Several new species of brachiopods, including *Bekkeromena semipartita* (Roemer), *Ilmarinia sinuata* (Pahlen), *Laticrura rostrata* Hints, *Sulevorthis lyckholmiensis* (Wysogorski), *Pseudolingula quadrata* (Eichwald), appear in the Nabala Stage. Some of these species are missing in the Saunja Formation, but appear again in the overlying Vormsi Stage. The micritic limestones of the Saunja Formation contain a notably abundant and diverse fauna of molluscs (about 30 gastropod and more than 10 cephalopod species).

The lower boundary of the stage is exposed only in the Paekne quarry (Nõlvak & Meidla 1990). It is marked by a series of uneven discontinuity surfaces, above which there appears a new association of chitinozoans, including the zonal *Armoricochitina reticulifera* (Grahn). The latter can be used as the most reliable fossil for the identification of the lower boundary of the Nabala Stage in the core sections in central and southern Estonia.

The composition of ostracodes changes remarkably on the Rakvere - Nabala transition. Several new taxa, including *Disulcina perita explicata* Sarv, *Tetrada neckajae* Meidla, *Oepikella luminosa* Sarv a.o., appear in the lower part of the Nabala Stage, but mainly somewhat higher of the lower boundary of the Paekna Formation (Fig. 50). For this reason the lower boundary of the stage is marked better in the ostracode record by the disappearance of the species *Disulcina perita perita* (Sarv) and *Daleiella admiranda* Meidla (a zonal species) in the uppermost part of the Rakvere Stage (Meidla 1996). Brachiopods are found mostly in the lower Paekna Formation and among them new faunal elements *Pseudolingula quadrata* (Eichwald) and *Sulevorthis lyckholmiensis* (Wysogorski) appear close to the lower boundary of the Nabala Stage (Fig. 54).

The Nabala Stage corresponds to the middle part of the *Pleurograptus linearis* graptolite Zone (Table 7) and the upper part of the North Atlantic *superbus* conodont Zone. In the ostracode record, the summary differences between the Paekna and Saunja formations are not significant, but the very uneven distribution of ostracodes in the Saunja Formation should be mentioned (Meidla 1996).

Vormsi Stage

The Vormsi Stage (Jaanusson 1944b, = middle part of the Lyckholm'sche Schicht, Schmidt 1858) consists of a facies succession of bioclastics limestones (Kõrgessaare Formation, up to 21 m) in northern Estonia, argillaceous limestones with glauconite (Tudulinna Formation, up to 17.1 m) in central Estonia and black shales (Fjäcka Formation, up to

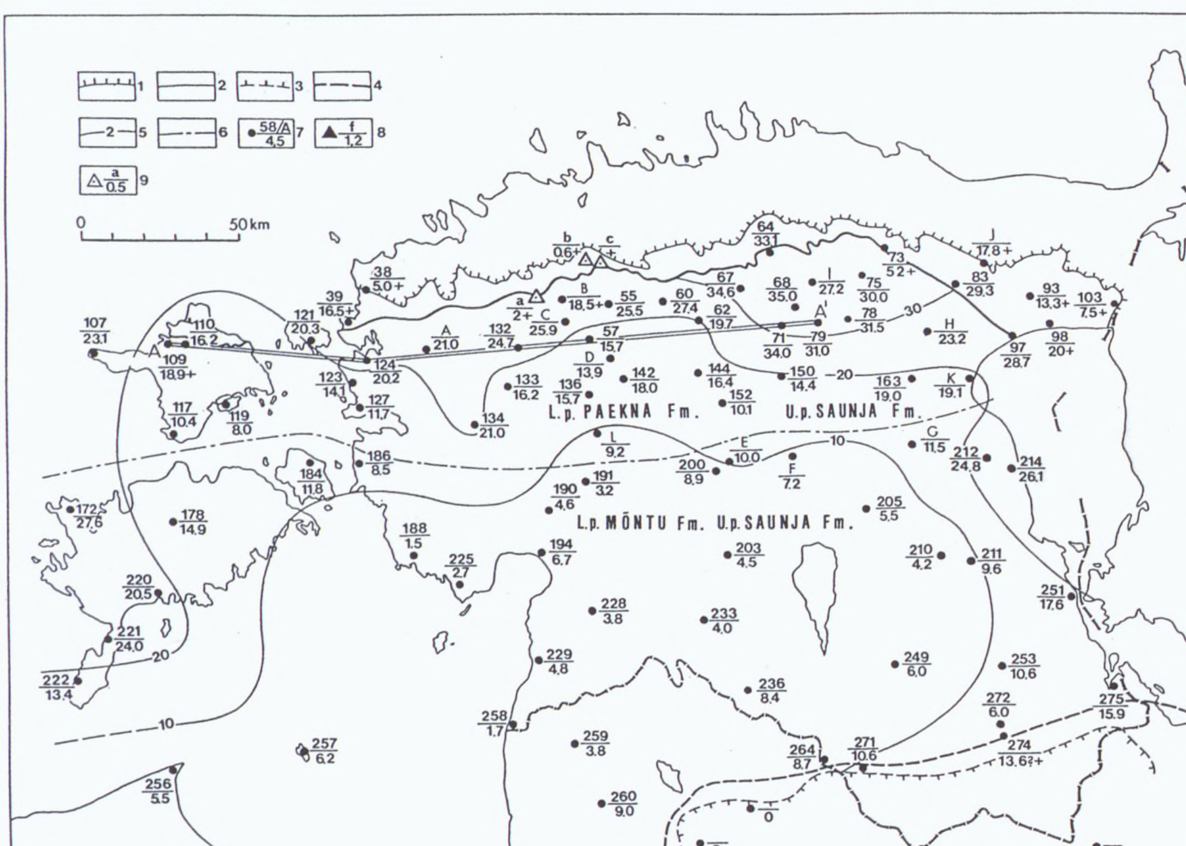


Fig. 53. Sketch-map showing the present distribution and the thickness of the Nabala Stage. Legend 1- 9 see Fig. 30. Outcrops: a - Mõnuste (Saunja), type section of the Saunja Formation; b - Paekna, type section of the Paekna Formation, c - Nabala (Nõmmeküla), type section of the Nabala Stage, core sections: A - Palivere, B - Adila, C - Oela, D - Keava, E - Koksvere, F - Põltsamaa, G - Palamuse, H - Venevere, I - Tamsalu, J - Piilsi, K - Mustvee, L - Kõnnu. The signs "L. p." and "U. p" show whether the unit forms the lower or upper part of the stage.

4.5 m) in southern Estonia. The thickness of the stage decreases from 10 - 20 m in northern Estonia to 0.3 m in southern Estonia (Ohesaare core, Fig. 55). In the transitional area between the Kõrgessaare and Tudulinna formations, interfingering of these units can be followed (Ораспыльд 1982a, figs. 3, 4).

The association of the diverse shelly fauna of corals, bryozoans, brachiopods, molluscs and trilobites includes some 200 species (Рыымусокс 1967) in northern Estonia in the Kõrgesssaare Formation. Southwards this fauna is replaced by a specific and less diverse association.

The Tudulinna Formation is characterized by an association of brachiopods, comprising species of the genera *Dicoelosia, Christiania, Skenidioides, Leptellina*?, and a facies dependent ostracode association prevailed by *Uhakiella curta* Sidaraviciene, *Medianella blidenensis* (Gailite) and *Rectella nais* Neckaja (Meidla 1996). The Fjäcka Formation comprises a brachiopod association typical of shally facies, consisting mainly of inarticulated small brachipods *Paterula, Hisingerella* a.o., and of a few articulate brachiopods, such as *"Sericoidea"* and *Onniella* (Fig. 56). In general, the association is similar to that in the Mossen Formation.

The lower boundary of the Vormsi Stage coincides with a lithologically sharp boundary in most of Estonia. Above that boundary the frequency of macrofossils and ostracodes increases notably. In the Kõrgessaare Formation, several new species appear, including the corals *Proheliolites dubius* (Schmidt), *Kenophyllum siluricum* (Dybowski), *Streptelasma (Streptelasma) distinctum* Wilson, the brachiopods *Eoplectodonta schmidti* (Lindström), *Equirostra gigas* Schmidt, *Sampo hiiuensis* Öpik, *Glyptorthis plana* Oraspõld, *Triplesia insularis* (Eichwald), the first *Dicoelosia* (Wright 1968) and the trilobites *Encrinurus moe* Männil a.o. Ostracodes are dominated by the species common with the older strata (Fig. 57). The earliest conodonts of the *ordovicicus* Zone occur also in the Vormsi Stage (Männik1992b).The zonal species *Amorphognathus ordovicicus* has been recorded from the basal part of the Kõrgessaare Formation and also from the Tudulinna Formation.

In spite of distinct lower and upper boundaries of the Vormsi Stage, the detailed correlation of the formations belonging to this stage is not yet very clear. The distribution of zonal chitinozoans allows to suppose that the oldest part of the Vormsi Stage is missing in central and southern Estonia (Nõlvak & Grahn 1994). In these areas the Vormsi Stage corresponds to the *Tanuchitina bergstroemi* Zone which forms the upper part of the stage in northern Estonia, overlying the *Fungochitina fungiformis* Zone (Table 7) of the lower part of the stage in this area.

The distribution of ostracodes (Meidla 1996) does not support the correlation schemes based on chitinozoans.

The topmost part of the Vormsi Stage correponds to the chitinozoa *Acanthochitina barbata* Subzone (Table 7). The level of the disappearance of the index species marks well the traditional upper boundary of the Vormsi Stage.

Pirgu Stage

The Pirgu Stage (Jaanusson 1944b) is a lithologically variable (Table 7) and thick (up to 66 m, Fig. 58) stratigraphical unit (=upper part of the "Lyckholm'sche Schicht", Schmidt 1858). The notable changes in the thickness of the stage, sometimes within a short distance, are due to various reasons, *e.g.* the development of mud mounds, denudation during late Pirgu and/or Porkuni time, intensive tectonical movements and changes in sea-level.

In northern Estonia, within the stage two succesive rock units of grey-coloured limestones are distinguished - the lower, Moe and the upper, Adila Formation (Рыымусокс 1960) which correspond approximately to the former Nyby (Мяэнниль 1966, Jaanusson 1944b) and Piirsalu (Jaanusson 1945) substages.

The Moe Formation, up to 40 m in thickness, consists of micritic and bioclastic nodular or bedded limestones with argillaceous intercalations. The lower part of the formation contains abundant calcareous alga *Palaeoporella? (=Dasyporella* in Мяэнниль 1966) and in some places (Hoitberg on Vormsi Island, Võhma core in central Estonia) typical carbonate mounds are developed, quite similar to the Boda mounds in the Siljan district of Sweden.

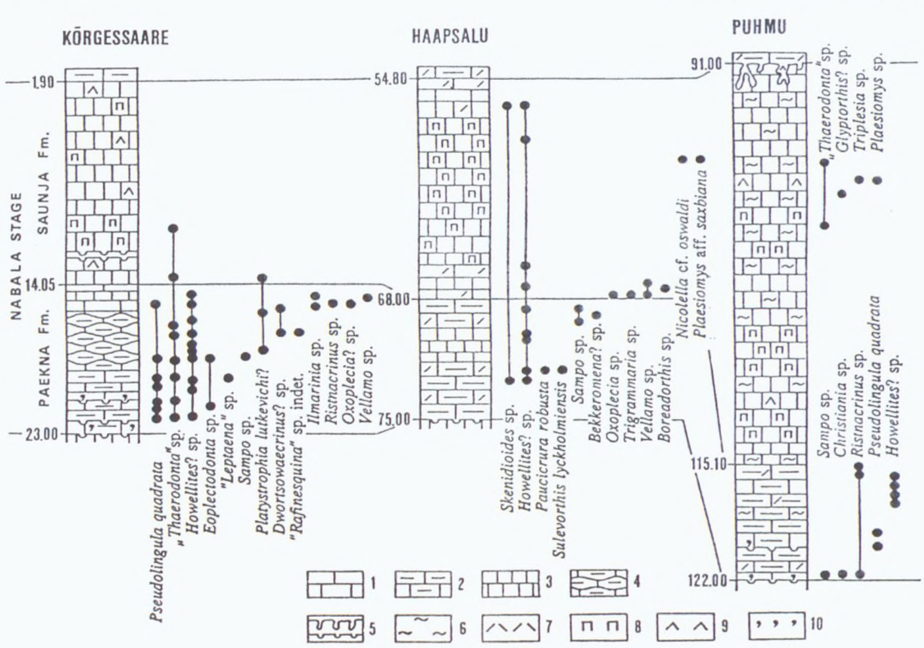


Fig. 54. Correlation of the Nabala Stage on the line A-A' (see Fig. 53) and the distribution of selected macrofossils. Legend: 1 - limestone; 2 - argillaceous limestone; 3 - micritic (aphanitic) limestone; 4 - nodular argillaceous limestone; 5 - discontinuity surface; 6 - burrows; 7 - skeletal detritus; 8 - pyrite mottles; 9 - kukersite kerogen; 10 - glauconite.

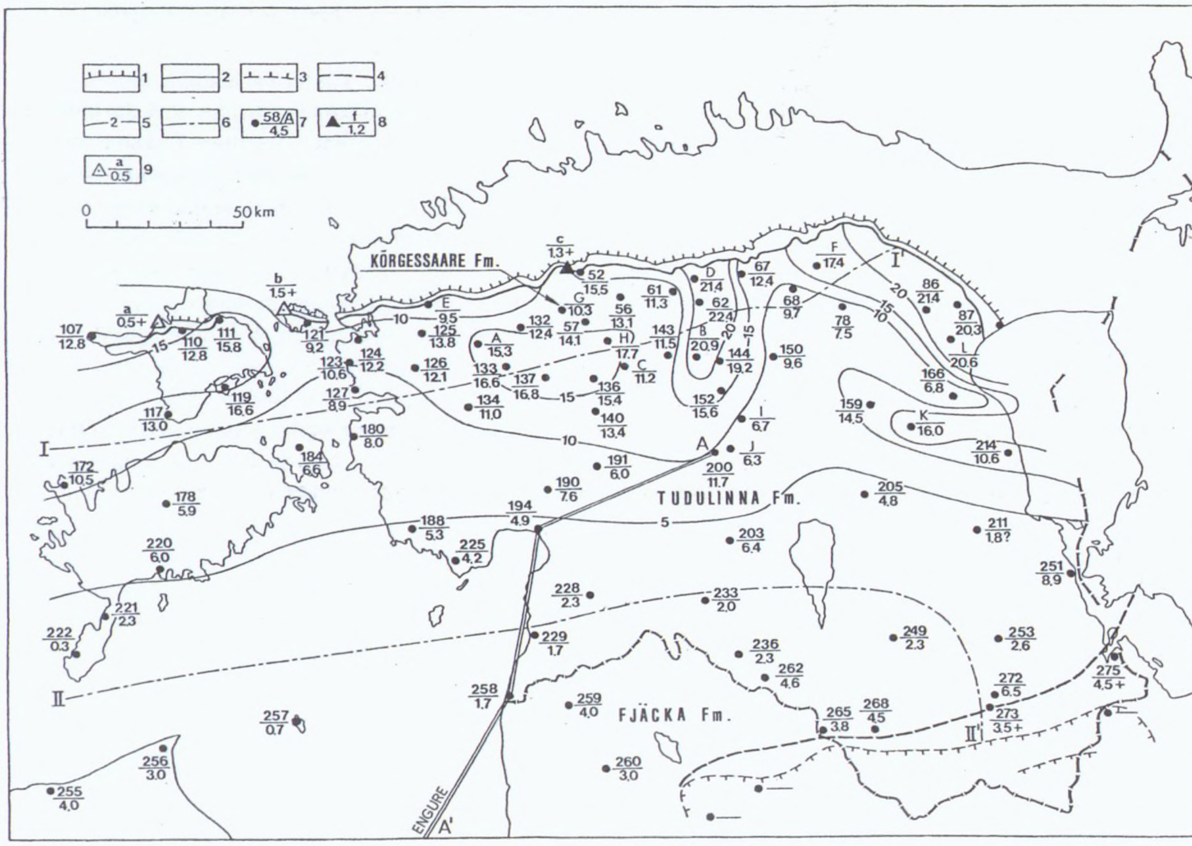


Fig. 55. Sketch-map showing the present distribution and the thickness of the Vormsi Stage. Legend 1- 9 see Fig. 30. Outcrops: a - Kõrgessaare, type section of the Kõrgessaare Formation, b - Saxby, type section of the Vormsi Stage, c - Urge; core sections: A - Maidla, B - Väätsa, C - Lelle, D - Atla, E - Seljaküla, F - Tamsalu, G - Oela, H - Keava, I - Kabala, J - Koksvere, K - Palamuse, L - Piilsi, 78 - Kamariku, type section of the Tudulinna Formation. Line I-I'- northern limit of the distribution of the Tudulinna Formation, II- II'- northern limit of the distribution of the Fjäcka Formation. Sections on the line A-A' see Fig. 56.

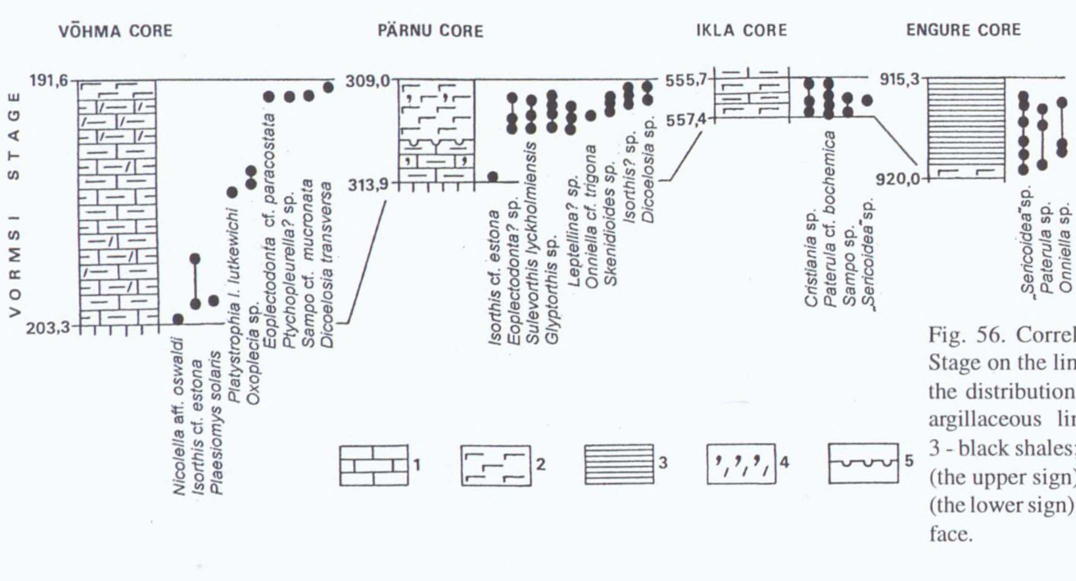


Fig. 56. Correlation of the Vormsi Stage on the line A-A' (Fig. 55) and the distribution of macrofossils: 1 - argillaceous limestones; 2 - marls; 3 - black shales; 4 - glauconite grains (the upper sign) and skeletal detritus (the lower sign); 5 - discontinuity surface.

The Adila Formation comprises predominantly bioclastic limestones with a thickness of 10-15 m. Numerous discontinuity surfaces and cyclically alternating pure and argillaceous limestones are characteristic to the upper part of the formation. In this topmost part of the formation, the pentamerid brachiopod *Holorhynchus* has been recorded in the Island of Hiiumaa (Hints 1993). The boundary between the Moe and Adila formations coincides with the boundary between the *rugata* and *bergstroemi* chitinozoan zones (Table 7).

Within a rather large area in central Estonia, the Pirgu Stage is characterized by the interfingering of different rock units (Орасмыльд 1975a, Table 7), whose correlation with the northernmost and southernmost sequences is complicated. In this transitional area, the lowermost part of the stage consists of argillaceous bioclastic limestones with glauconite (Tootsi Member), overlying the upper part of the Vormsi Stage, corresponding to the chitinozoan *barbata* Zone. The

Tootsi Member contains a distinct association of shelly fauna (Fig. 59). This unit is succeeded upwards by the grey-coloured, sometimes red mottled marls and highly argillaceous limestones of the main part of the Halliku Formation whose relationship with the Moe Formation is not yet very clear (Männil & Meidla 1994). The diverse and abundant ostracode fauna of the Halliku Formation comprises taxa (Fig. 60) common with the upper part of the Moe Formation (Meidla 1996). Most uncertain is the age of the Kabala Member on the transition between the Pirgu and Porkuni stages in central and westernmost Estonia. According to the ostracode record, the formation contains two different associations. The older association comprises the Pirgu species *Brevibolbina pontificans* Schallreuter, *Bullaeferum tapaensis* (Sarv), whereas the species *Apatochilina falacata* Sarv and *Gryphiswaldensia plicata* Schallreuter from the supposedly younger part are common with the fossiliferous part of the Ärina Formation of the Porkuni Stage.

In southern Estonia, in the limits of the central confacies belt, the Pirgu Stage is represented by red-coloured or mottled argillaceous limestones and mudstones of the Jonstorp and Jelgava formations, including the Kuili Member (Table 7).

According to the traditional understanding, the lower boundary of the Pirgu Stage coincides with the lower boundary of the Moe Formation in northern Estonia. This level is underlain by the chitinozoan *Acanthochitina barbata* Zone. The zone has also been established under the red-coloured limestones of the Jonstorp Formation (Ruhnu core). On this basis, the lower boundary of the latter formation has been equalized with the lower boundary of the Pirgu Stage in southern Estonia. In terms of graptolite zonation, the stage boundary corresponds to the base of the *D. complanatus* Zone (Männil 1990). The presence of *Climacograptus supernus* Elles *et* Wood in the upper part of the Pirgu Stage suggests its correlation with the *Dicellograptus anceps* Zone (Мяннил 1976, Männil 1990).

The lower boundary of the Pirgu Stage is poorly reverberated in the distribution of most shelly fossil groups. As a rule, the new faunal elements appear 1 - 2 m above (or even higher) of the distinct lithological changes (Fig. 59). In northern Estonia, the biostratigraphical boundary is best expressed

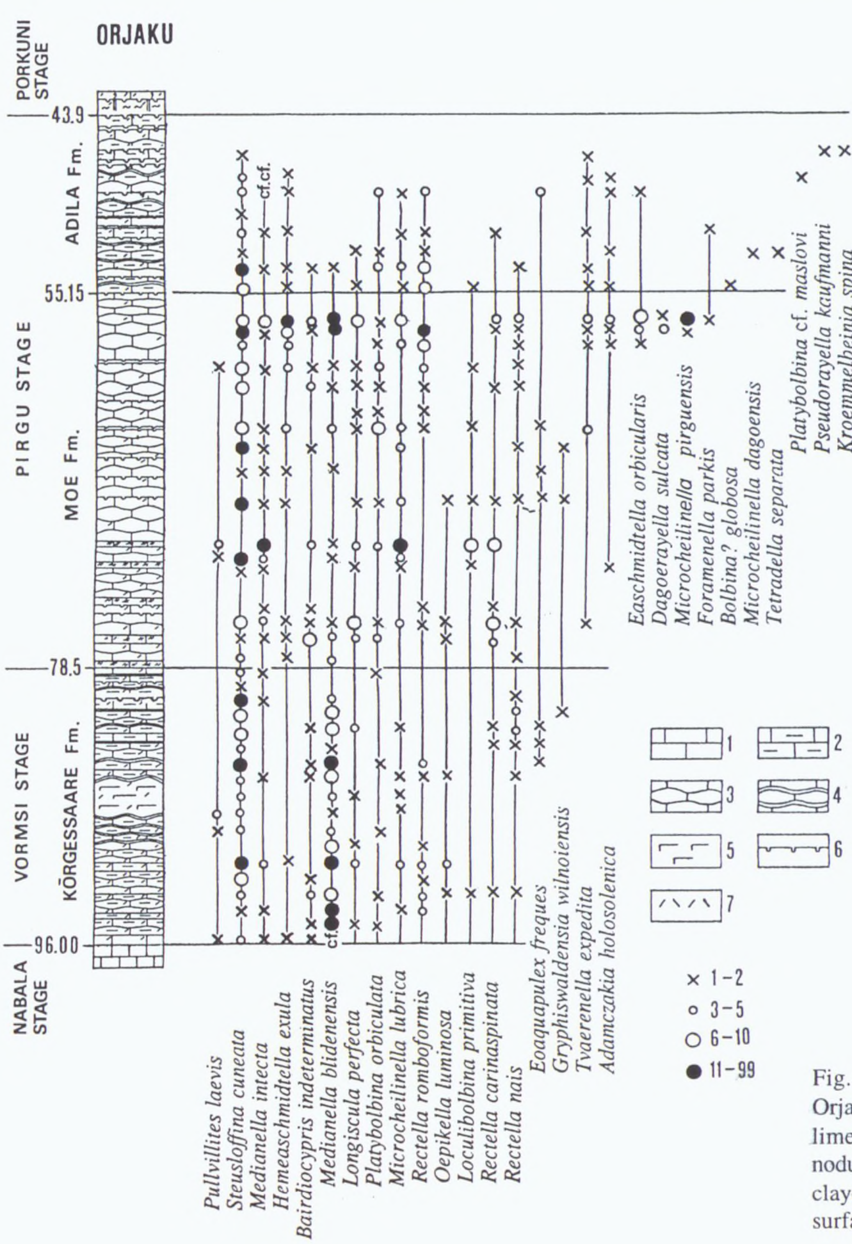


Fig. 57. Distribution of ostracodes in the Orjaku core (by Meidla 1996). Legend: 1 - limestone; 2 - argillaceous limestone; 3 - nodular limestone; 4 - nodular limestone with clayey interbeds; 5 - marl; 6 - discontinuity surface; 7 - skeletal detritus.

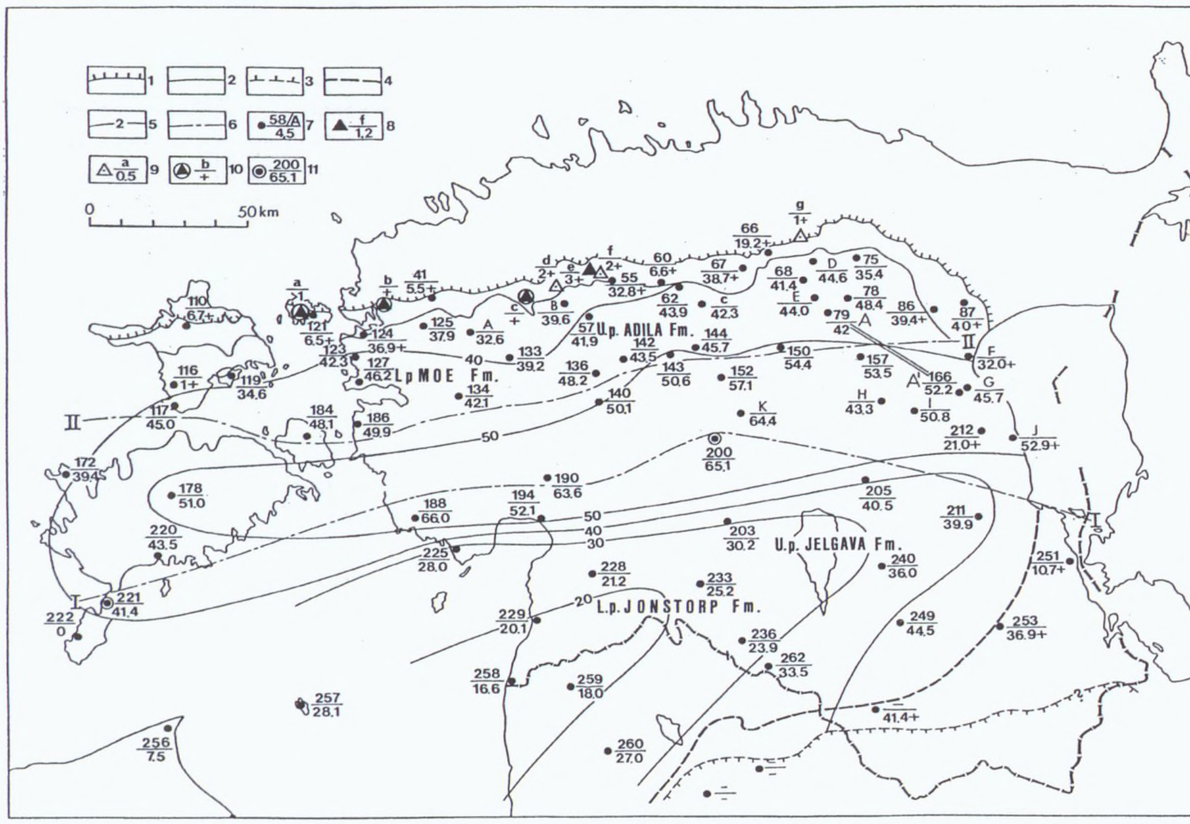


Fig. 58. Sketch-map showing the present distribution and the thickness of the Pirgu Stage. Legend 1- 9 see Fig. 30; 10 - outcrop of mud mound and reef, index in the numerator, thickness in the denominator; 11 - the same in the core section. I - I - northern limit of the distribution of the red-coloured and mottled limestones (Jonstorp and Jelgava formations), II - II - southern limit of the distribution of the sections of North Estonian type (Moe and Adila formations). Sections on the line A-A' see Fig. 59. Outcrops: a - Hoitberg, b - Niibi, c - Ruunavere, d - Adila, e - Lohu, f - Pirgu, type section of the Pirgu Stage, g - Moe, type section of the Moe Formation; core sections: A - Risti, B - Oela, C - Mustla, D - Tamsalu, E - Raigu, F - Mustvee, G - Kose, H - Ellavere, I - Palamuse, J - Alatskivi. The signs "L. p." and "U. p" show whether the unit forms the lower or upper part of the stage. Sections on line A - A' see Fig. 59.

by the appearance of the concentrations of the alga *Palaeoporella*?. In central and southern Estonia, the faunal renovation is best revealed in the ostracode record (Meidla 1996, see also Fig. 60).

The Pirgu Stage comprises three different assemblages of macrofauna, related to different facies zones of the palaeobasin. In northern Estonia, in the grey-coloured Moe and Adila formations a rich assemblage of large articulated brachiopods *Plaesiomys solaris* (Buch), *Equirostra gigas* Schmidt, *Triplesia insularis* (Eichwald), *Luhaia vardi* Rõõmusoks, corals *Sarcinula, Catenipora, Palaeofavosites*, and stromatoporoids, together with different molluscs, is distributed (Fig. 59). In the ostracode composition nonpalaeocopes are dominating: the associations of *Steusloffina cuneata-Medianella blidenensis* and *S. cuneata-Olbianella fabacea* (Meidla 1996, Fig. 60) occur. In central Estonia, in the Halliku Formation the most common representatives of the shelly fauna seem to be brachiopods and rugose corals (Fig. 59). In the red-coloured deposits of southern Estonia, only a few macrofossils, mainly brachiopods and trilobites, have been recorded, while trilobite and echinoderm fragments dominate in the skeletal sand (Мянниль и др. 1968).

Porkuni Stage

The Porkuni Stage (Borkholm'sche Schicht by Schmidt 1881) represents the topmost Ordovician stage (Raymond 1916, Bekker 1922). Up to the 1960s, the stage was included to the Silurian System (Öpik 1930b, 1934, Аалоэ и др. 1958, Алихова 1960, Рыымусокс 1960). In Estonia, the Porkuni Stage is represented by variable deposits of shallow-water facies (Мянниль 1966; Ораспыльд 1975б, 1982б; Rõõmusoks 1983), with a thickness of about 10 m in northern and up to 18 m in southern Estonia (Fig. 61). In northern Estonia, the stage is supposedly represented by its older part only, because afterwards, during late Porkuni time, this area turned into dry land as a result of the glacioeustatic sea-level lowering (Ораспыльд 1975б).

In northern Estonia, the Porkuni Stage is represented by the Ärina Formation comprising a succession of dolomites (Röa Member), stromatoporoid-tabulate reefs (with surrounding facies) and oolitic or sandy limestones (Kamariku Member) in the top. The assignment of the Röa Member (0.5-5.5 m of dolomites) has been problematic over the years. Some researchers have assigned it to the Pirgu Stage (Рыымусокс 1991), others to the Porkuni Stage (Rosenstein 1943, Jaanusson 1956, Решения… 1987). The unit, usually poor in fossils, yields some species common with the Pirgu

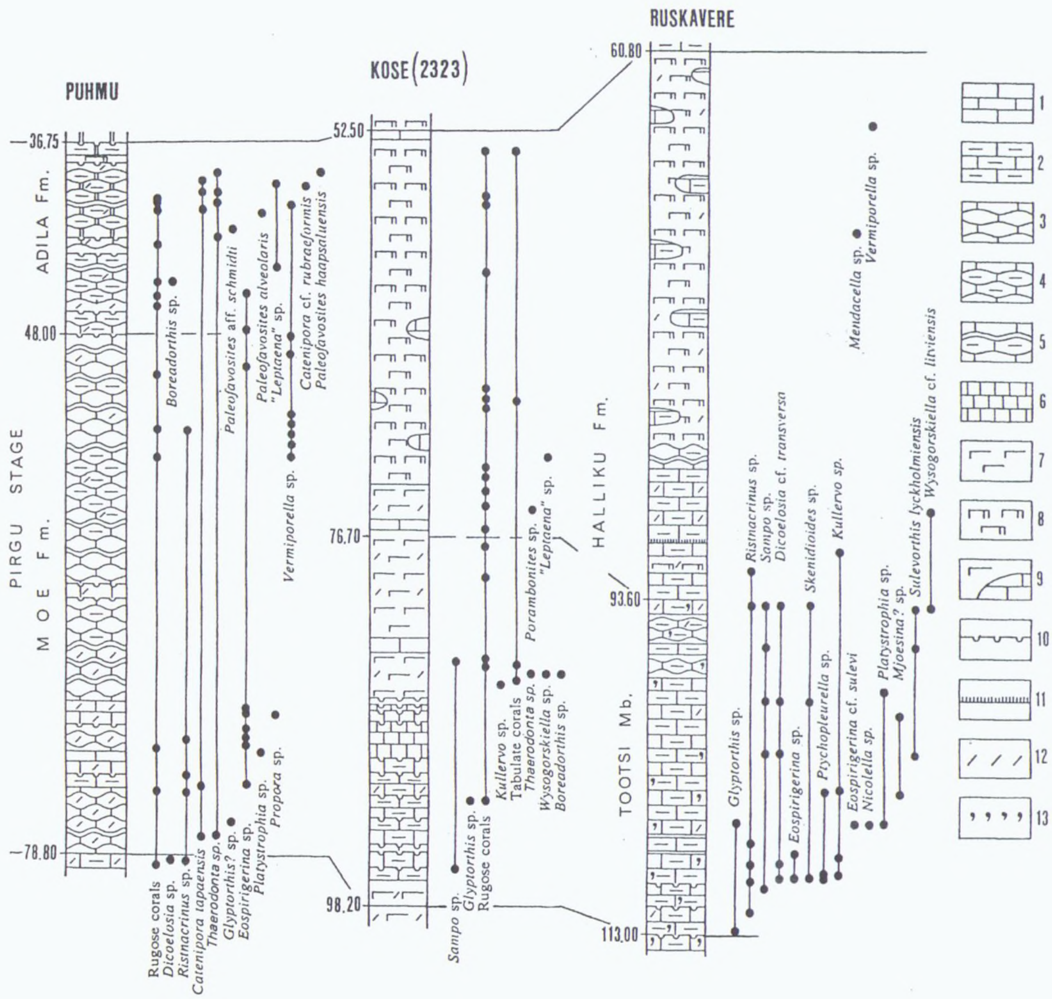


Fig. 59. Correlation of the Pirgu Stage and distribution of macrofossils. Legend: 1 - limestone; 2 - argillaceous limestone; 3 - nodular limestone; 4 - argillaceous nodular limestone; 5 - nodular limestone with clayey intercalations; 6 - micritic (aphanitic) limestone; 7 - marl; 8 - dolomitic marl; 9 - marl with limestone nodules; 10 - discontinuity surface; 11 - K-bentonite; 12 - skeletal detritus; 13 - glauconite.

Stage (Рыымусокс 1989). In many sections the lower boundary of the member is lithologically sharp, except the areas where the topmost part of the Adila Formation is dolomitized. The upper boundary is transitional. Here, the assignment of the Röa Member to the Porkuni Stage is conventional.

Small reef bodies, recorded in the middle part of the Ärina Formation (2-3 m high, up to 20 m wide, traditionally treated as the Tõrevere Member), yield the tabulate corals of the *Palaeofavosites rugosus* community (Клааманн 1986) and the stromatoporoids *Clathrodictyon mammillatum* (Schmidt), *Ecclimadictyon porkuni* (Riabinin) a.o. The reefs are surrounded by skeletal limestones (Vohilaid Member, up to 3.7 m) and kerogenous limestones (Siuge Member, up to 2.6 m; see Ораспыльд 1975б, Rõõmusoks 1983), apparently representing the pre-reef and inter-reef facies, respectively. In the western part of mainland Estonia, the Vohilaid Member, which is often represented by pure skeletal sand in sparry (?) calcite matrix, contains thin (up to 20 cm) layers of oolitic limestone, whereas the ooids make up 10-45% of the rock volume (Ораспыльд 1975б).

In core sections, the common succession of the three lithotypes of the "reef complex" begins with skeletal limestones which are overlain by kerogenous and reef limestones (Fig. 62). The rocks contain a rich and diverse macrofauna of corals, brachiopods, gastropods *etc.*, (more than 150 species and subspecies; Мянниль 1962, Рыымусокс 1970). The associations characteristic of the particular lithotypes have many species in common: rugose corals *Konodophyllum rhizobolon* (Dybowski), *Streptelasma (Streptelasma) giganteum* (Kaljo), brachiopods *Streptis undifera* (Schmidt), *Schmidtomena acuteplicata* (Schmidt), trilobite *Platylichas mastocephalus* (Öpik) and others. Among microfossils, ostracodes are abundant (Meidla 1996), whilst conodonts are extremely rare (Männik 1992b).

South of the distribution area of the Ärina Formation, distinction of the Porkuni Stage is complicated. In some sequences (Ohesaare core) the Porkuni Stage is obviously missing. In many sections in central Estonia (Are and Kahala *etc.*, Fig. 62), the topmost part of the Ordovician sequence is represented by 1- 2m-thick dolomites, which may correspond to some part of the Ärina Formation (?Röa Member). The distribution area of these dolomites coincides roughly with the area of the pre-Silurian (early Silurian?) channeling where the erosion reached the pre-Porkuni rocks (Perens 1995).

In southern Estonia, the Porkuni Stage is represented by the peripheral parts of the Kuldiga and Saldus formations (Ульст и Гайлите 1982). The Kuldiga Formation of bioclastic limestones and marls, overlain by the silty and sandy limestones of the Saldus Formation, comprises the cosmopolitan *Hirnantia* fauna (Rong & Harper 1988). *Hirnantia sagittifera* (M'Coy), *Dalmanella testudinaria* (Dalman),

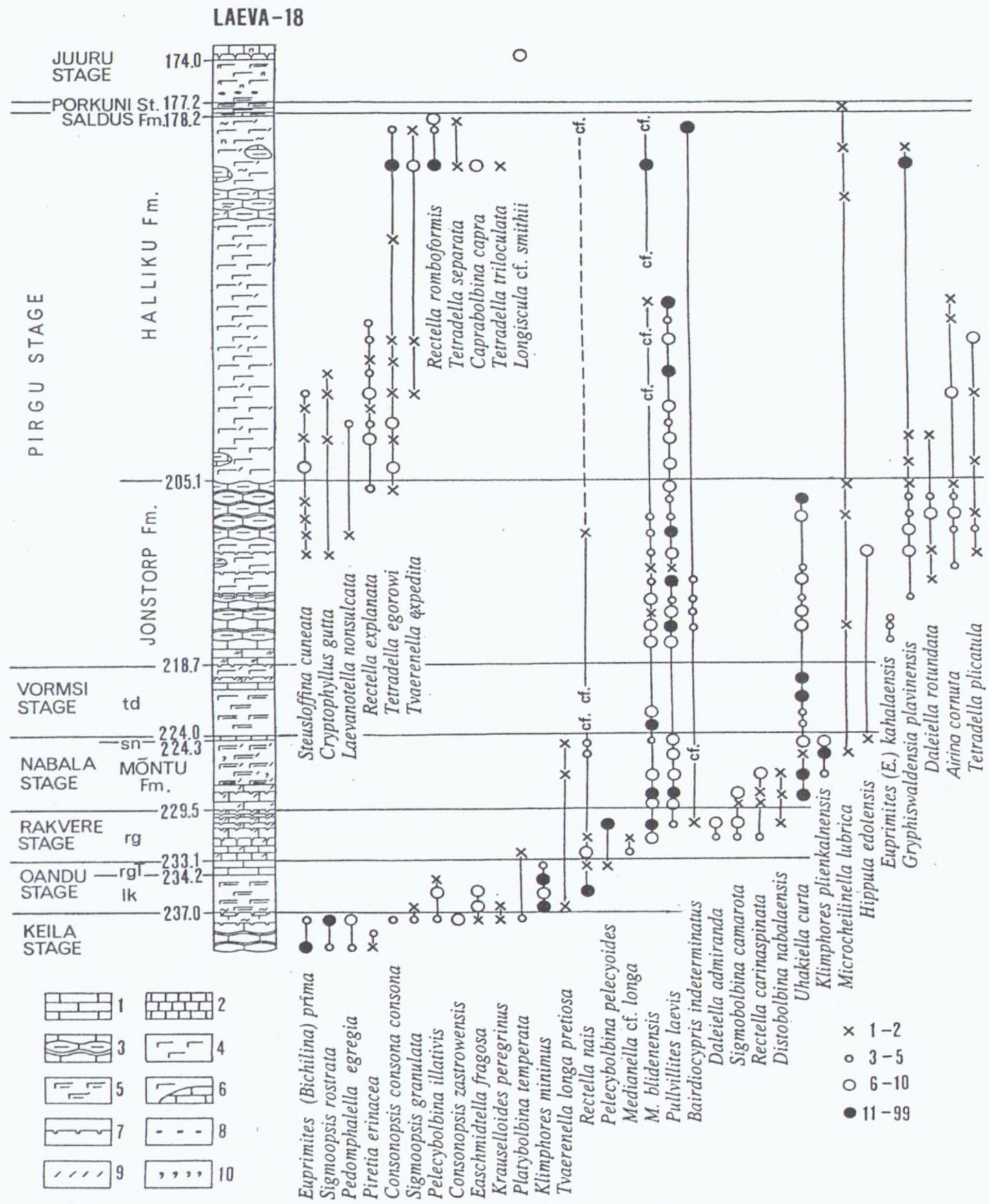


Fig. 60. Distribution of selected ostracodes in the Laeva-18 core section (by Meidla 1996). Legend: 1 - nodular argillaceous limestone; 2 - micritic (aphanitic) limestone; 3 - nodular limestone with clayey intercalations; 4 - marl; 5 - argillaceous marl; 6 - marl with carbonate nodules; 7 - discontinuity surface; 8 - pebbles; 9 - skeletal detritus; 10 - glauconite. The symbols showing the approximate number of specimens are given in the lower right corner. Abbreviations of stratigraphical units: formations – lk - Lukštai, rg - Rägavere, sn - Saunja, td - Tudulinna; rgT - Tõrremägi Member of the Rägavere Formation.

Plectothyrella crassicosta (Dalman), typical elements of that fauna, have been identified in the core sections of Ruhnu, Ikla and Taagepera. These species appear in the lower part of the Kuldiga Formation roughly on the level where the ostracodes common with the Ärina Formation and a zonal chitinozoa *Spinachitina taugourdeaoui* (Eisenack) disappear. The new ostracodes appearing in the Kuldiga Formation seem to have an extraordinarily wide geographical distribution and probably form a part of the Hirnantia fauna *sensu lato* (Meidla 1996).

The youngest Ordovician deposits corresponding to the *Glyptograptus persculptus* graptolite Zone are identified only on the western coast of the East Baltic (Ульст 1992). There is no certain evidence on the occurrence of shallow-water deposits of the *persculptus* Zone in Estonia, although they may be present as the unfossiliferous topmost Ordovician or even in strata assigned to the lowermost Silurian (Kaljo & Hints, in press).

Concluding the data on the terminal Ordovician in Estonia, it should be mentioned that the Porkuni Stage, in the presents limits, comprises rocks of different age. The oldest part of the stage is present in the stratotype area in northern Estonia, while the most complete sequences presumably occur in southern Estonia. The appearance level of the *Hirnantia* fauna, which may be correlated with the lower boundary of the Hirnantia Stage in Scandinavia, lies seemingly in the lower part of the Porkuni Stage in the East Baltic.

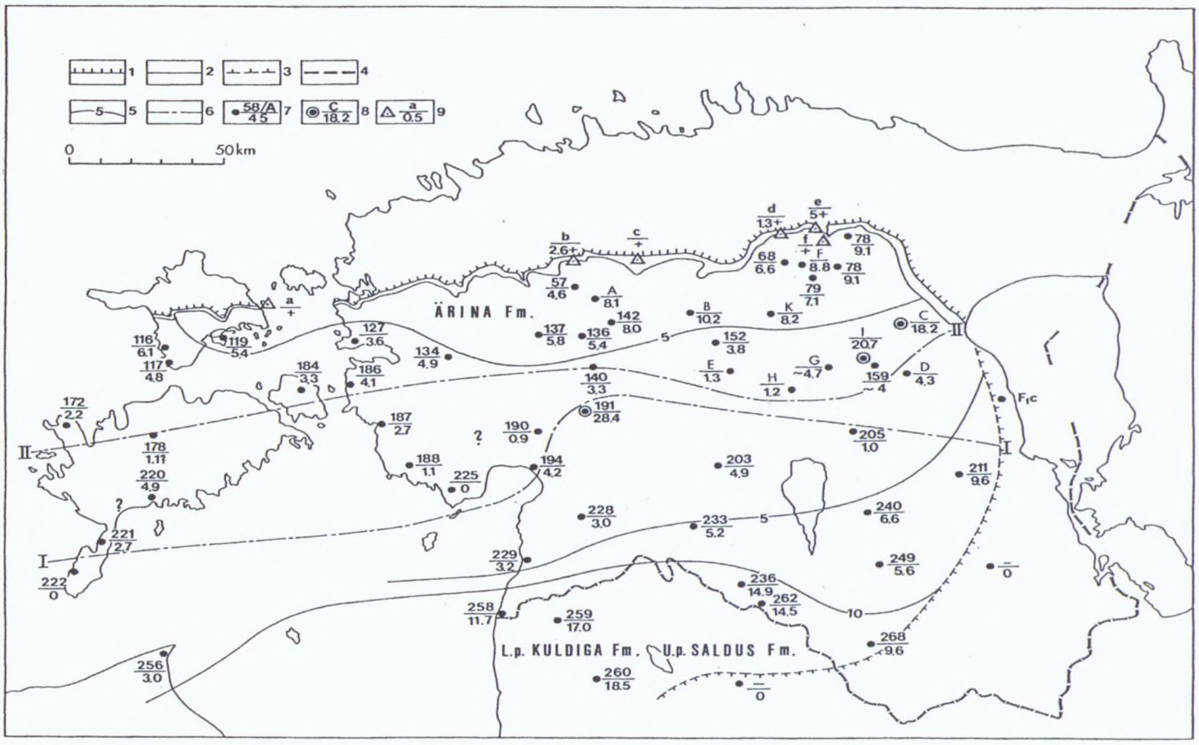


Fig. 61. Sketch-map showing the present distribution and the thickness of the Porkuni Stage. Legend 1 - 7 see Fig. 30; 8 and 9 - core and outcrop with type section, number or index in the numerator, thickness in the denominator. Outcrops: a - Vohilaid, b - Röa, type section of the Röa Member, c - Siuge, type section of the Siuge Member, d - Tõrevere, type section of the Tõrevere Member, e - Porkuni, type section of the Porkuni Stage, f - Ärina, type section of the Ärina Formation; core sections: A - Keava, B - Paide, C - Sadala, D - Palamuse, E - Kabala, F - Raigu, G - Aidu, H - Põltsamaa, I - Jõgeva, K - Kahala. Line I – I - approximate northern boundary of the Saldus Formation (Ораспыльд 1975б), II - II - approximate southern boundary of the reef complex (Ärina Formation without the Röa Member). The signs "L. p." and "U. p." show whether the unit forms the lower or upper part of the stage.

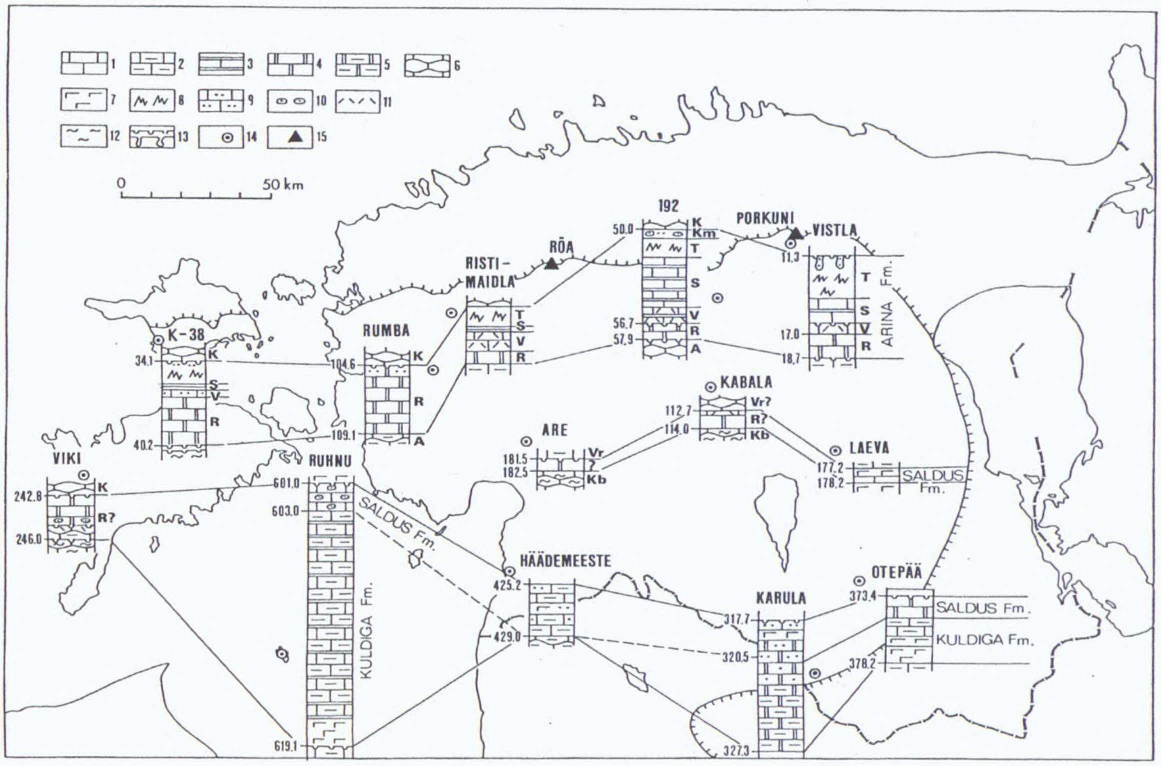


Fig. 62. Correlation and subdivisions of the Porkuni Stage. Letters on the right side of the logs mark formations: A - Adila and Vr - Varbola, and members: Kb - Kabala, R - Röa, V - Vohilaid, S - Siuge, T - Tõrevere, K - Koigi, Km - Kamariku. Legend: 1 - limestone; 2 - argillaceous limestone; 3 - limestone with clayey intercalation; 4 - dolomite; 5 - argillaceous dolomite; 6 - nodular limestone; 7 - marl; 8 - reef; 9 - silty limestone; 10 - carbonate ooids; 11 - skeletal detritus; 12 - burrows; 13 - discontinuity surface; 14 - core section; 15 - outcrop.

The first stratigraphical classification of the Silurian rocks in Estonia was worked out by Schmidt (1858, 1881, 1892). Bekker (1922, 1925) and Luha (1930, 1933, 1946) established the present nomenclature of the Silurian regional chronostratigraphical units - regional stages. Lithostratigraphical divisions have been adequately defined in the monograph "The Silurian of Estonia" (Кальо 1970в). Further amendments to the stratigraphical nomenclature and correlation with the sequences of the adjacent areas have been published in the unified regional stratigraphical charts of the Baltic Republics (Решения... 1978) and of the East-European Platform (Решения... 1987). The latest version of the Silurian stratigraphical chart, approved by the Stratigraphical Commission of Estonia, was published by H. Nestor (1995a) and is followed in the present publication (Table 8).

The Silurian sequence in Estonia consists of ten regional stages grouped directly into the series of the global chronostratigraphical standard. In most cases the boundaries of the regional stages and series have been considered more or less congruent, based on the graptolite or conodont datings (Кальо 1962, Viira 1982). An exception is the Wenlock/Ludlow boundary which is only conventionally fitted with the junction of the Rootsiküla and Paadla stages. The lower limit of the Silurian System coincides with the boundary between the Porkuni and Juuru stages. It is proved by the presence of the Hirnantian trilobites and brachiopods in the Kuldiga and Saldus formations of the Porkuni Stage and records of *Stricklandia lens prima* Williams from the lowermost beds of the Varbola Formation of the Juuru Stage (Kaljo *et al.* 1988b).

Based on the sharply expressed lateral, facies changes of the Silurian rocks, the Mid-Estonian and South-Estonian confacies belts have been distinguished (Кальо 1977). The Mid-Estonian Confacies Belt is dominated by various lime- and dolostones, rich in shelly fauna. The belt covers the islands of the West-Estonian Archipelago and the western and central parts of mainland Estonia (Table 8). In the latter area, the Silurian sequence is less complete; its upper part has undergone severe dolomitization. The South-Estonian Confacies Belt consists mostly of marl- and mudstones with a more unilateral deeper-water shelly fauna, graptolites and planktonic microfossils (chitinozoans). Within the confacies belts separate sets of lithostratigraphical units have been established.

Many parts of the Silurian sequence have a clearly expressed cyclical nature, especially in the more shallow-water Mid-Estonian Confacies Belt. In such cases a cyclostratigraphical unit, the so-called beds consisting of alternating types of rocks with a certain trend of succession, has been distinguished and treated as a subdivision of formation. In some cases formations can be subdivided into members.

LLANDOVERY SERIES

Juuru Stage

The Juuru strata were established by Schmidt (1858) as the Bed ("Jördensche Schicht"), later transferred to the rank of Stage ("Stufe") (Schmidt 1892). Nestor and Kala (Нестор и Кала 1968) determined the present stratigraphical extent of the stage and worked out its classification. With the Juuru Stage they united the Tamsalu Formation, earlier treated as an independent stage, and the lowermost beds of the Raikküla Stage (now the Karinu Member). The former Hilliste Member of the Juuru Stage was recently expanded and raised into formation rank partly corresponding to the Raikküla Stage (Männik 1992b, Nestor 1995a).

The Juuru drill core in the interval of 0.4-16.2 m has been selected as a neostratotype for the Juuru Stage (Nestor 1993). The Juuru Stage spreads on the islands of Hiiumaa and Saaremaa and in the western, central and southern parts of mainland Estonia. The outcrop extends as a west-eastwards widening belt (4 to 25 km) from midsouthern Hiiumaa as far as the eastern slope of the Pandivere Upland. The main localities are ancient coastal cliffs at Kallasto and Pullapää, quarries at Hilleste, Kirimäe, Karinu, Tamsalu and Rakke (Kamariku) and a well in the ancient Varbola stronghold (Fig. 63). The full thickness of the stage varies from 20.1 m in the Asuküla borehole to 63.7 m in the Viljandi borehole (Fig. 63).

The stage is dominated by biomicritic limestones (packstones, wackestones) rhythmically intercalating with thin layers of marl- and mudstones (argillites, clays) and containing interlayers of sparitic limestones (grain- or rudstones). The proportion of marlstones increases southwards and the number of sparitic interlayers towards the north-west and upwards in the sequence.

The lower boundary of the stage coincides with the base of a thin band of micro- to cryptocrystalline limestone of the Koigi Member or, if the latter is absent, with the base of the marl- or mudstones of the Varbola and Õhne formations overlying various sparitic limestones of the Porkuni Stage, including bioclastic and oolitic grainstones, lithoclastic rudstones of shallow-water origin. Above the boundary, the brachiopod *Stricklandia lens*, the chitinozoans *Ancyrochitina laevaensis* and *Spinachitina fragilis* or the conodont *Ozarkodina* ex gr. *oldhamensis* appear.

The Juuru Stage contains a rather rich benthic shelly fauna, whereas planktonic fossils are rare. The most characteristic species are (abbreviations in brackets: *vr* - Varbola Formation, *tm* - Tamsalu Formation, *õh* - Õhne Formation, pt. - part) *Clathrodictyon boreale* Riabinin (*vr, tm*), *Paleofavosites paulus* Sokolov (*vr, tm, õh*), *Stricklandia lens prima* Williams (*vr*, lower pt.), *S. lens lens* Williams (*vr*, upper pt.), *Zygospiraella duboisi* (Verneuil) (*vr*), *Borealis borealis* (Eichwald) (*tm*), *Acernaspis estonica* Männil (*õh*), *Calymene ansensis* Männil (*vr, tm*), *Aitilia senecta* Sarv (*vr*), *Steusloffia eris* Neckaja (*vr, tm, õh*), *Ozarcodina* ex gr. *oldhamensis* (Rexroad) (*vr, tm, õh*), *Distomodus kentuckyensis* (Branson et Mehl) (*vr, tm, õh*), *Ancyrochitina laevaensis* Nestor (*õh*, basal pt.), *Spinachitina fragilis* Nestor *õh*, basal pt.), *Conochitina postrobusta* Nestor (*õh*), *Dimorphograptus confertus* Nicholson) (*õh*, top), *Pribylograptus incommodus* (Toernquist) (*õh*, top). Records of *S. lens prima*, *A. laevaensis* and *S. fragilis* from the basal part of the stage suggest that the base of the Juuru Stage lies on the level of the *Parakidograptus acuminatus* Zone (Cocks 1971, V. Nestor 1994). Graptolites

Table 8. The Silurian of Estonia

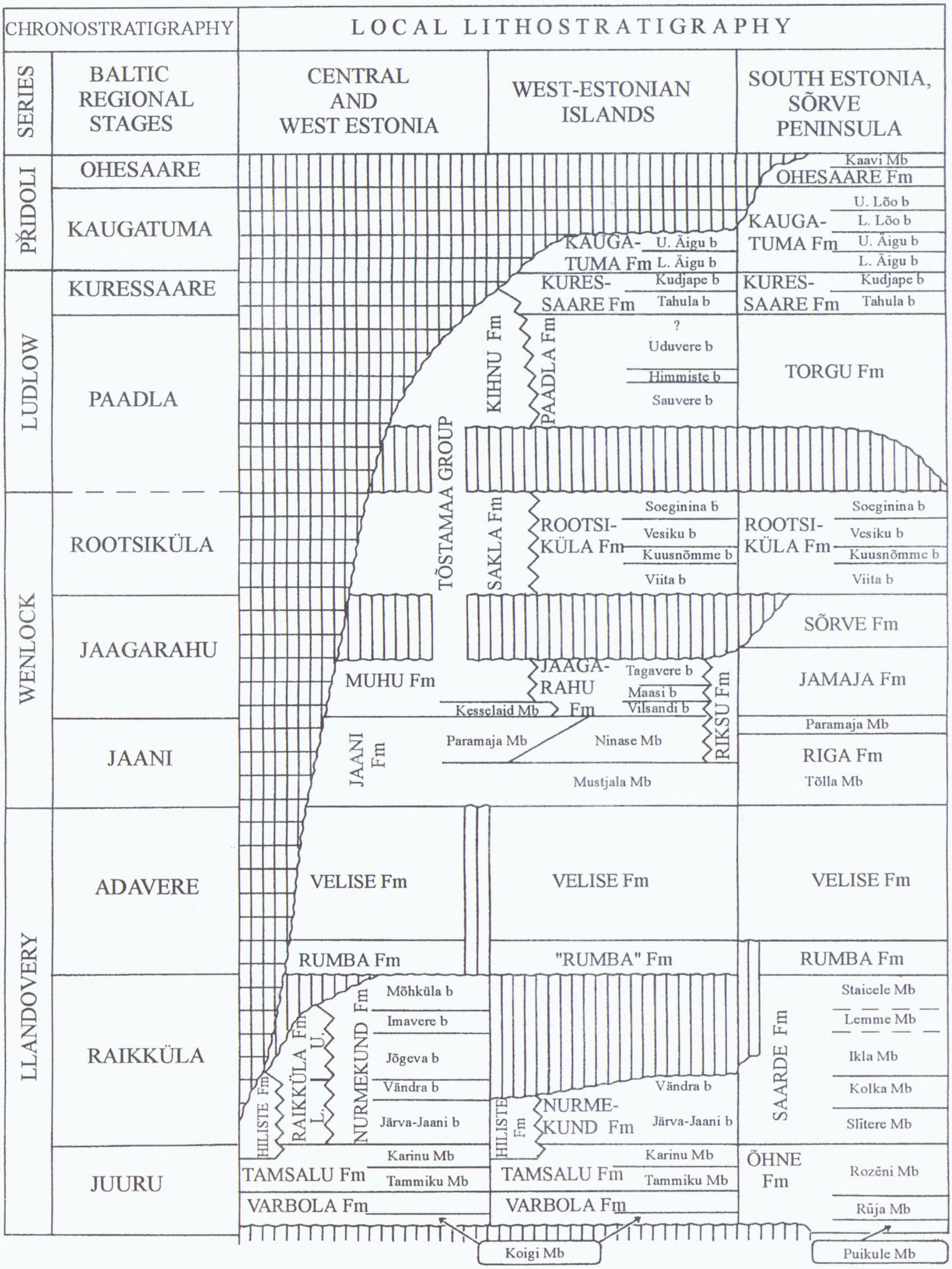


Table 8. The Silurian of Estonia (continued)

BIOZONES					
GRAPTOLITE (Кальо и др. 1984)	CONODONT (Männik, Viira) subzones offshore	CONODONT nearshore	CHITINOZOAN (V. Nestor)	OSTRACODE (Sarv)	VERTEBRATE (Märss)
-	*O. elegans detorta*		*Urochitina*	*N. protuberans*	*K. timanicus*
	O. rem-scheidensis	*O. r. rem-scheidensis*	*E. filifera-E. pistilliformis*	*N. tuberculata*	*Por. punctatus*
ultimus		*O. r. eosteinhorensis*	*A. fragilis*	*F. groenvalliana*	*N. gracilis*
formosus	*O. s. para-snajdri*	*O. aff. scanica*	*S. sphaerocephala* / *B. granosa*	*P. numerosa-U. balticum*	*T. sculptilis (P. ludlowiensis)*
balticus	*O. crispa*		*E. philipi-E. lagenomorpha*	*N. ctenophora-N. lauensis*	*A. hedei*
tauragensis			*B. latifrons*		*Phl. elegans*
scanicus-progenitor	*O. snajdri*	*O. roopaensis*	*G. militaris - C.* sp. 2		*Phl.' ornata*
nilssoni	*K. variabilis*		*Conochitina* sp. 1	*C. ezerensis*	*P. martinssoni*
ludensis	*O. b. bohemica*	*C. murchisoni*	Interzone	*B. cubornata*	
nassa			*S. indecora*		
testis	*K. absidata*		*C. cribrosa* / *C. subcyatha*	*L. quadricuspidata*	*L. einari*
radians / *perneri*	*K. amsdeni*		*C. pachycephala* / *E. lagena*		*L. grossi*
flexilis	*K. walliseri*		*C. cingulata*		
antennularius		*O. s. rhenana*	*C. tuba*		
riccartonensis	*K. ranu-liformis*		*C.* cf. *mamilla*	*C. mucronulata*	
murchisoni			Interzone		
bohemicus			*M. margaritana*		?
spiralis	*P. a. amorphognathoides*		*C. proboscifera*	*L. caudalis-T. walensis*	
griestoniensis	*P. celloni*		*A. longicollis*		
crispus	?	?		?	
turriculatus	*P. eopennatus*		Interzone	*B. valguensis*	*L. scotica*
sedgwickii	?		*C. emmastensis*	?	
convolutus	*D. staurognathoides*		Interzone	?	
gregarius-triangulatus	?		*C.* cf. *protracta*		
	P. tenuis		*A. convexa*	*B. sarvi*	?
cyphus			*C. electa*		
Interzone	*D. kentuckyensis*		*B. postrobusta*	*M. edita-S. eris*	
			A. laevaensis		

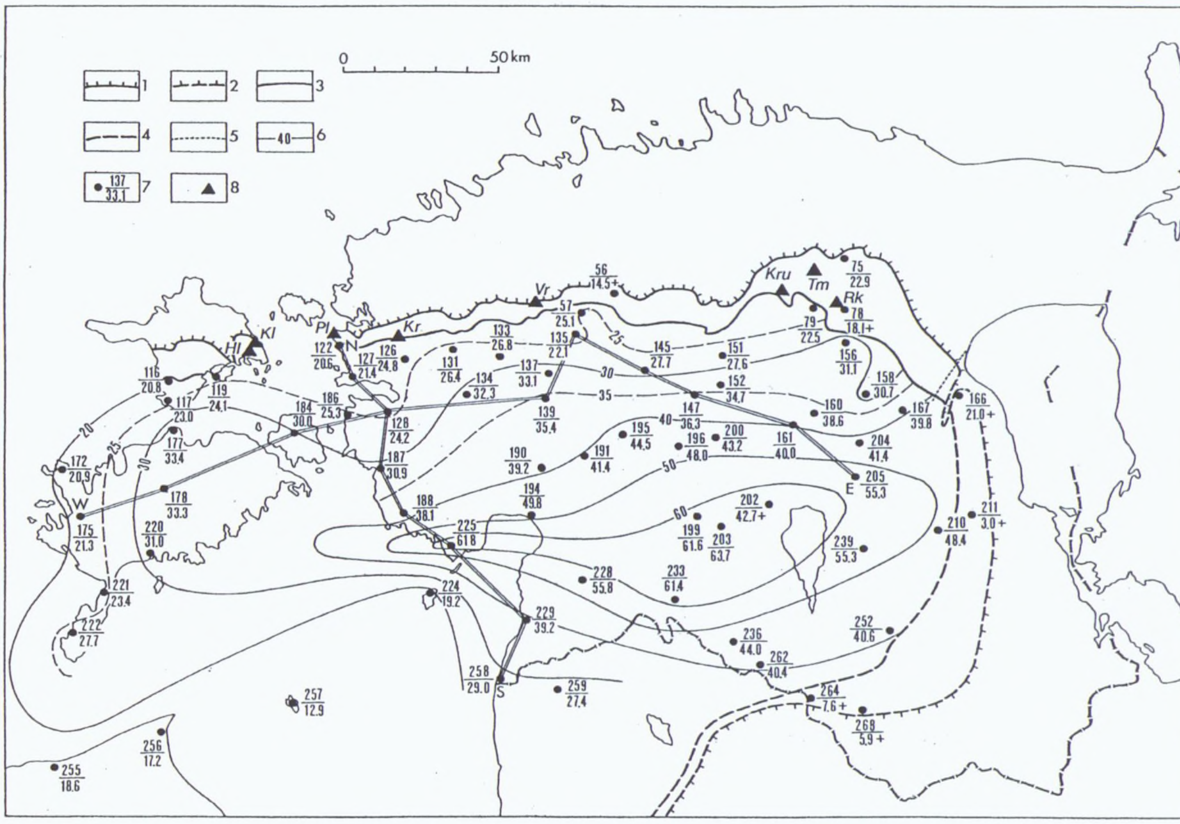


Fig. 63. Sketch-map showing the present distribution and the thickness of the Juuru Stage in Estonia. NS and WE indicate transects of stratigraphical cross-sections presented in Figs. 64 and 65, respectively. Numerator marks the number of the borehole, denominator shows the thickness of the stage in metres. 1 - contour of rock distribution; 2 - contour of subsurface distribution; 3 - limit of outcrop belt; 4 - limit of pre-Devonian erosion; 5 - margin of the Devonian cover; 6 - isopach; 7 - borehole; 8 - exposure locality: Kl - Kallasto, Hl - Hilleste, Pl - Pullapää, Kr - Kirimäe, Vr - Varbola, Kru - Karinu, Tm - Tamsalu, Rk - Rakke.

D. confertus and *P. incommodus* from the top of the stage confirm that the upper boundary of the stage roughly coincides with the boundary between the *Orthograptus vesiculosus* and *Coronograptus cyphus* zones (Кальо и Вингисаар 1969).

In the Mid-Estonian Confacies Belt, the Juuru Stage is divided into the Varbola (below) and Tamsalu (above) formations. In the South-Estonian Confacies Belt, the Õhne Formation corresponds to both of them (Figs. 64, 65).

The **Varbola Formation** is represented by nodular biomicritic limestones (skeletal to coquinoid pack- and wackestones) with thin intercalations of marlstone. The formation contains tempestitic interlayers of skeletal grainstones, often with intraclasts, the number of which increases upwards in the sequence and northwestwards in the space. Brachiopods of the *Stricklandia* Community are characteristic to the formation. The thickness of the formation varies from 8.8 m in the Pusku borehole to 24.6 m in the Käru borehole. The 0.1—3.5-m-thick Koigi Member of micritic (aphanitic) limestones is developed at the base of the Varbola Formation.

The **Tamsalu Formation** consists of various, prevailingly sparitic limestones (skeletal and pelletal grainstones, coquinoid or lithoclastic rud- and floatstones). The thickness of the formation varies from 8.8 m in the Pusku 2 borehole to 18.5 m in the Rumba borehole. The formation is subdivided into the Tammiku (below) and Karinu (above) members.

The Tammiku Member is typically represented by a bank of coquinoid limestone consisting of shells and debris of the brachiopod *Borealis borealis*. The thickness of the bank reaches 13.5 m on the Pandivere Upland. In the same area, the Karinu Member consists of skeletal and pelletal grainstones and bio- or lithoclastic rudstones. South- and westwards the latter are replaced by fine-grained grain- and packstones with numerous hardgrounds.

The **Hilliste Formation** consists of a highly variable assemblage of rock types in which the most characteristic are crinoidal limestones (grainstones) with coral-stromatoporoid bioherms. The formation also contains fine-grained pelletal and skeletal grain- and packstones and micritic limestones. The formation corresponds to the upper part of the Tamsalu Formation (Karinu Member) and to the lower part of the Raikküla Stage (Nestor 1995a). It occurs on Hiiumaa Island and in the vicinity of Haapsalu - Rohuküla and Rapla - Käru in mainland Estonia.

The **Õhne Formation** is represented by marlstones, mudstones and micritic limestones. It corresponds to the whole stratigraphical extent of the Juuru Stage in southern Estonia. The rather poor fauna corresponds to the brachiopod *Clorinda* Community. The maximum thickness (63.7 m) has been fixed in Viljandi 91 borehole. The thin, up-to-2.7-m-thick Puikule Member of marlstones and the overlying, up-to-8-m-thick Ruja Member of micritic limestones occur in the basal part of the Õhne Formation along the southern and eastern margins of the area of distribution of the formation.

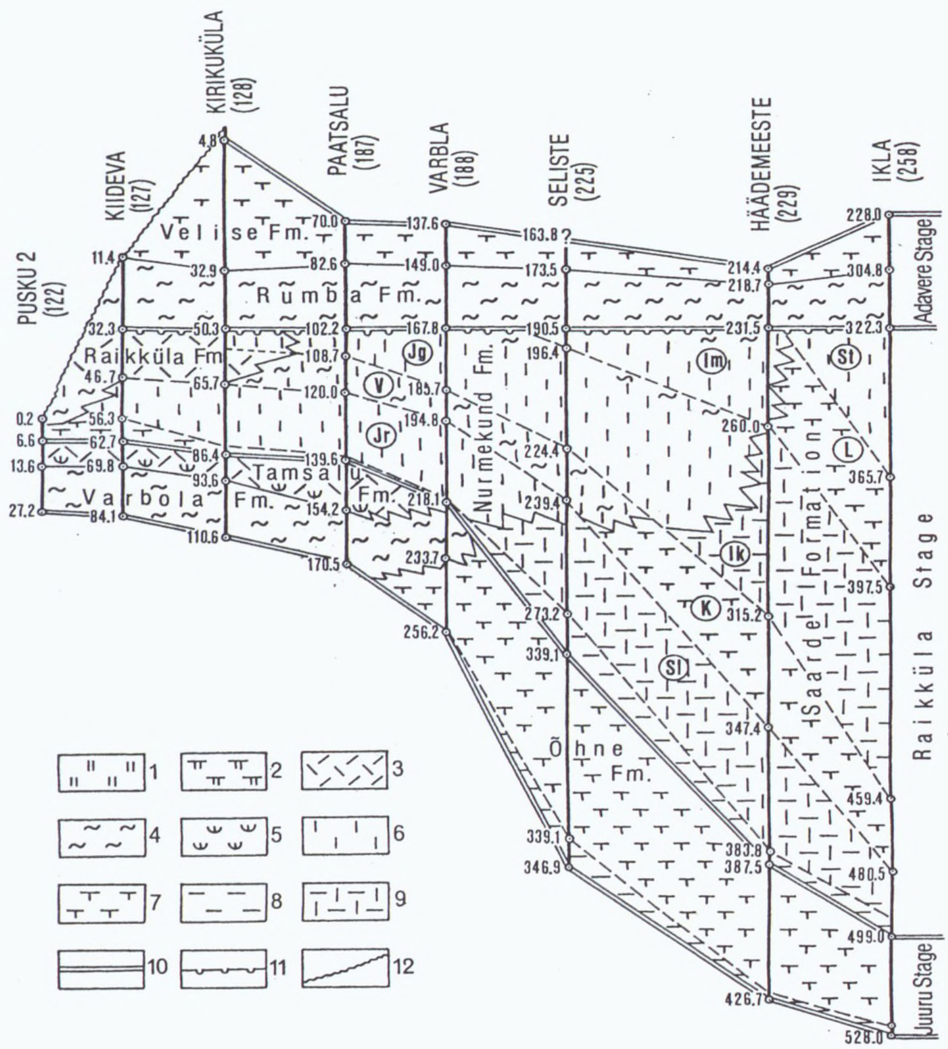


Fig. 64. Meridional (NS) stratigraphical cross-section of the Llandovery: Juuru, Raikküla and Adavere regional stages. For location of cross-section see Fig. 63. Stratigraphical abbreviations of beds: Jr - Järva-Jaani, V - Vändra, Jg - Jõgeva, Im - Imavere; and members: Sl - Slītere, K - Kolka, Ik - Ikla, L - Lemme, St - Staicele.
1 - dolostone; 2 - argillaceous dolostone; 3 - sparitic limestone; 4 - biomicritic limestone; 5 - coquinoid limestone; 6 - micritic limestone; 7 - marlstone; 8 - mudstone; 9 - limestone/mudstone intercalation; 10 - stage boundary; 11 - stratigraphical hiatus; 12 - the present denudation surface.

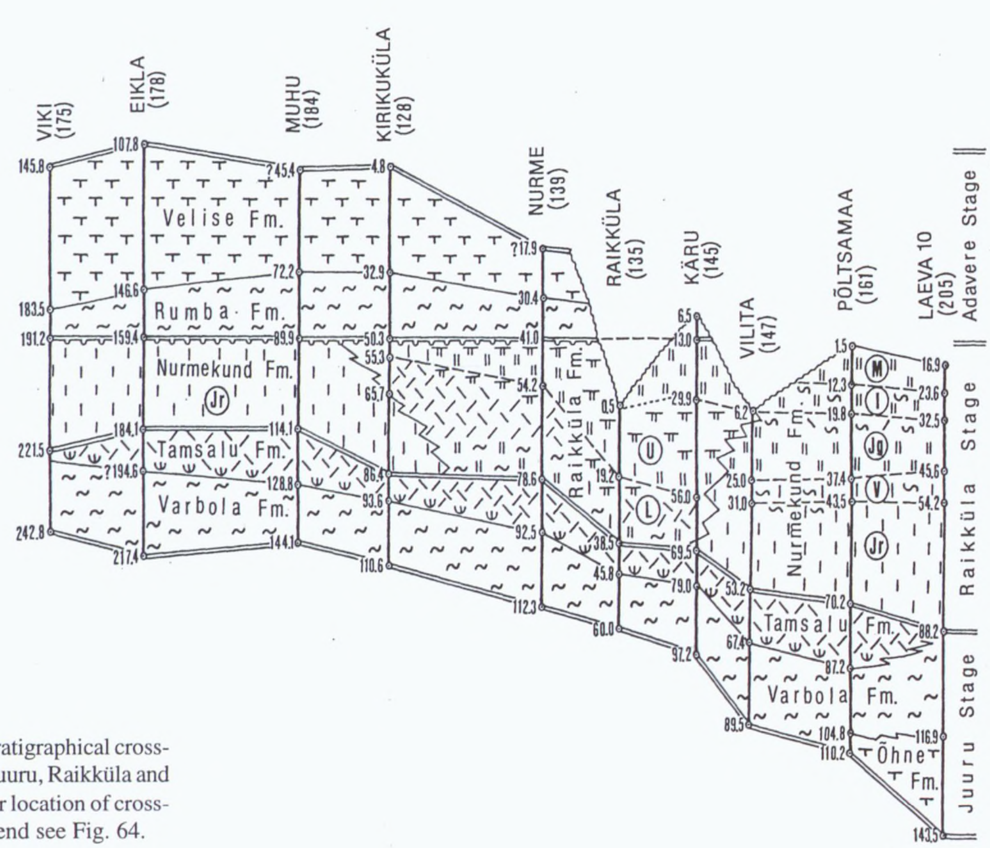


Fig. 65. Latitudinal (WE) stratigraphical cross-section of the Llandovery: Juuru, Raikküla and Adavere regional stages. For location of cross-section see Fig. 63, for legend see Fig. 64.

Raikküla Stage

The Raikküla beds were originally defined (Schmidt 1858) as the "Intermediate zone" (Zwischenzone) between the strata with *Pentamerus borealis* and *P. oblongus*. In 1881, Schmidt introduced the geographical name - Raikküllsche Schicht. Kaljo and Vingisaar (Кальо и Вингисаар 1969) presented the currently used subdivision of the stage for southern Estonia. Perens (1992) and H. Nestor (1995a) modernized the classification for the outcrop area. The Mõhküla beds, earlier attributed to the Adavere Stage, were replaced into the Raikküla Stage as they are separated from the rest of the Adavere Stage by a structural disconformity (Nestor 1995a). However, since the stratigraphical level of the Môhküla beds was changed only recently, it is not yet reflected in the limit between the outcrops of the Raikküla and Adavere stages on the printed geological maps.

The Raikküla-Paka scarp and Raikküla drill core in the interval of 0.5 to 35.0 m have been defined as the composite stratotype of the stage (Nestor 1993). The Raikküla Stage is distributed on the islands of the West-Estonian Archipelago and in the western, central and southern parts of mainland Estonia. The outcrop extends as a latitudinal, eastward widening belt (6 to 45 km) from southern Hiiumaa as far as the southeastern slope of the Pandivere Upland near Palamuse. The main localities are active or abandoned quarries at Pusku, Orgita, Keava, Mündi, Kalana and Rôstla, and ancient coastal scarps (inland cliffs) at Pakamägi and Raikküla-Paka. The thickness of the stage varies from 16.3 m in the Murika borehole to 176.3 m in the Ikla borehole (Fig. 66) and decreases abruptly in the northwest direction due to the end-Raikküla denudation of the upper layers of rocks.

The Raikküla Stage consists of a variety of carbonate rocks. The most characteristic are micritic (micro- and cryptocrystalline) limestones cyclically interbedding with marl- or mudstones in the south and with different bioclastic limestones (wacke-, pack- and grainstones) in the north. In the northernmost sections of central Estonia, the shallowing-up sedimentary cycles may end with argillaceous primary dolomites. In the southernmost sections, the marl- and mudstones contain graptolites on certain levels. In central Estonia, in the Paide - Pärnu belt of faults and eastwards, the Raikküla rocks are strongly dolomitized.

The lower boundary of the stage coincides with the base of a band of marl- or mudstones overlain by thick deposit of monotonous micritic limestones of the Järva-Jaani beds in the north and Slītere Member in the south. In the area of distribution of the most shallow-water sequences of the Hilliste and Raikküla formations the boundary is less definite. Above the boundary, sparse graptolites of the *Pristiograptus cyphus* Biozone and chitinozoans of the *Conochitina electa* Biozone (*C. electa*, *C. maennili, etc.*) appear.

Fossils are of uneven distribution in the rocks of the Raikküla Stage. The widespread micritic limestones and different dolostones (from pure dolomite to dolomitic marl) contain occasional macrofossils. The most characteristic species are (abbreviations in brackets: *rk* - Raikküla Formation, *nr*-

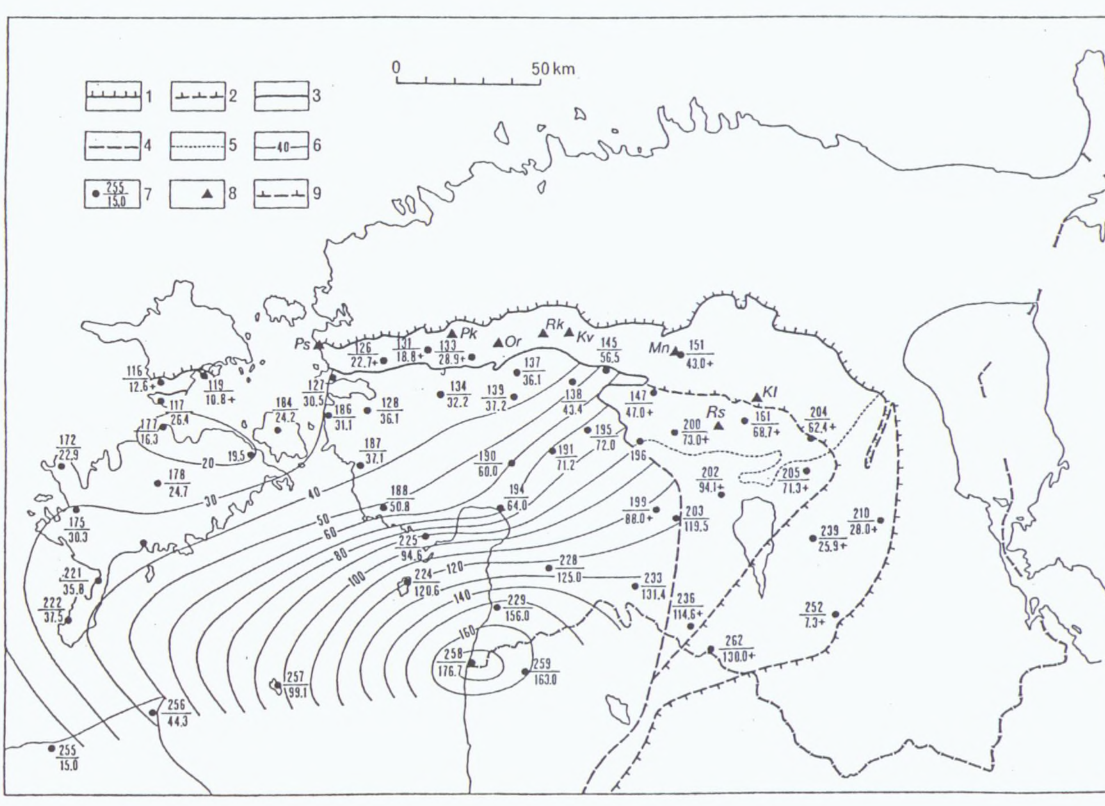


Fig. 66. Sketch-map showing the present distribution and the thickness of the Raikküla Stage in Estonia. Numerator marks the number of the borehole, denominator shows the thickness of the stage in metres. 1 - contour of rock distribution; 2 - contour of subsurface distribution 3 - limit of outcrop belt; 4 - limit of pre-Devonian erosion; 5 - margin of the Devonian cover; 6 - isopach; 7 - borehole; 8 - exposure locality; 9 - extension limit of the Mõhküla beds. Localities: Ps - Pusku, Pk - Pakamägi, Or - Orgita, Rk - Raikküla-Paka, Kv - Keava, Mn - Mündi, Kl - Kalana, Rs - Röstla.

Nurmekund Formation, *sr* - Saarde Formation, u.pt. - upper part, m.pt. - middle part, l.pt. - lower part): *Clathrodictyon clivosum* Nestor (*rk*, u. pt.), *Parastriatopora celebrata* Klaamann (*rk*, u. pt.), *Borealis pumilus* (Eichwald) (*nr*), *Borealis borealis osloensis* Mjork (*nr*), *Meifodia ovalis* Williams (*sr*), *Hermannina hisingeri* (Schmidt) (*rk, nr*), *Bythrocyproidea sarvi* Neckaja (*nr*), *Icriognathus cornutus* Männik (*rk*, l. pt.), *Kockelella manitoulinensis* (Pollack, Rexroad et Mehl), (*rk, nr*), *Conochitina electa* Nestor (*rk, nr, sr*, l. pt.), *C. iklaensis* Nestor (*nr, sr*), *Spinachitina maennili* Nestor (*sr*), *Coronograptus cyphus* (Lapworth) (*sr*, l. pt.), *C. gregarius* (Lapworth) (*nr, sr*, m. pt.), *Demirastrites triangulatus* (Harkness) (*sr*, m. pt.), *D. convolutus* (Hisinger) (*sr*, u. pt.). The presence of zonal graptolites shows that the Raikküla Stage spans from the *C. cyphus* Biozone to the *D. convolutus* Biozone.

The Raikküla Stage consists of the Raikküla, Nurmekund and Saarde formations, laterally replacing each other from north to south (Figs. 64, 65). The upper part of the Hilliste Formation is of Raikküla Age (Table 8).

The **Raikküla Formation** is distributed in central and western Estonia, in the Lääne, Rapla and Järva counties. It is represented by two shallowing-up sedimentation cycles starting with biomicritic or micritic limestones, succeeded by skeletal grainstones, pelletal or coral-stromatoporoid limestones, and ending with argillaceous lagoonal dolostones. These cycles are treated as the lower and upper subformations (Nestor 1995a). The thickness of the formation varies from 30 m in the Kiideva borehole to 56 m in the Käru borehole. The upper layers of the formation have undergone considerable denudation and in the westernmost sections of mainland Estonia the upper subformation thins totally out.

The **Nurmekund Formation** south and east of the Raikküla Formation consists of five sedimentary cycles which begin with a relatively thin layer of marlstone or argillaceous limestone. The main, middle part of the cycle is represented by wavy-bedded micritic limestone, the upper part by bioclastic limestones containing numerous discontinuity surfaces. In central Estonia, the formation is strongly dolomitized, particularly its upper half. The first, third and fifth cycles from below are thicker and more complete, the second and fourth being thinner and less typical. In ascending order, the cycles are termed the Järva-Jaani, Vändra, Jõgeva, Imavere and Mõhküla beds (Table 8). Westwards the upper beds gradually thin out and on Saaremaa Island only the Järva-Jaani and, partly, the Vändra beds are present. The thickness of the formation ranges from 16 m in the Murika borehole to 73+ m in the Võhma borehole.

The **Saarde Formation** is distributed in southwestern Estonia. It consists of cyclically alternating deposits of horizontally-bedded micritic lime- and marlstones or mudstones and is subdivided into six members. The lowermost, rather thin mudstone member, comprised mostly of argillites, has no name and was earlier included in the Õhne Formation of the Juuru Stage. In ascending order, the Slītere, Kolka, Ikla, Lemme and Staicele members follow. The shaly mudstone interlayers in the Ikla Member abound in graptolites of the *Demirastrites triangulatus* Zone. In other members graptolites are less frequent. In its full thickness (176.3 m) the Saarde Formation occurs only in the Ikla borehole.

Adavere Stage

The Adavere Stage as a stratigraphical unit was established by Schmidt (1858) as the uppermost unit (zone 6) of the group of smooth pentamerids ("Gruppe der glatten Pentameren"). Afterwards it was termed the Esthonus-Schicht (Schmidt 1881), Addifer Formation (Twenhofel 1916), Adavere Stage (Bekker 1922). Kaljo (Кальо 1962) fitted the upper boundary of the stage with the Llandovery and Wenlock boundary and included in it the marlstones of the present Velise Formation. Recently, Perens (1992) and Nestor (1995a) excluded the Mõhküla beds and replaced them into the Raikküla Stage.

The Päri quarry in western Estonia has been selected as the neostratotype of the stage (Nestor 1993, 1995a) and the Kirikuküla core at the depth of 50.3 m may be treated as the boundary stratotype of the stage. The Adavere Stage is distributed in the southernmost part of Hiiumaa Island, on Saaremaa and Muhu islands and in the southwestern part of mainland Estonia as far as the Viljandi fault. The outcrop extends as a 10—15-km-wide belt from the southernmost Hiiumaa Island and the Soela Strait over Matsalu Bay up to the vicinity of Türi - Vändra being denudated eastwards the Paide-Pärnu belt of disturbances. The main localities are the Saastna coast, Päri quarry, river banks at Päärdu, Jädivere, Velise, Valgu, Vändra and ditches at Lätiküla and Valgu (Fig. 67). The thickness of the stage increases westwards — from 10.7 m in the Ristiküla borehole to 56.3 m at Nässumaa.

The Adavere Stage is represented by thin-bedded to nodular biomicritic limestones (wackestones to packstones) with *Pentamerus oblongus* (below) and marl- to mudstones (above). The former unit is treated as the Rumba Formation and the latter as the Velise Formation. The clay content increases westwards. The lower boundary of the stage coincides with the strongly pyritized erosion surface at the base of the nodular biomicritic limestones of the Rumba Formation, transgressively overlying different strata of the Raikküla Stage. The Adavere Stage contains rather rich shelly fauna of *Pentamerus* (below) and *Clorinda* (above) communities. Microfossils (chitinozoans, ostracodes, conodonts) are more frequent in the mud- and marlstones of the Velise Formation, almost devoid of corals and stromatoporoids. The most characteristic species are as follows (abbreviations in brackets: *rm* - Rumba Formation, *vl* - Velise Formation): *Clathrodictyon variolare* (Rosen) (*rm*), *Mesofavosites obliquus* Sokolov (*rm*), *Angopora hisingeri* (Jones) (*vl*), *Palaeocyclus porpita* (Linnaeus) (*vl*), *Prodarwinia speciosa* (Dybowski) (*rm*), *Pentamerus oblongus* (Sowerby) (*rm*), *Stricklandia laevis* (Sowerby) (*rm*), *Dicoelosia baltica* Musteikis *et* Puura (*vl*), *Encrinurus (Nucleurus) rumbaensis* Rosenstein (*rm*), *Calymene frontosa* Lindström (*vl*), *Beirichia valguensis* Sarv (*rm*), *Longiscella caudalis* (Jones) (*vl*), *Conochitina emmastensis* Nestor (*rm*), *Eisenackitina dolioliformis* Umnova (*rm, vl*), *Angochitina longicollis* Eisenack (*vl*), *Pterospathodus celloni* Walliser (*vl*), *P. amorphognathoides* Walliser (vl), *Spirograptus turriculatus* (Barrande) (*vl*), *Monograptus discus* Törnquist (*vl*), *Monoclimacis griestoniensis* (Nicol) (*vl*).

The presence of the index species of graptolites (*S. turriculatus*, *M. griestoniensis*) and conodonts (*P. celloni, P. amorphognathoides*) in the upper half of the Adavere Stage demonstrates that most probably the stage corresponds to the *Monograptus sedgwickii* to *Monoclimacis crenulata* biozones.

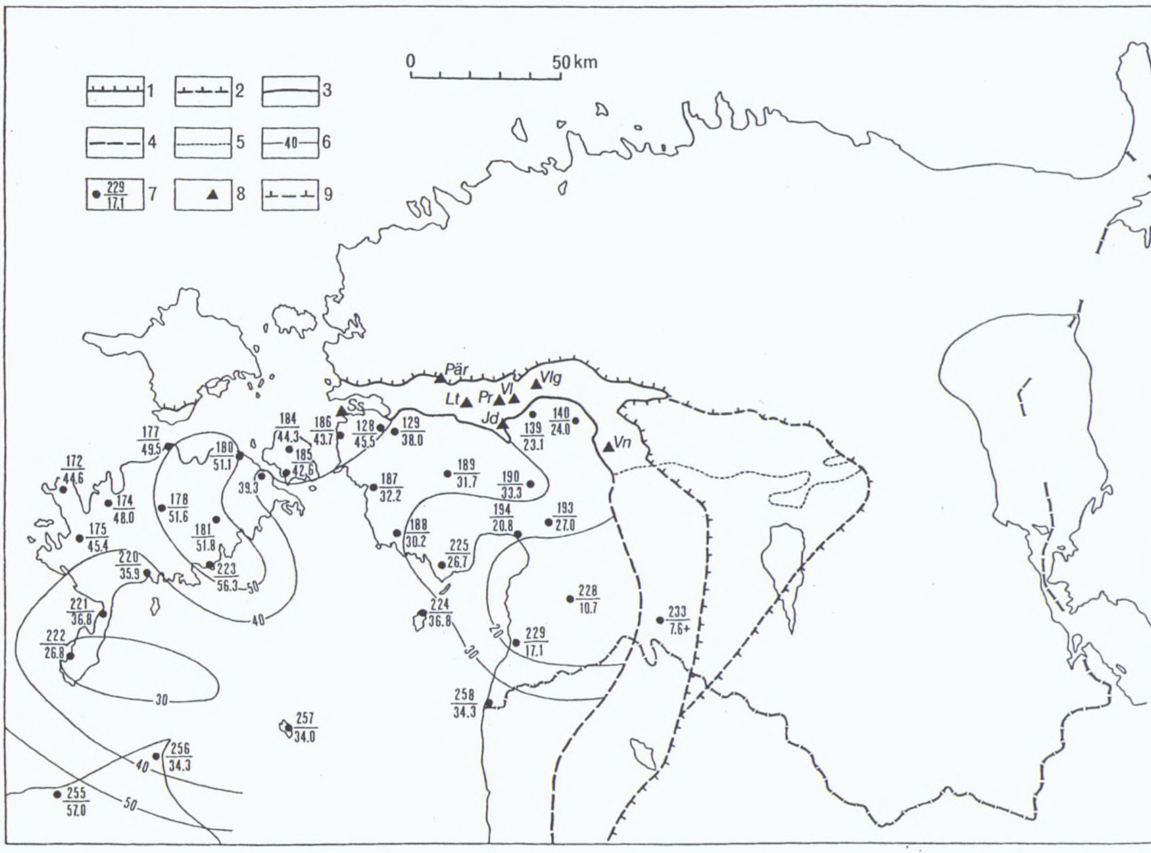


Fig. 67. Sketch-map showing the present distribution and the thickness of the Adavere Stage in Estonia. Numerator marks the number of the borehole, denominator shows the thickness of the stage in metres. 1 - contour of rock distribution; 2 - contour of subsurface distribution; 3 - limit of outcrop belt; 4 - limit of pre-Devonian erosion; 5 - margin of the Devonian cover; 6 - isopach; 7 - borehole; 8 - exposure locality; 9 - extension limit of the Mõhküla beds. Localities: Ss - Saastna, Pär - Päri, Lt - Lätiküla, Pr - Päärdu, Vl - Velise, Vlg - Valgu, Vn - Vändra.

In Estonia, the Adavere Stage consists of the Rumba (below) and Velise (above) formations. The **Rumba Formation** spreads on the islands of the West-Estonian Archipelago and in the southwestern part of mainland Estonia. It is represented by horizontally-bedded to nodular biomicritic limestones (wackestones, packstones) with clayey partings and scattered shells or tempestitic accumulations of the brachiopod *Pentamerus oblongus*. The formation consists of twelve low-grade sedimentary cycles beginning with argillaceous rocks (marlstones, argillaceous limestones) and ending with a layer of pure, hard limestone (Эйнасто и др. 1972). Westwards the clay content of the rocks increases and on Saaremaa Island marlstones are prevailing in the sequence of the Rumba Formation. A characteristic yellowish-green tuffaceous (metabentonite) interlayer (8 to 18 cm) occurs at the level of the base of the upper third of the sequence.

The thickness of the Rumba Formation is mostly 15 to 19 m and it decreases at the western and eastern margins of the distribution area. Local hiatuses occur in the Ohesaare and Are sections.

The **Velise Formation** overlies the Rumba Formation and consists of different marlstones and mudstones up to plastic clays. The mostly greenish- to bluish-grey rocks are south- and eastwards replaced by red-coloured (purple) varieties. In the southwesternmost sections (Ohesaare, Ruhnu) graptolites are present in the dark-grey interlayers of argillite. Thin (0.5 to 5.0 cm) metabentonite interlayers are characteristic to the formation. The thickness of the formation is greatest in northwestern Saaremaa, reaching 37 - 38 m in the Viki and Eikla sections. In the southeast direction, it decreases until thinning out in the Ristiküla section, eastern Pärnumaa.

WENLOCK SERIES

Jaani Stage

The Jaani Stage was defined by Luha (1933) as a marlstone unit corresponding to the lower part of the "Untere Oeselsche Gruppe (Stufe)" by Schmidt (1858, 1892). Kaljo (Кальо 1962) separated the lower part of the marlstones (now the Velise Formation), corresponding to the uppermost Llandovery, and joined it with the Adavere Stage. V. Nestor (Нестор В. 1984) determined the position of the upper boundary in the subsurface area. Aaloe (Аалоэ 1960, 1961) subdivided the Jaani Stage into the Mustjala, Ninase and Paramaja members. Later, Aaloe and Kaljo (Аалоэ и Кальо 1962) distinguished the Tõlla Member for the South-Estonian subsurface area.

A historical stratotype of the Jaani Stage is the sea shore with the Paramaja Cliff in the vicinity of the Jaani Church (Решения... 1987, Nestor 1993). The Ohesaare drill core at the depth of 345.8 m may be treated as the boundary stratotype of the stage. The Jaani Stage spreads on Saaremaa and Muhu islands and in the southwestern part of mainland Estonia (Pärnumaa and southern Läänemaa). The outcrop runs along

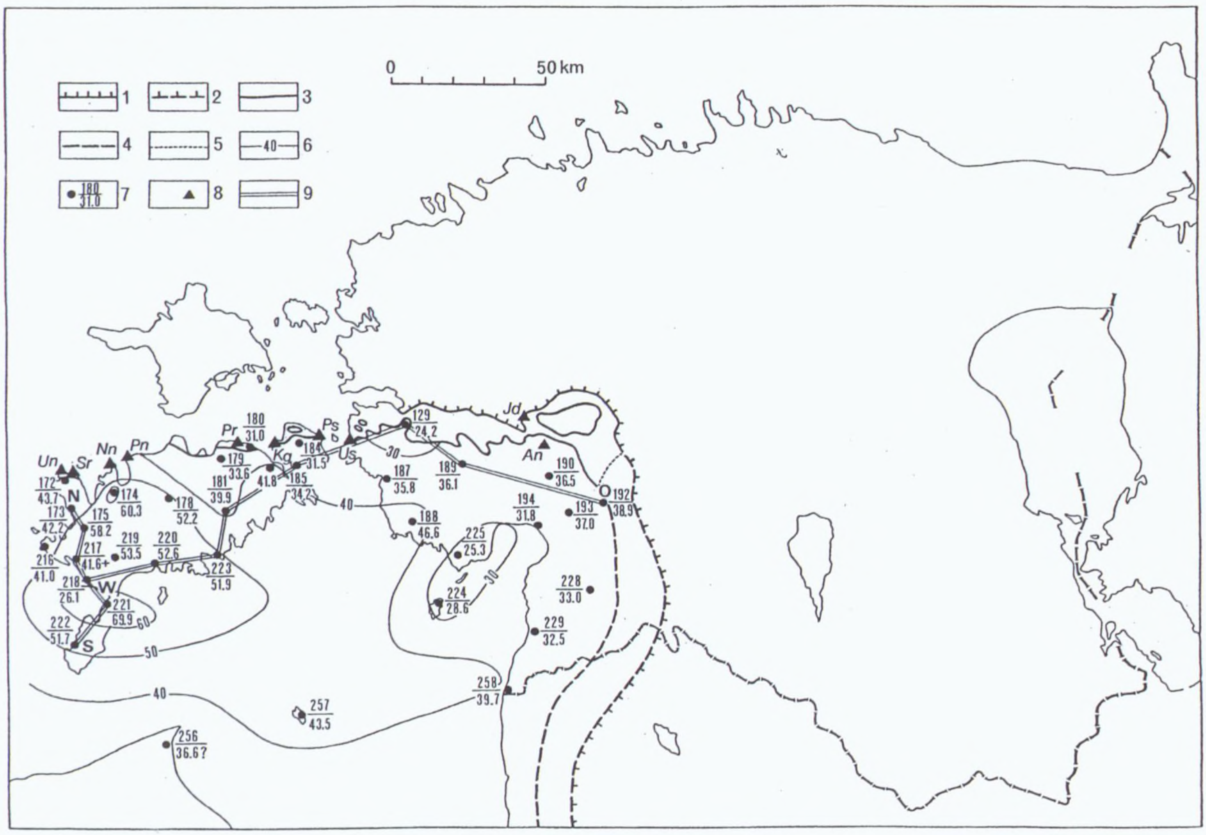


Fig. 68. Sketch-map showing the present distribution and the thickness of the Jaani Stage in Estonia. NS and WO indicate transects of stratigraphical cross-sections presented respectively in Figs. 69 and 70. Numerator marks the number of the borehole, denominator shows the thickness of the stage in metres. 1 - contour of rock distribution; 2 - contour of subsurface distribution; 3 - limit of outcrop belt; 4 - limit of pre-Devonian erosion; 5 - margin of the Devonian cover; 6 - isopach; 7 - borehole; 8 - exposure locality: Un - Undva, Sr - Suuriku, Nn - Ninase, Pn - Panga, Pr - Paramaja, Kg - Koguva, Ps - Püssina, Us - Uisu, Jd - Jädivere, An - Anelema, 9 - transect of cross-section.

the northern coast of Saaremaa and Muhu islands and the southern coast of Matsalu Bay towards the Vändra Borough (Fig. 68). The main localities are the cliffs at Undva, Suuriku, Ninase, Panga, Liiva, Paramaja, Kautliku, Püssina and Uisu, the quarries at Koguva and Anelema (lower part), and the river bank at Jädivere. The thickness of the stage increases westwards and varies from 24.2 m in the Lihula borehole to 70 m in the Kaugatuma borehole.

The stage consists mainly of various marl- and mudstones. Limestones (skeletal wacke-, pack-, grain- and boundstones) are of minor importance and occur only in the upper half of the stage in the northwestern part of Saaremaa Island (Ninase Member). The lower boundary of the stage has been made congruous with the Llandovery/Wenlock boundary (Кальо 1962), established by the appearance of the graptolite *Cyrtograptus murchisoni* in the Ohesaare drill core above the depth level 345.8 m and in other sections by chitinozoans of the *Margachitina margaritana* Zone (Nestor, V. 1994). Lithologically, it usually coincides with a certain increase in the carbonate content of rocks.

The Jaani Stage contains rather rich shelly and planktonic faunas with characteristic deeper-water elements (graptolites, chitinozoans, trilobites). The most typical species are as follows (abbreviations: *M* - Mustjala Member, *P* - Paramaja Member, *N* - Ninase Member, *T* - Tõlla Member, u.pt. - upper part, l.pt. - lower part): *Stromatopora impexa* Nestor (*M*, u. pt.), *Halysites senior* Klaamann (*M*), *Thecia podolica* (*P*), *Neocystiphyllum keyserlingi* (Dybowski) (*P*), *Leptaena rhomboidalis* (Wahlenberg) (*M,P*), *Eocoelia angelini* (Lindström) (*N*), *Pseudobollia krekenawaiensis* Neckaja (*T,P*), *Craspedobolbina (C.) mucronulata* Martinsson (*N,P*), *Beirichia (B.) suurikuensis* Sarv (*N,P*), *Calymene orthomarginata* Schrank (*T,P*), *Encrinurus punctatus* (Wahlenberg) (*P*), *Conochitina* cf. *mamilla* Laufeld (*T,N,P*), *Calpichitina acollaris* (Eisenack) (*P*), *Pterospathodus amorphognathoides* Walliser (*T,M*, l.pt.), *Kockelella ranuliformis* (Walliser) (*M,N,P*), *Cyrtograptus murchisoni* Carruthers (*T*), *Monograptus riccartonensis* Lapworth (*T*), *M. flexilis* Elles (*P*). The presence of the index species of graptolites proves that the Jaani Stage spans from *the C. murchisoni* Biozone to the *M. flexilis* Biozone.

In Estonia, the Jaani Stage is mainly represented by the Jaani Formation. Only in the southernmost sections, the lower part of the stage has been treated as the Tõlla Member of the Riga Formation (Figs. 69,70).

The **Jaani Formation** consists of marlstones and, to a lesser extent, of bioclastic and biohermal limestones. The lower part is formed by the Mustjala Member comprising argillaceous marlstones (Figs. 69,70), which are often dolomitized, particularly in eastern sections. In the middle of the sequence of the Jaani Stage, the carbonate content increases abruptly and, respectively, the upper half of the Jaani Formation is represented by calcareous marlstones or argillaceous limestones of the Paramaja Member in the eastern part of

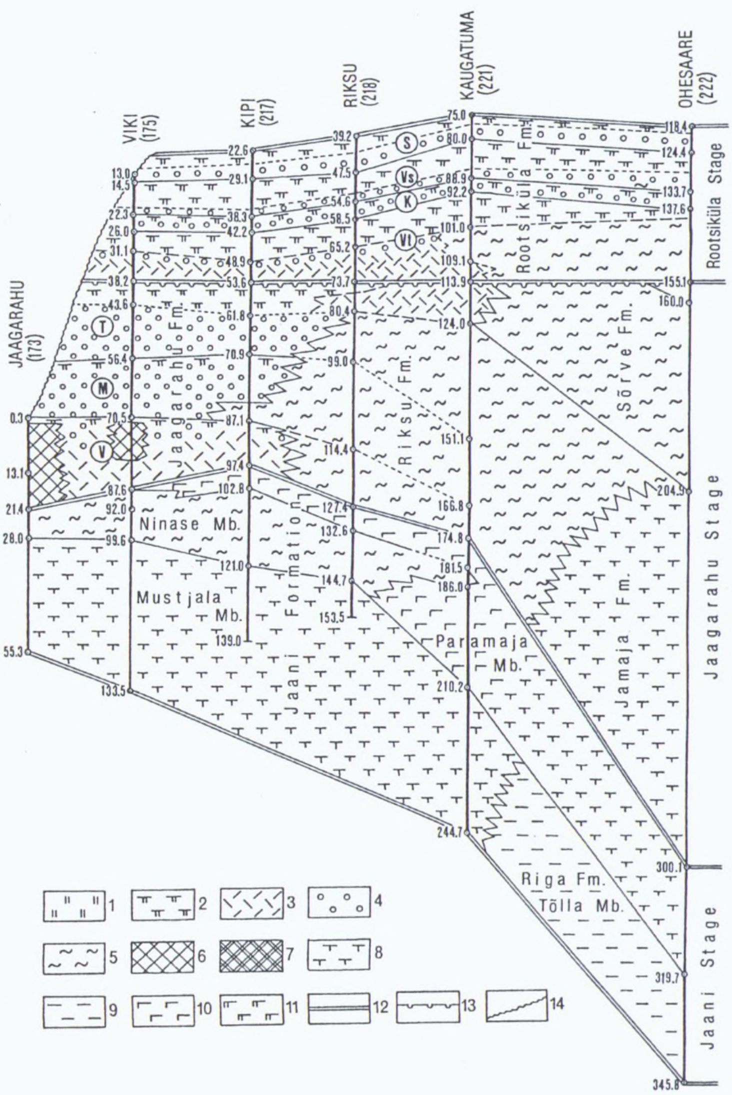


Fig. 69. Meridional (NS) stratigraphical cross-section of Wenlock: Jaani, Jaagarahu and Rootsiküla regional stages. For location of cross-section see Fig. 68. Abbreviations of beds: V - Vilsandi, M - Maasi, T - Tagavere, Vt - Viita, K - Kuusnõmme, Vs - Vesiku, S - Soeginina. 1 - dolostone; 2 - argillaceous dolostone; 3 - skeletal grainstone; 4 - pelletal-skeletal limestone; 5 - biomicritic limestone; 6 - reef limestone; 7 - reef dolomite; 8 - argillaceous marlstone; 9 - mudstone; 10 - argillaceous limestone, calcareous marl; 11 - dolomitic marl; 12 - stage boundary; 13 - stratigraphical hiatus; 14 - the present denudation surface.

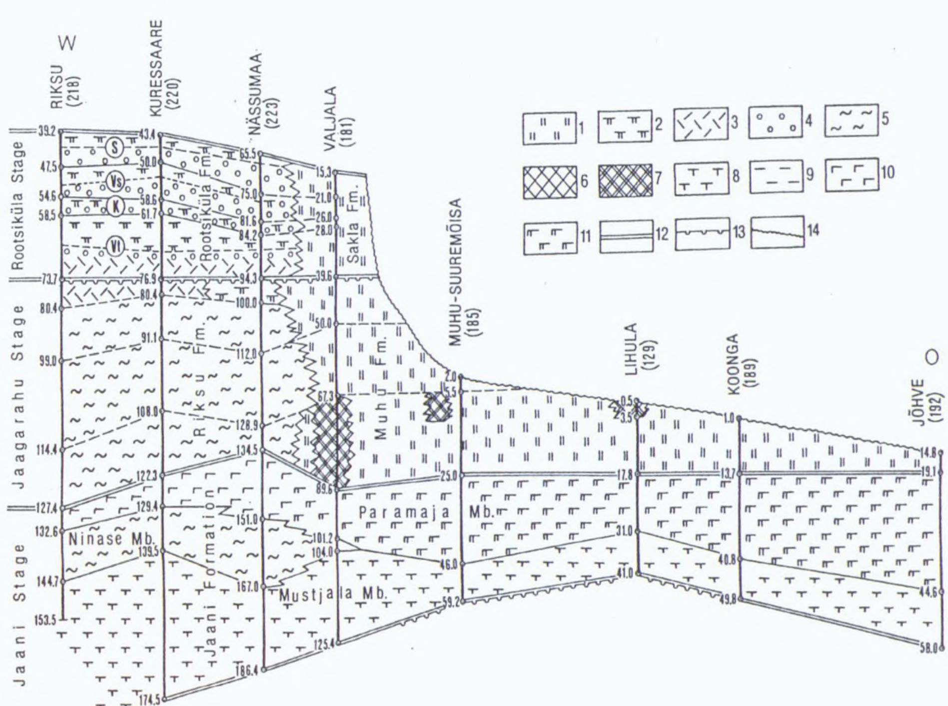


Fig. 70. Latitudinal (WE) stratigraphical cross-section of the Wenlock Jaani, Jaagarahu and Rootsiküla regional stages. For location of cross-section see Fig. 68. Abbreviations of beds: Vt - Viita, K - Kuusnõmme, Vs - Vesiku, S - Soeginina. 1 - dolostone; 2 - argillaceous dolostone; 3 - skeletal grainstone; 4 - pelletal-skeletal limestone; 5 - biomicritic limestone; 6 - reef limestone; 7 - reef dolomite; 8 - argillaceous marlstone; 9 - mudstone; 10 - argillaceous limestone, calcareous marl; 11 - dolomitic marl; 12 - stage boundary; 13 - stratigraphical hiatus; 14 - the present denudation surface.

Saaremaa Island, on Muhu Island and in mainland Estonia. In the northwest direction the Paramaja Member is laterally replaced by bioclastic limestones (wackestones to grainstones) of the Ninase Member containing also bioherms. In many sections, a tongue of the Paramaja marlstones overlaps the Ninase limestones.

In Estonia (Tõlla, Ikla, Ruhnu, Ohesaare drill sections) the **Riga Formation** is represented only by its lower part - the Tõlla Member, which is characterized by graptolite-bearing grey mudstones. Northwards it is replaced by the greenish-grey marlstones of the Mustjala Member and upwards with the marlstones of the Paramaja Member, both belonging to the Jaani Formation (Figs. 69, 70).

Jaagarahu Stage

The present unit was established by Luha (1933) provisionally as the Muhu-Kurevere Stage, later as the Jaagarahu Stage (Luha 1946). It corresponds roughly to the upper half of the "Untere Oeselsche Gruppe (Schicht)" by Schmidt (1858, 1881). The subdivision of the stage has been recurrently changed (Bekker 1925, Luha 1930, Аалоэ 1970, Аалоэ и др. 1958, 1976, *etc.*). Recently, some additional units, including the Jamaja and Riksu formations, were introduced (Решения... 1987, Nestor 1995a) and the Muhu dolomites by Luha (1930) were re-established as a formation (Nestor 1995a). V. Nestor (Нестор В. 1984) determined the scope of the stage in subsurface area.

The historical stratotype of the stage is an abandoned quarry at Jaagarahu supplemented with the Jaagarahu drill core in the interval of 0.3 to 21.4 m (Аалоэ 1970). The Jaagarahu Stage spreads on Saaremaa and Muhu islands and in the southwestern part of mainland Estonia (Pärnumaa and southern Läänemaa). The outcrop extends as a 10—30-m-wide belt from Vilsandi and Vaika islands through northern Saaremaa and Muhu as far as Eidapere and Tori at Tallinn - Pärnu railway (Fig. 71). The main localities are the quarries at Jaagarahu, Tagavere, Koguva and Anelema, and recent and ancient coastal cliffs at Vilsandi, Abula, Panga (Photo 22), Pulli (Oiu), Üügu, Püssina, Kesselaid, Salevere and Kirbla (Photo 23). The thickness of the stage is variable and increases southwestwards from 32.3 m in the Viki core to 145.0 m in the Ohesaare core.

In the western part of Saaremaa, the Jaagarahu Stage is dominated by comparatively pure limestones, while dolomites are prevailing in the eastern part of Saaremaa, on Muhu Island and in mainland Estonia. Reefs (bioherms and mounds) are widespread in the Jaagarahu Stage, especially in its lower part (Vilsandi beds and Kesselaid Member). In the South-Estonian Confacies Belt, the lower part of the Jaagarahu Stage is represented by the marlstones of the Jamaja Formation, and the upper part by the nodular biomicritic limestones of the Sõrve Formation. Temporal analogues of the latter formation are absent in northern sequences due to the long stratigraphical hiatus (Nestor & Nestor 1991). In the Mid-Estonian Confacies Belt, the lower boundary of the stage has been drawn by an abrupt increase in the carbonate content of the rocks coinciding with the base of the Jaagarahu and Muhu formations. In the more argillaceous sequences of the South-Estonian Confacies Belt and transition area, the lower boundary is determined by the appearance of chitinozoans of the *Linochitina cingulata* Biozone (Nestor, V. 1994) at the base of the Jamaja and Riksu formations.

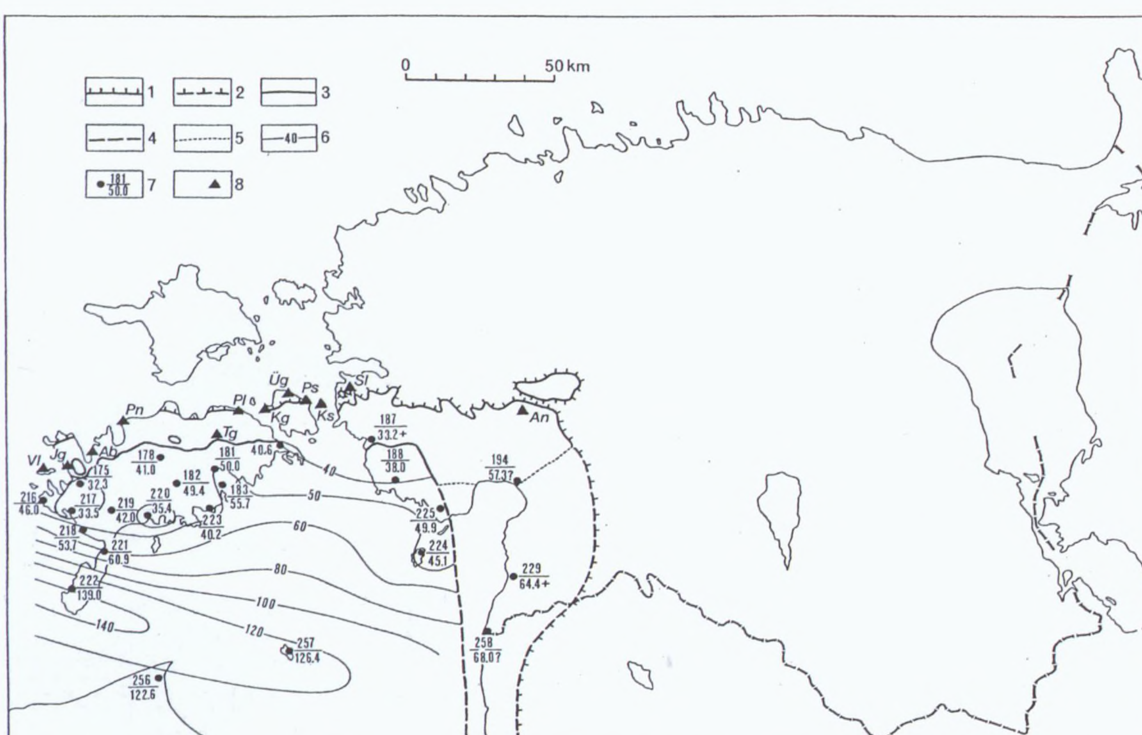


Fig. 71. Sketch-map showing the present distribution and the thickness of the Jaagarahu Stage in Estonia. Numerator marks the number of the borehole, denominator shows the thickness of the stage in metres. 1 - contour of rock distribution; 2 - contour of subsurface distribution; 3 - limit of outcrop belt; 4 - limit of pre-Devonian erosion; 5 - margin of the Devonian cover; 6 - isopach; 7 - borehole; 8 - exposure locality: Vl - Vilsandi, Jg - Jaagarahu, Ab - Abula, Pn - Panga, Tg - Tagavere, Pl - Pulli, Kg - Koguva, Üg - Üügu, Ps - Püssina, Ks - Kesselaid, Sl - Saluvere, An - Anelema.

Photo 22. Panga Cliff on Saaremaa Island. The Jaani Stage crops out in the lower and the Jaagarahu Stage in the upper part. *Photo by Kaarel Orviku.*

The Jaagarahu Stage contains a wide spectrum of fossils from lagoon-related eurypterids and thelodonts to deep-water communities of chitinozoans, ostracodes and trilobites. Severe dolomitization has destroyed skeletal remains over a vast area in the eastern part of the stage (*e.g.* in the Muhu Formation). The most characteristic fossils are as follows (abbreviations: *jg* - Jaagarahu Formation, *jm* - Jamaja Formation, *srv* - Sõrve Formation, *rks* - Riksu Formation, *mh* - Muhu Formation, *V* - Vilsandi beds, *M* - Maasi beds, u.pt. - upper part): *Vikingia tenuis* (Nestor) (*jgV*), *Ecclimadictyon astrolaxum* Nestor (*jgM*), *Favosites mirandus* Sokolov (*jgV*), *Thecia confluens* (Eichwald) (*jgM*), *Coenites juniperinus* Eichwald (*jgV,M*), *Acervularia ananas* (L.) (*jgV*), *Kodonophyllum truncatum* (L.) (*jgM*), *Dolerorthis rustica* (Sowerby) (*jm*), *Howellella cuneata* Rubel (*jg,mh,srv*), *Encrinurus balticus* Männil (*jm*), *Warburgella estonica* Männil (*jgM*), *Craspedobolbina insulicola* Martinsson (*jm*), *Leptobolbina quadricuspidata* Martinsson (*srv*), *Conochitina lagena* Eisenack (*jm, rks*), *C. pachycephala* Eisenack (*jm, srv*), *C. cribrosa* Nestor (*srv*), *Ozarkodina sagitta rhenana* Walliser (*jgV*), *Kockelella walliseri* Helfrich (*jm*), *K. amsdeni* Barrick et Klapper (*srv*), *Monograptus flemingii* Salter (*jm, srv*), *Gothograptus nassa* (Holm) (*srv*, u.pt.), *Logania taiti* (Stetson) (*srv*). Taking into account the few findings of zonal species of graptolites in the Ohesaare core, it seems that the Jaagarahu Stage probably spans from the *M. flexilis* Biozone (partly) to the *Gothograptus nassa* Biozone (Table 8). The main, lower part of the Jaagarahu Stage consists of the Jaagarahu, Muhu, Riksu and Jamaja formations laterally replacing one another (Figs. 69,70). The upper part of the stage is represented by the Sõrve Formation which is distributed only in the South-Estonian Confacies Belt; in the northern sequences a stratigraphical cap corresponds to it.

Photo 23. The West-Estonian Klint at Kirbla with reef-mounds of the Muhu Formation (Jaagarahu Stage). *Photo by A. Raukas.*

The **Jaagarahu Formation** occurs in northwestern Saaremaa and consists of very variable, prevailingly sparitic limestones of shallow-water origin. Coral-stromatoporoid limestones, including reefs or bioherms, and fine-grained skeletal and pelletal grainstones are the most widespread rocks. In some places they are dolomitized. The formation contains some bands of lagoonal argillaceous dolostones, the so-called eurypterus and pattern dolomites which divide the formation into three subunits: the Vilsandi, Maasi and Tagavere beds. The Vilsandi beds comprise an abundance of large bioherms. The Maasi beds contain biomicritic interlayers of deeper-water genesis. The Tagavere beds are capped by the thickest (5-8 m) deposit of lagoonal dolomites treated sometimes as the Selgase Member. The thickness of the Jaagarahu Formation varies from 32 to 46 m (Fig. 71).

The **Muhu Formation** is distributed in northeastern Saaremaa, on Muhu Island and on mainland Estonia north of the Pärnu latitude. It consists mostly of rather monotonous flaggy dolomites containing numerous large massive reef-mounds (Fig. 70, Photo 23) in its lower part (Kesselaid Member). Almost everywhere the thickness (20 to 40 m) of the formation is uncomplete due to the post-Silurian denudation.

The **Riksu Formation** bounds the Jaagarahu and Muhu formations from the south, and spreads along the southern coasts of Saaremaa, the Tõstamaa Peninsula and around Pärnu Bay. It is mostly represented by nodular biomicritic to micritic limestones containing layers of argillaceous limestones and marlstones causing a cyclical nature of the sequence. In the eastern part of the distribution area, the rocks are usually dolomitized. In places (Pärnu, Kihnu, Seliste), the Riksu Formation is underlain by a tongue of the Jamaja Formation and, in most places, it is overlain by the tongue of the Jaagarahu or Muhu formations. The thickness of the Riksu Formation varies from 34.5 m in the Nässumaa borehole to 50.8 m in the Kaugatuma borehole.

The **Jamaja Formation** forms the lower part of the Jaagarahu Stage in the South-Estonian Confacies Belt. The formation is represented by different marl- and mudstones. The thickness of the formation reaches 95.2 m in the Ohesaare section.

The **Sõrve Formation** overlies the Jamaja Formation in the southernmost sections of Estonia (Ohesaare, Ruhnu, Ikla). It is represented by biomicritic to micritic nodular limestones (pack- and wackestones) similar to the Riksu Formation but lying stratigraphically higher in the sequence and corresponding to the hiatus in northern sequences. The thickness of the formation reaches 49.8 m in the Ohesaare boring.

Rootsiküla Stage

The Rootsiküla strata were established by Bekker (1925) as the Rootsiküla-Kaarma Substage of the Saaremaa Stage (= "Obere Oeselsche Gruppe" by Schmidt 1858, 1881). Later Luha (1933) raised the unit into the stage rank and introduced the name Kaarma. Einasto (Эйнасто 1970) motivated the use of the name Rootsiküla. He also defined the boundaries of the stage and subdivided it into beds. Viita quarry, the historical stratotype, has been destroyed. The Kipi drill core in the interval of 25.6 to 53.6 m has been treated as the hypostratotype of the stage (Эйнасто 1970, Nestor 1993). The Rootsiküla Stage spreads in middle and southern Saaremaa, in the western part of the Tõstamaa Peninsula and on Kihnu and Ruhnu islands. The outcrop forms a 4–10-km-wide belt running through the central part of Saaremaa from Atla to the Kübassaare Peninsula. On mainland, it reaches the Seliste Village on the Tõstamaa Peninsula (Fig. 72). The main localities are the coastal cliffs at Elda, Soeginina, Anikaitse and Kübassaare, the Vesiku Rivulet and an abandoned quarry at Pamma. In Estonia, the full thickness of the stage varies from 20 to 40 m and increases rapidly southwards, towards the Kurzeme Peninsula (Fig.72).

The Rootsiküla Stage consists of various skeletal, pelletal, lithoclastic, coquinoid and micritic limestones cyclically interbedding with argillaceous sedimentary dolomites (the so-called *Eurypterus* and pattern dolomites). Limestones form the lower and dolomites the upper part of the shallowing-up sedimentary cycles. Microbial-algal structures (oncolites, stromatolites) are frequent. The limestones are often dolomitized and in the eastern part of the distribution area the whole sequence consists completely of dolomites. The lower boundary of the stage has been determined at the base of a stratum of skeletal pack- or grainstones forming the lowermost part of the Viita beds which disconformably overlie the first thick stratum of lagoonal dolomites (Selgase Member of the Jaagarahu Formation) in the Middle-Estonian Confacies Belt and unnamed members of skeletal grainstones at the top of the Sõrve and Riksu formations in the South-Estonian Confacies Belt. In the Ohesaare boring these basal-Rootsiküla nodular packstones contain the *Beirichia subornata* ostracode fauna characteristic to the Mulde Marls on Gotland and correlatable with the *Gothograptus nassa* graptolite zone.

The Rootsiküla Stage contains sparsely distributed and specific fossil biota. Eurypterids, thelodonts, leperditian ostracodes, specific gastropods, bivalves, oncolites, stromatolites are common indicating the shallow, near-shore environments. The most typical species are (abbreviations: *rt* - Rootsiküla Formation, *Vt* - Viita beds, *K* - Kuusnõmme beds, *Vs* - Vesiku beds, *S* - Soeginina beds): *Araneosustroma stelliparratum* (Nestor) (*rtK*), *Parastriatopora commutabilis* Klaamann (*rtK,S*), *Howellella cuniculi* Rubel (*rtVt*), *Straparollus (S.) helicites* (Sowerby) (*rtVs*), *Murchisonia (Hormotoma) compressa* Lindström (*rt*), *Hermannina phaseola* (Hisinger) (*rt*), *Bingeria vesikuensis* Sarv (*rtVt*), *Beirichia subornata* Martinsson (*rtVt*), *Balteurypterus remipes tetragonophtalmus* (Fischer) (*rtVt, Vs*), *Ctenognathodus murchisoni* (Pander) (*rt*), *Ozarkodina bohemica bohemica* (Walliser) (*rtVt*), *Logania martinssoni* Gross (*rt*), *Tremataspis schmidti* (Rohon) (*rtVt,K,Vs*). The presence of *Ozarkodina bohemica bohemica* enables to date the Rootsiküla Stage as top Wenlock - basal Ludlow.

In Estonia, the Rootsiküla Stage is represented by the Rootsiküla and Sakla formations laterally replacing each other (Fig. 70). The **Rootsiküla Formation** is distributed on Saaremaa Island, except its easternmost part. The formation is represented by cyclically alternating limestones (often secondarily dolomitized) and argillaceous sedimentary dolostones. Limestones are prevailingly skeletal and pelletal grainstones containing in places oolites, oncolites and intraclasts. Biomicritic and micritic limestones also occur on some levels. Argillaceous dolomites and dolomitic marlstones form the upper part of the sedimentary cycles. They are laminated *Eurypterus* dolomites or massive bioturbated pattern dolomites. Four cycles have been distinguished in the

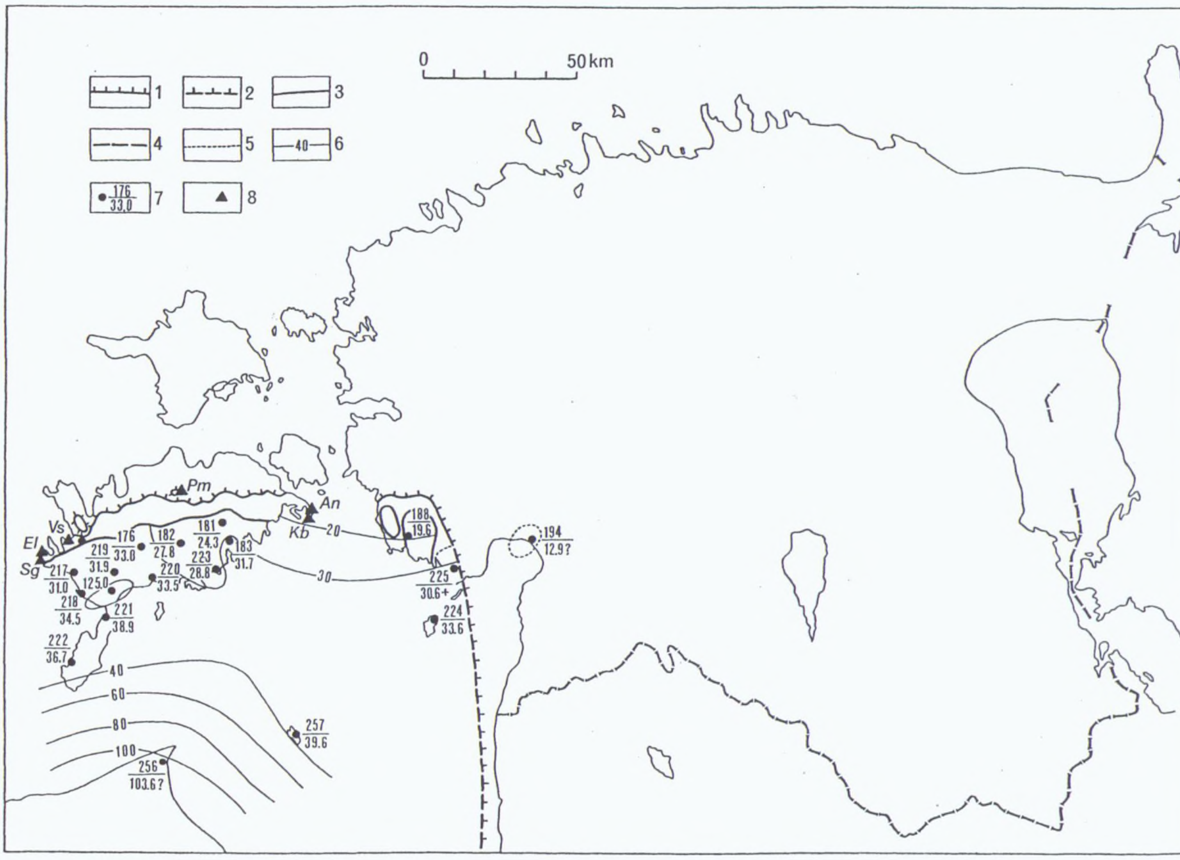


Fig. 72. Sketch-map showing the present distribution and the thickness of the Rootsiküla Stage in Estonia. Numerator marks the number of the borehole, denominator shows the thickness of the stage in metres. 1 - contour of rock distribution; 2 - contour of subsurface distribution; 3 - limit of outcrop belt; 4 - limit of pre-Devonian erosion; 5 - margin of the Devonian cover; 6 - isopach; 7 - borehole; 8 - exposure locality: Sg - Soeginina, El - Elda, Vs - Vesiku, Pm - Pamma, Kb - Kübassaare, An - Anikaitse.

Rootsiküla Formation (Эйнасто 1970) defined as beds. The lowermost, Viita beds begin with comparatively normal-marine biomicritic limestones and end with typical *Eurypterus* and pattern dolomites. The Kuusnõmme beds form a thin uncomplete cycle with coral-stromatoporoid or oncolitic limestone in the lower and pattern dolomite in the upper part. The Vesiku beds begin with sparitic limestones, pelletal, oolitic, coquinoid or lithoclastic grain- or floatstones and their dolomitized counterparts and end with the thickest band of *Eurypterus* and pattern dolomites. The Soeginina beds form an untypical cycle with totally dolomitized porous grainstones in the lower, thick stromatolite band in middle and pattern dolomites in the upper part. The **Sakla Formation** is developed in southeastern Saaremaa, on Kihnu Island, and on the Tõstamaa Peninsula in mainland Estonia. It is represented by comparatively monotonous, thick-bedded, bioturbated dolomites with numerous pyrite patterns and undefinite cyclicity.

LUDLOW SERIES

Paadla Stage

The present unit was treated by Schmidt (1892) as the *Ilionia* Schichten (Beds) of the "Obere Oeselsche Gruppe". Bekker (1925) introduced the geographical name Paadla and considered these beds as a substage of the Saaremaa Stage. Aaloe (Аалоэ 1963a) included equivalents of the present-day Himmiste beds earlier correlated with the Kaarma Stage. Aaloe *et al.* (Аалоэ и др. 1976) redefined the upper boundary of the stage excluding the Tahula beds and gave the current stratification of the stage (Table 8).

The historical stratotype of the stage - Paadla quarry, has been destroyed. The Kuressaare-GI (Kingissepa) drill core in the interval of 19.8 to 43.4 m has been chosen as the neostratotype of the stage (Nestor 1993). The rocks of the Paadla Stage occur in middle and southern Saaremaa, on Kihnu and Ruhnu islands and in the western part of the Tõstamaa Peninsula in mainland Estonia. The outcrop forms a 12–20-km-wide belt passing through midsouthern Saaremaa from the vicinity of Karala to Kõiguste and extending eastwards as far as the Tõstamaa Settlement on mainland (Fig. 73B). The main localities are cliffs at Roopa and Katri, quarries at Lümanda, Himmiste-Kuigu, Kogula, Kaarma and Uduvere and the walls of the Kaali meteorite crater. The full thickness of the stage varies from 2.8 m in the Kihnu borehole to 28.4 m in the Kaugatuma borehole, increasing westwards (Table 9).

The Paadla Stage consists of various bioclastic, pelletal and argillaceous limestones containing coral-stromatoporoid bioherms and biostromes in the west and different primary and secondary dolomites in the east. The base of the stage coincides with the top of the Soeginina pattern dolomites of the Rootsiküla Stage, overlain by argillaceous limestones and dolomites with *Didymothyris didyma* and *Ilionia prisca*.

In Estonia, the Paadla Stage contains a rather specific shallow-water fauna of corals, stromatoporoids, agnathans, brachiopods and molluscs in the north-west, more diverse shelly

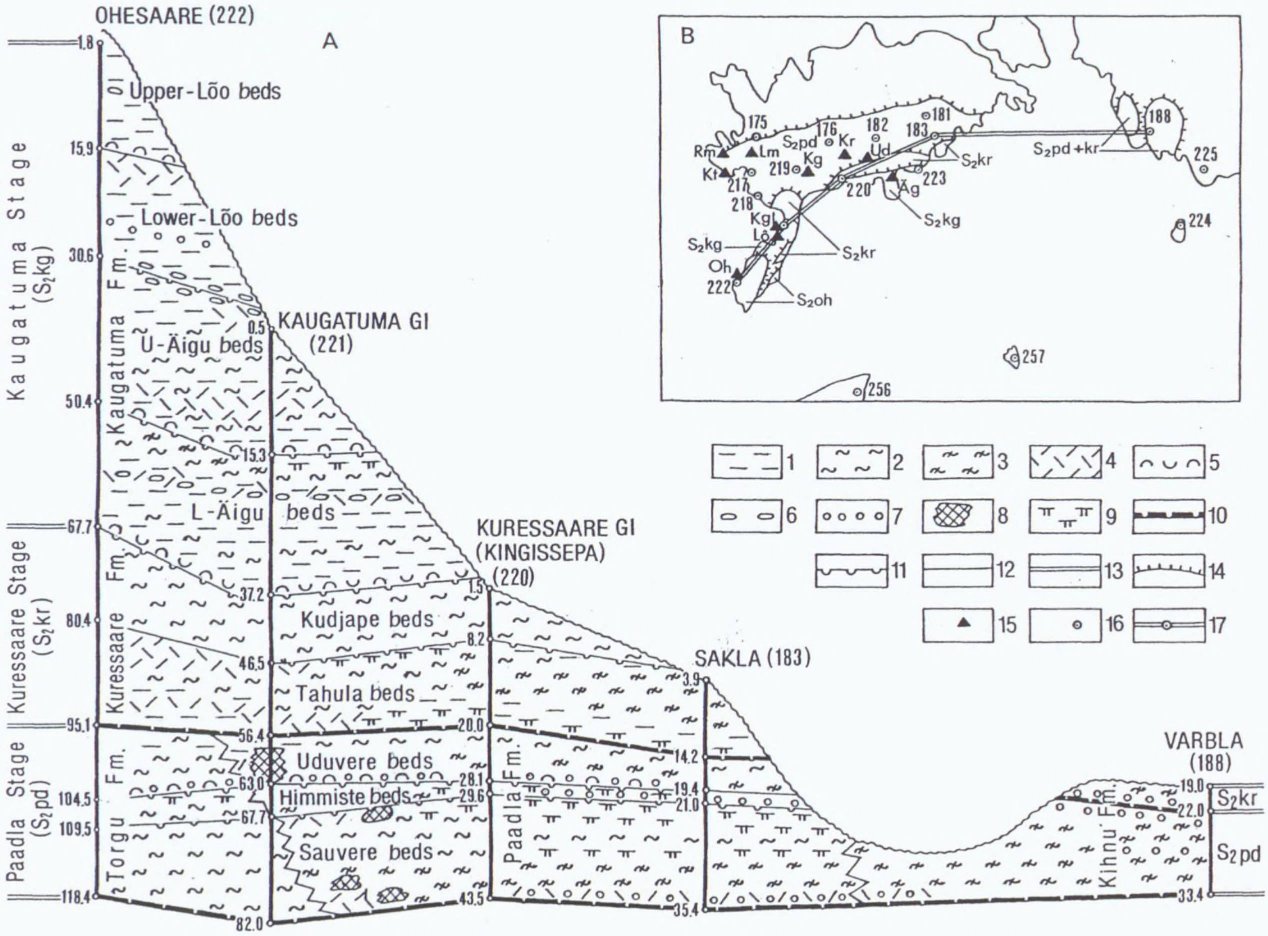


Fig. 73. Stratigraphical cross-section of the Upper Silurian Ludlow and Přidoli rocks (A) and location of outcrops, boreholes and exposures (B) (by R. Einasto). For the names and numbers of boreholes see Table 9: 1 - marlstone; 2 - nodular biomicritic limestone; 3 - argillaceous dolomite; 4 - skeletal (crinoidal) grainstone; 5 - coquinoid limestone; 6 - lithoclastic rudstone (conglomerate); 7 - pelletal-skeletal limestone; 8 - reef limestone; 9 - dolomitic marlstone; 10 - stratigraphical hiatus; 11 - discontinuity surface; 12 - lithostratigraphical boundary; 13 - stage boundary; 14 - limits of outcrops of the regional stages: Paadla (S_2pd), Kuressaare (S_2kr), Kaugatuma (S_2kg), Ohesaare (S_2oh); 15 - localities: Kt - Katri, Rm - Riiumägi, Lm - Lümanda, Kg - Kogula, Kr - Kaarma, Ud - Uduvere, Äg - Äigu, Kgt - Kaugatuma, Lõ - Lõu, Oh - Ohesaare; 16 - borehole; 17 - transect of stratigraphical cross-section (A).

fauna in the south-west, and almost barren dolomites in the east. The most typical species are as follows (abbreviations: *pd* - Paadla Formation, *tr* - Torgu Formation, *kh* - Kihnu Formation, *S* - Sauvere beds, *H* - Himmiste beds, *U* - Uduvere beds, m.pt. - middle part, u.pt. - upper part): *Conochitina latifrons* Eisenack *(tr), Angochitina elongata* Eisenack *(tr), Parallelostroma typicum* (Rosen) *(pd), Lophiostroma schmidtii* (Nicholson) *(pd), Thecia swindereniana* (Goldfuss) *(pd), Laceripora cribrosa* Eichwald *(pdU), Phaulactis cyathophylloides* Ryder *(pd), Didymothyris didyma* (Dalman) *(pd, tr), Howellella elegans* Muir-Wood *(pd, tr), Ilionia prisca* Hisinger *(pd, tr); Cardiola interrupta* Sowerby *(pd, tr), Megalomphala taenia* (Lindström) *(pdU), Hemsiella hemsiensis* Martinsson *(pdU, tr* u.pt*.), Neobeirichia nutans* (Kiesow) *(tr), Hammariella pulchrivelata* Martinsson *(pdU, tr), Amygdalella paadlaensis* Sarv *(pd, tr), Balizoma obtusus* (Angelin) *(pdU), Ozarkodina crispa* (Walliser) *(pdU, tr* u.pt.*), O.* cf. *snajdri* (Walliser) *(tr* m.pt*.), Tremataspis mammillata* Pander *(pdH, kh), Phlebolepis elegans* Pander *(pd, tr, kh), Andreolepis hedei* Gross *(pdU, tr* u.pt.*)*. The presence of chitinozoan species *Conochitina latifrons* and *C. lauensis* enables to correlate the strata of the Paadla Stage in Estonia with *scanicus* to *tauragensis* (*leintwardinensis*) graptolite zones of middle Ludlow in Latvia which suggests that the basal Ludlow beds are probably absent (Nestor & Nestor 1991).

In Estonia, the Paadla Stage is represented by the Paadla, Torgu and Kihnu formations (Аалоэ и др. 1976) laterally replacing one another (Fig. 73). The **Paadla Formation** occurs in the southern part of Saaremaa, except the Sõrve Peninsula. It is dominated by argillaceous biomicritic to sparitic limestones and dolomites with bands of marlstones, coral-stromatoporoid biostromes, pelletal and coquinoid (*Didymothyris*) limestones. The formation is subdivided into the Sauvere, Himmiste and Uduvere beds (Клааманн 1970a). The Sauvere beds are represented by nodular argillaceous bioturbated biomicritic limestones (pack- and wackestones), containing small bioherms in the west and being gradually replaced by argillaceous dolomites towards the east. The Himmiste beds are mainly represented by micro- to cryptolaminated argillaceous dolomites with the remains of eurypterids and agnathans. They also contain bands of pelletal-skeletal grainstones at the base and (less often) at the top. The well-known Kaarma building dolomite is tentatively attributed to these beds now (Einasto in Kaljo & Nestor 1990, p. 173). The Uduvere beds are represented by variable rocks

Table 9. Base depths and thicknesses of the upper Silurian regional stages in Estonian boreholes

No.in Figs.	Borehole	Paadla Stage	Kuressaare Stage	Kaugatuma Stage	Ohesaare Stage
176	Kaarmise	14.8/14.5+	-	-	-
217	Kipi-GI	22.6/22.0+	-	-	-
218	Riksu-803	39.8/20.0+	-	-	-
219	Kõrkküla-863	17.6/10.3+	-	-	-
	Tehumardi-864	39.0/14.0+	-	-	-
	Möldri-865	43.0/23.0	20.0/9.0+	-	-
220	Kuressaare-GI	43.4/23.4	20.0/18.5+	-	-
	Tahula-709	36.6/26.6	10.0/8.5+	-	-
	Pihtla-816	40.5/23.5	17.0/13.0+	-	-
183	Sakla-GI	34.6/20.4	14.2/8.8+	-	-
188	Varbla-502	33.4/11.4	22.0/3.0+	-	-
224	Kihnu-526	93.5/2.8?	90.7/4.9+	-	-
221	Kaugatuma-GI	85.0/28.4	56.6/19.4	37.2/36.6+	-
	Muratsi-805	55.9/16.1	39.8/19.2	20.6/16.7+	-
	Kailuka-817	65.5/21.5	44.0/22.9	21.1/14.6+	-
223	Nässumaa-825	65.5/23.5	42.0/21.2	20.8/14.8+	-
222	Ohesaare-GI	118.4/23.3	95.1/27.4	67.7/65.9	-
	Sõrve-514	151.0/6.4+	144.6/23.4	121.2/85.7	35.5/5.7+
	Kaavi-869	-	150.0/19.0+	131.0/72.0	59.0/29.0+
257	Ruhnu-500	248.0/26.0	222.0/5.4	216.6/41.6	175.0/28.9+
256	Kolka-54	305.4/19.7	285.7/20.7	265.0/59.6	205.4/39.0

of shallow-water genesis: skeletal-, pelletal- lithoclastic-, oncolitic grainstones, packstones and rudstones, interbedded with bands of marlstones, coral-stromatoporoid biostromes, *etc*. Rocks are partly or totally dolomitized, particularly east of the Kuressaare Town.

The **Torgu Formation** spreads on the Sõrve Peninsula and Ruhnu Island. It mainly consists of nodular argillaceous biomicritic limestones with rather rich shelly fauna, but corals and stromatoporoids are rare.

The **Kihnu Formation** is distributed on the Tõstamaa Peninsula and Kihnu Island. It is represented by monotonous platy dolomites (below) and argillaceous dolomites (above) of reduced thicknesses containing agnathans of Paadla and Kuressaare ages, respectively (Эйнасто и др. 1977), and consequently spanning from the Paadla to Kuressaare Stage.

Kuressaare Stage

The Kuressaare Stage was separated from the Kaugatuma Stage by Klaamann (Клааманн 1970a). Later on, the Tahula beds were added to the stage from among the Paadla Stage (Аалоэ и др.1976). The stratotype section is the Kuressaare-GI (Kingissepa) drill core in the interval of 1.5 to 19.8 m (Аалоэ и др. 1976). The Kuressaare Stage spreads in the southernmost Saaremaa, on Ruhnu and Kihnu islands and in the southwestern part of the Tõstamaa Peninsula. The outcrop forms a 2-to-10km-wide belt along the southern coast of Saaremaa Island (Fig. 73). The rocks of the stage crop out in temporary excavations and ditches in the Town of Kuressaare and its surroundings. The full thickness of the stage varies from 5.4 m in the Ruhnu to 27.4 m in the Ohesaare borehole (Tab. 9).

The Kuressaare Stage consists of different marlstones (below) and nodular argillaceous biomicritic limestones (above), both containing interlayers of skeletal, lithoclastic and coquinoid grain-, float- and rudstones. The base of the stage coincides with a sharp increase in the clay component and appearance of the elements of a new microfossil assemblage: *Pterochitina perivelata, Ozarkodina remscheidensis* aff. *scanica, Calcibeirichia altonodosa, Thelodus sculptilis*.

The Kuressaare Stage contains a rich assemblage of shelly fossils, especially ostracodes.The brachiopod *Atrypoidea prunum* is extremely numerous and forms coquina banks in the upper, Kudjape beds of the stage. The most typical species are as follows (abbreviations: *T* - Tahula beds, *K* - Kudjape beds): *Pterochitina perivelata* (Eisenack) *(T, K), Conochitina granosa* Laufeld *(T, K), "Parallelopora" ornata* Mori *(K), "Paleofavosites" moribundus* Sokolov *(K), Entelophyllum articulatum* (Wahlenberg) *(K), Tryplasma loveni* (M.Edw. et Haime) *(K), Atrypoidea prunum* (Dalman) *(T, K), Calcaribeirichia altonodosa* Sarv *(T, K), Plicibeirichia numerosa* Sarv *(K), Retisaculus sulcatus* Gailite *(K), Limbinariella malornata* Sarv *(K), Calymene flabellata* Männil *(K), Pulcherproetus kuressaarensis* (Männil) *(K), Ozarkodina remscheidensis* aff. *scanica* (Jeppsson) *(T, K), O. snajdri parasnajdri* Viira et Aldridge *(T, K), Thelodus sculptilis* Gross *(T, K)*. The Kuressaare Stage has been indirectly correlated with the upper part of the Ludlow Series.

In Estonia, the Kuressaare Stage is represented by the

Kuressaare Formation which is subdivided into Tahula beds (below) and Kudjape beds (above) (Аалоэ и др. 1976). The Tahula beds mainly consist of argillaceous or dolomitic marlstones with bands of various bio- and lithoclastic limestones. The content of the calcareous component increases northeastwards.

The Kudjape beds are represented by nodular argillaceous biomicritic limestones containing coquinoid interlayers with *Atrypoidea prunum* and numerous colonial rugose corals.

PŘIDOLI SERIES

Kaugatuma Stage

Twenhofel (1916) introduced the name Kaugatoma in the sense of the upper subdivision (Zone) of his Oesel Formation (="Upper Oeselsche Gruppe" by Schmidt 1858, 1881). The present-day limits and stratification of the stage were proposed by Klaamann (Клааманн 1970a) who separated the Kuressaare Stage as an independent unit. The historical stratotype of the stage is the Kaugatuma Cliff supplemented by Kaugatuma-GI drill core in the interval of 0.6 to 37.2 m (Решения... 1987, Nestor 1993). The rocks of the Kaugatuma Stage are distributed on the southern peninsulas of Saaremaa Island, and also on Ruhnu and Abruka islands. They crop out in the northern part of the Sõrve Peninsula and on the Roomassaare, Muratsi, Vätta and Leina peninsulas (Fig. 73). The main localities are the cliffs at Kaugatuma and Lõu and the abandoned quarries at Muratsi, Väike-Rootsi and Äigu. The full thickness of the stage varies from 41.6 m in the Ruhnu borehole to 85.7 m in the Sõrve-514 borehole (Table 9).

The Kaugatuma Stage is represented by interbedded marlstones and bioclastic to coquinoid limestones displaying certain cyclicity. In the lower part of the cycle, marlstones are dominating; in the upper part the limestone interlayers become more frequent and the cycle ends with a thick (2-4 m) deposit of crinoidal limestones. In the upper cycles and southwards, the role of limestone layers decreases. The lower boundary of the stage coincides with a notable increase in the clay component. Higher in the sequence, there appear species of ostracodes *Amygdalella nasuta, Sleia equestris, Frostiella groenvalliana, Neobeirichia buchiana*; chitinozoans *Ancyrochitina fragilis*; conodonts *Ozarkodina remscheidensis eosteinhornensis etc.*

The Kaugatuma Stage contains a rich shelly fauna, particularly ostracodes. The guide fossils of the stage are as follows (abbreviations: *Ä* - Äigu beds, *L* - Lõo beds): *Ancyrochitina fragilis* Eisenack *(Ä, L), Fungochitina pistilliformis* (Eisenack) *(L), Densastroma astroites* (Rosen) *(Ä), Actinostromella vaiverensis* Nestor *(Ä), Parallelostroma tuberculatum* (Yavorsky) *(Ä), Favosites pseudoforbesi muratsiensis* Sokolov *(Ä), Syringopora blanda* Klaamann *(Ä, L), Cystiphyllum cylindricum* Lonsdale *(Ä), Atrypoidea prunum* (Dalman) *(Ä), Stegerchynchus pseudobidentatus* (Rybnikova) *(Ä, L), Acaste dayiana* Richter et Richter *(Ä), Pulcherproetus nieszkowskii* (Männil) *(Ä), Amygdalella nasuta* Martinsson *(Ä, L), Sleia equestris* Martinsson *(Ä), Frostiella groenvalliana* Martinsson *(Ä), Nodibeirichia tuberculata* (Klöden) *(L), Crotalocrinites rugosus* (Miller) *(Ä, L), Ozarkodina remscheidensis eosteinhornensis* (Walliser) *(Ä), O. remscheidensis remscheidensis* (Walliser) *(L), O. confluens nasutus* (Viira) *(L), Thelodus admirabilis* Märss *(Ä), Nostolepis gracilis* Gross *(Ä, L)*. The presence of *Ozarkodina remscheidensis eosteinhornensis* in the lower part of the Kaugatuma Stage shows that its base roughly corresponds to the Ludlow/Přidoli boundary.

In Estonia, the Kaugatuma Stage is represened by the **Kaugatuma Formation** which is subdivided into the Äigu (below) and Lõo beds (above) (Fig. 73). The Äigu beds consist of two regressive sedimentary cycles, sometimes regarded as the Lower and Upper Äigu beds (Nestor 1995a). In the lower part of these cycles, marlstone layers are prevalent; in their upper part limestones dominate. Among the latter, coarse-grained crinoidal limestones are the most typical rocks, but interlayers of coquinoid or bio-lithoclastic limestones are also quite common, among these bands with *Atrypoidea prunum*. Both cycles are capped by a thick deposit of crinoidal limestones. The Äigu beds roughly correspond to the ostracode *Frostiella groenvalliana* Biozone.

The Lõo beds also consist of two sedimentary cycles of the same type but marlstones prevail throughout the whole sequence. Bioclastic to coquinoid limestones occur as thin intercalations. A thicker band of crinoidal limestones occurs at the top of the lower cycle considered sometimes as the Lower Lõo beds. The Upper Lõo beds lack crinoidal limestones at the top. The Lõo beds roughly correspond to the *Nodibeirichia tuberculata* Biozone.

Ohesaare Stage

The Ohesaare strata were originally established by Bekker (1925) as a substage of the Saaremaa Stage and were raised into the stage rank by Luha (1933). Klaamann (Клааманн 1970a) defined the lower boundary of the stage. Aaloe *et al.* (Аалоэ и др. 1976) distinguished the Kaavi Member. The Ohesaare Cliff is the historical stratotype of the stage. Ohesaare-2 drill core at the depth of 4.10 m has been selected as the boundary stratotype of the stage.

The Ohesaare Stage crops out in the southern part of the Sõrve Peninsula and spreads also on Ruhnu Island under the Devonian cover. The only exposures of the stage are the Ohesaare and Loode cliffs (Fig. 73). In Estonia, the upper limit of the stage is erosional and the stage does not reach its full thickness anywhere. The thickest section (33.7+ m) has been recorded in Kaavi-568 boring.

In Estonia, the Ohesaare Stage is represented by the **Ohesaare Formation** which mostly consists of argillaceous-dolomitic marlstones or calcareous mudstones with thin intercalations of partly to totally dolomitized bio- to lithoclastic limestones. At the base of the stage there is a rather thick (4-5 m) deposit of various thin-bedded bioclastic to micritic limestones with thin intercalations of marlstone. In the upper part of the sequence the argillaceous-dolomitic marlstones are of red colour and contain silt and sand admixture. This part of the sequence is regarded as the Kaavi Member. The lower boundary of the stage coincides with the junction between the marlstones of the Lõo beds and the platy bioclastic limestones in the basal part of the Ohesaare Stage. Above this level there appear some new elements among ostracodes *(Juviella piltenensis, Nodibeirichia protuberans)*, chitinozoans *(Urochitina sp. sp., Eisenackitina lagenicula)* and vertebrates *(Poracanthodes punctatus, Goniporus alatus), etc.*

The Ohesaare Stage contains a rather rich shelly fauna and a diverse association of agnathans and fish remains. Most of the palaeontological records come from the Ohesaare locality and characterize the lowermost part of the stage. From the Kaavi Member (*K*) only vertebrate fossils have been identified up to now. The species characteristic of the whole stage include *Eisenackitina lagenicula* (Eisenack), *Urochitina* cf. *simplex* Eisenack, *U. verrucosa* Eisenack, *Favosites forbesi ohesaarensis* Klaamann, *F. vectorius* Klaamann, *Fistulipora tenuilamellata* (Bassler), *Eridotrypa parvulipora* Ulrich et Bassler, *Shaleria dzwinogrodensis* (Kozlowski), *Collarothyris collaris* (Rubel), *Grammysia obliqua* (McCoy), *Tentaculites scalaris* (Schlotheim), *Lonchidium inaequale* Eichwald, *Calymene conspicua* Schmidt, *Eophacops serotinus* Männil, *Juviella piltenensis* Gailite, *Nodibeirichia protuberans* (Boll), *Klodenia leptosoma* Martinsson, *Orcofabella testata* (Gailite), *Ozarkodina confluens nasutus* (Viira), *Poracanthodes punctatus* Brozen, *Tylodus deltoides* Rohon, *Goniporus alatus* (Gross) *(K)*, *Nostolepis alta* Märss *(K)*.

The presence of the conodont species *Ozarkodina remscheidensis remscheidensis* allows to correlate the Ohesaare Stage with the upper Přidoli.

DEVONIAN

Introduction

The first data about the Devonian of Estonia and fossils date from the first half of the 19th century (Engelhardt & Ulprecht 1830, Kutorga 1835, 1837). A great contribution was made by Asmuss (1856), professor of Tartu University, with his valuable collection of fish fossils from the Aruküla caves (Photo 13). Grewingk (1861, 1879) was the first to describe and correlate the Devonian strata with neighbouring areas. Eichwald (1854) presented the first reasonably complete description of the cross bedding of the Devonian outcropping strata.

Systematic research into the Devonian was started in the first half of the 20th century. During the 1920s-1940s, the outcropping Devonian strata were described and correlated by Bekker (1924a), Orviku (1930c, 1932, 1935b, 1946, 1948), Gross (1930, 1931, 1933, 1934, 1940b, 1942), Obruchev (Обручев 1931, 1933) and Bölau (1943, 1944). In general outline, the classification and correlation dating from that period are valid todate (Сорокин 1981).

In the second half of the century, the lists of the fish fossils related to Devonian strata have been essentially improved (Обручев и Марк-Курик 1965) and new stages differentiated (Марк 1958, 1964). Age correlation of the Devonian strata in Estonia has been revised and adjusted to the internationally acknowledged scale (Mark-Kurik 1991a, 1993c, Valiukevičius 1988, 1994).

In connection with a medium-scale geological mapping thousands of boreholes were made which enabled research into buried Devonian strata (Каяк и др.1970; Каяк и Каяк 1983, 1986, Кырвел и др. 1970, Вяярси и др. 1971, 1981, Ванамб и др. 1975, Кала и др. 1981a, Поливко и др. 1981, Арвисто и др. 1987). In geological mapping, the stratigraphical schemes worked out by the Commission of Stratigraphy of the Baltic and East-European Platform were taken into account (Решения... 1978, Ржосницкая и Куликова 1990).

Mineralogical studies initiated by H. Viiding and carried on by A. Kleesment imparted much valuable information for subdividing and correlating the strata (Вийдинг 1962, 1964, 1965, 1976, Клеесмент 1976, 1977, 1984). The zonal scheme was worked out by J. Valiukevičius (1988, 1994, Валюкевичюс и др. 1986) on the basis of acanthodian scales.

The present chapter is based on the whole bulk of data available on the subject. The sources include the publications and original material of the authors of the chapter, descriptions of sections stored in the Estonian Geological Fund, the results of grain-size and mineralogical analyses, based mainly on fraction 0.1-0.05 mm. Use has also been made of H. Viiding's unpublished results. The recent stratigraphical scheme of the Devonian in Estonia (Table 10) is based on complex studies and was accepted in 1995 by the Devonian Working Group of the Estonian Stratigraphic Commission.

LOWER DEVONIAN

The Lower Devonian with a total thickness of up to 51.5 m is spread in restricted areas in southern Estonia. It is represented by three stratigraphical units of different age, separated from each other by stratigraphical gaps (Table 10, Figs. 74, 79, 80).

Tilžė Stage

Liepinš was the first to acknowledge the Tilžė beds as an independent stratigraphic unit - the Lower Stoniškiai Formation (Лиепиньш 1955). Paškevičius (Пашкевичюс 1959) determined the present stratigraphic extent of the stage and gave it the present name. Faunistically, it was determined by Karatajūtė-Talimaa (Каратаюте-Талимаа 1962). The stratotype of the Tilžė Stage is in the interval of 1104.5-1212 m of the drill core Stoniškiai-1 in southwestern Lithuania.

The Tilžė Stage is spread in southeastern Estonia and covered with rocks of the Rēzekne Stage. It has been determined only in the Laanemetsa and Värska drill cores, but is assumed to be present also in the Väimela drill core. The sediments are absent in the Mõniste High. The total thickness of the stage varies from 2.1 to 17.7 m (Fig. 74).

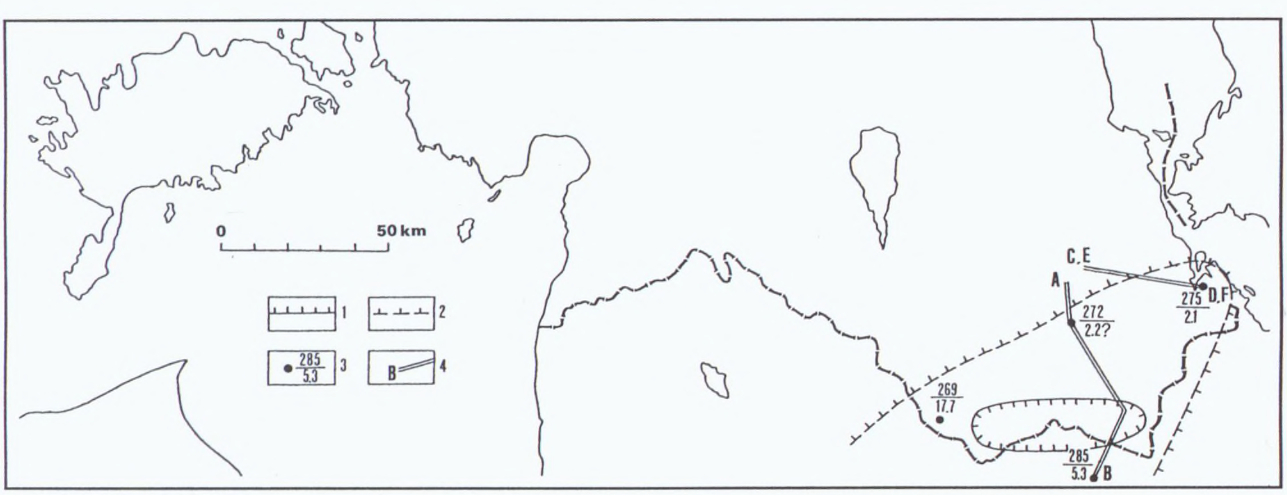


Fig.74. Sketch-map showing the present distribution and the thickness of the Tilžė Stage in Estonia. AB, CD and EF indicate part of the transects of stratigraphical cross-sections presented in figures 75, 76 and 77. Numerator at the borehole sign marks the number (see Fig. 3 for section names), denominator shows the thickness of the stage in metres. 1 - contour of rocks distribution today; 2 - contour of presumable rock distribution; 3 - borehole; 4 - transect of cross-section.

Table 10. The Devonian of Estonia

Series	Stage	Regional Stage	Substage	Group	Formation	Member/Beds*
UPPER	Frasnian	DAUGAVA			DAUGAVA	
		DUBNIKI			DUBNIKI	
		PĻAVIŅAS	Chudovo		PĻAVIŅAS	Izborsk
			Pskov			
			Snetnaya Gora			Snetnaya Gora
MIDDLE	Givetian	? AMATA		Šventoji	AMATA	
		GAUJA			GAUJA	Lode
						Sietiņi
		BURTNIEKI	Abava	Tartu	BURTNIEKI	Abava*
			Salaca			Koorküla*
						Härma*
	Eifelian	? ARUKÜLA			ARUKÜLA	Tarvastu*
						Kureküla*
						Viljandi*
		NARVA	Kernavė		NARVA	Kernavė
			Leivu			Leivu
			Vadja			Vadja
		PÄRNU			PÄRNU	Tamme
						Tori
LOWER	Emsian	RĒZEKNE			LEMSI, RĒZEKNE	
	Pragian	ĶEMERI?			ĶEMERI?	
	Lochkovian	TILŽĖ			TILŽĖ	

The Tilžė Stage lies with a stratigraphical unconformity on the rocks of different stages of Ordovician age (Figs. 75, 76, 77). It has yielded several thelodonts of stratigraphical value: *Turinia pagei* (Powrie), *Turinia* sp., *Nikolivia gutta* Kar.-Tal. and *N. elongata* Kar.-Tal. The other fossil fishes (psammosteid heterostracans, acanthodians) are identified only on the group level (Сорокин 1981).

In the Baltic region, the Tilžė Stage is represented by the Tilžė Formation which in Estonia is composed of grey and purplish-grey horizontally bedded silt- and sandstones with interlayers of grey clay and yellowish-grey dolomite. Silt- and sandstones are predominantly strongly cemented with dolomitic or gypsum (Värska) matrix. Siltstone is often mottled, conglomeritic sandstone occurs in the basal part.

The rocks of the Tilžė Formation are quartzose or feldspatic arenites with the content of quartz 60-85%. Micaceous arenites (content of micas up to 60%) occur in some places.

The heavy fraction is dominated by the group of transparent allothigenic minerals (50-70%). Garnet with a considerable supplement of zircon is prevailing. Tourmaline and apatite are also important (Fig. 78).

Ķemeri Stage

The stage was established by Liepinš (Лиепиньш 1960, 1964). The Ķemeri drill core in the interval of 461 to 547.65 m has been selected as a neostratotype for the Ķemeri Stage (Сорокин 1981). The probable Ķemeri rocks in Estonia have not revealed any fossils which makes the correlation of sections rather difficult.

In Estonia the presumable Ķemeri Stage occurs in a limited area in the southwestern part of the Republic and is identified only in Ikla, Ipiku, Tõlla and Abja drill cores where its thickness varies from 5.9 to 8.4 m (Fig. 79). It lies with a stratigraphical discordance on the Silurian rocks and is covered with deposits of the Rēzekne Stage (Figs. 76,77).

In the Baltic Region, the Ķemeri Stage is represented by

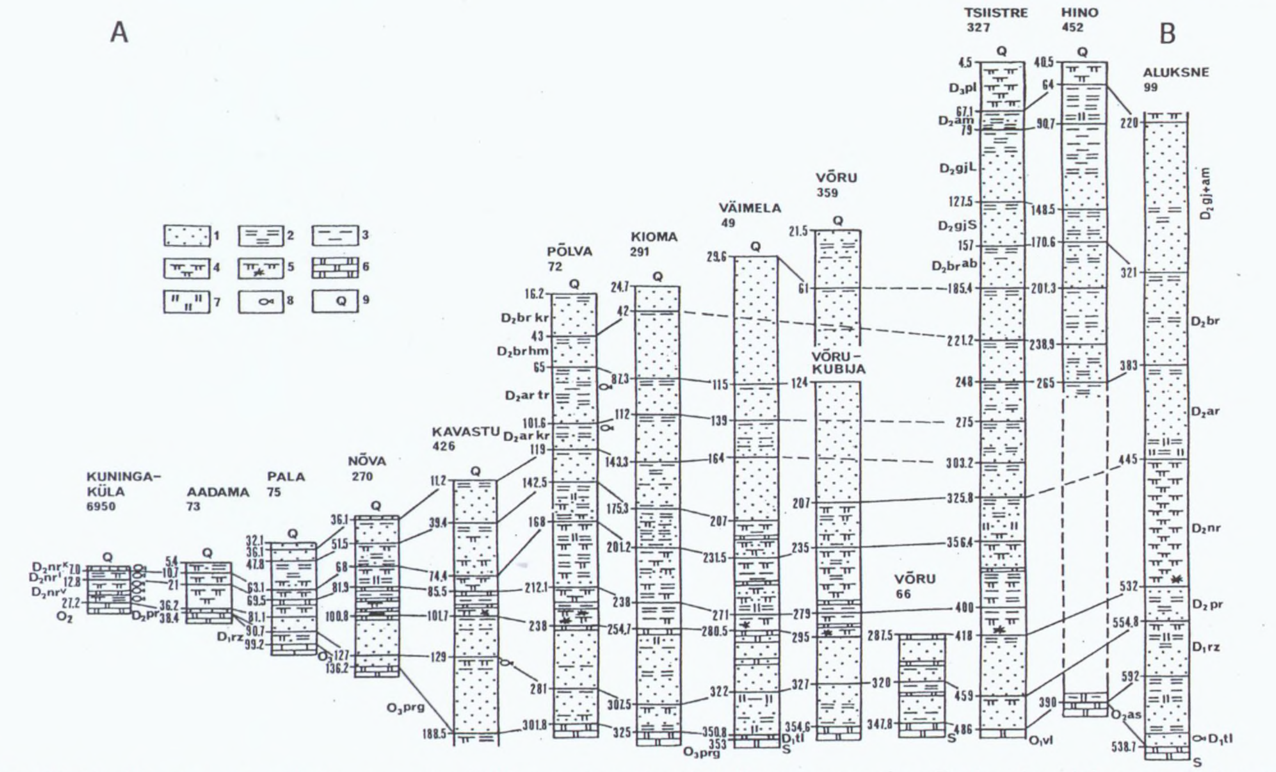


Fig.75. Meridional stratigraphical cross-section (AB) comprising the whole Devonian sequence. For location of the cross-section see Fig. 82. Boreholes: Kuningaküla - 103, Aadama - 169, Pala - 171, Nõva - 212, Kavastu - 215, Põlva - 253, Kioma - 250, Väimela - 272, Võru - 273, Võru-Kubija - 274, Tsiistre - 279, Hino - 283 and Aluksne - 285. Numbers under the name of borehole correspond to its initial number. Separated are Volkhov Stage (O_1vl), Middle Ordovician (O_2), Aseri Stage (O_2as), Upper Ordovician (O_3), Pirgu Stage (O_3prg), Silurian (S); Tilžė (D_1tl), Rēzekne (D_1rz), Narva (D_2nr), Aruküla (D_2ar), Burtnieki (D_2br), Gauja (D_2gj), Amata (D_2am) and Pļaviņas (D_3pl) stages; Vadja (D_2nr^v), Leivu (D_2nr^l) and Kernavė (D_2nr^k) substages of the Narva Stage; Viljandi (D_2ar vl), Kureküla (D_2ar kr) and Tarvastu (D_2ar tr) beds of the Aruküla Stage; Härma (D_2br hm) and Koorküla (D_2br kr) beds and Abava Substage (D_2br^{ab}) of the Burtnieki Stage; Sietiņi (D_2gj S) and Lode (D_2gj L) members of the Gauja Stage. 1 - sandstone; 2 - siltstone; 3 - clay; 4 - dolomitic marl; 5 - sedimentary breccia; 6 - dolomite; 7 - dolomitic matrix; 8 - fish remains; 9 - Quaternary deposits.

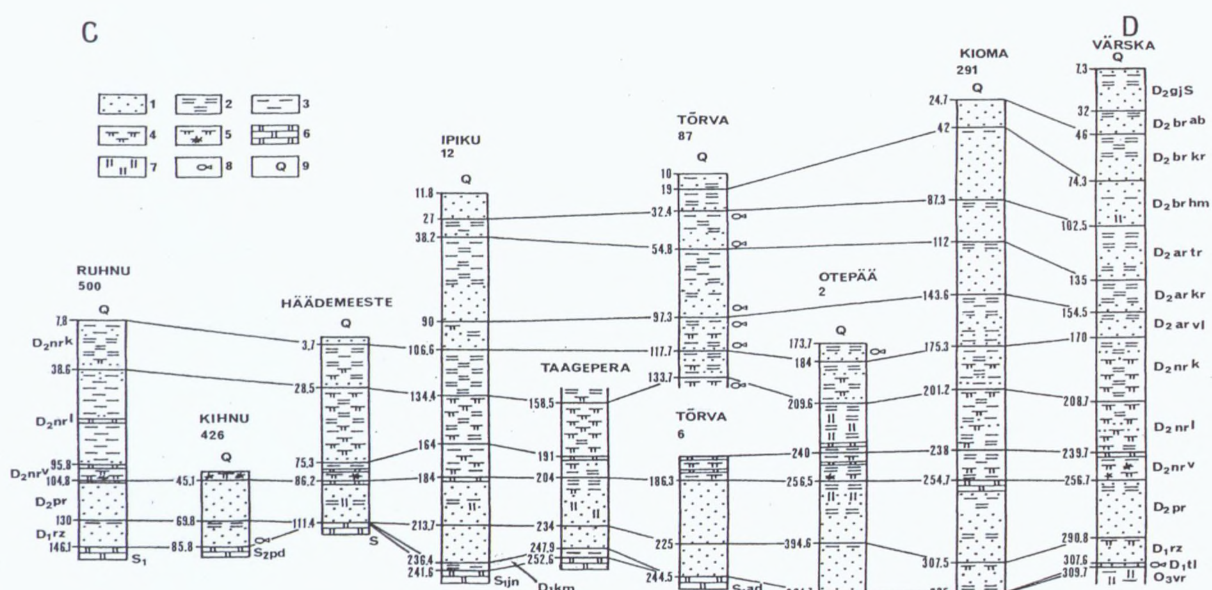


Fig. 76. Latitudinal stratigraphical cross-section (CD) comprising the whole Devonian sequence. For location of the cross-section see Fig. 80. Boreholes: Ruhnu - 257, Kihnu - 224, Häädemeeste - 229, Ipiku - 234, Taagepera - 236, Tõrva (two boreholes) - 237, Otepää - 249, Kioma - 250 and Värska - 275. Numbers under the name of borehole correspond its initial number. Separated are Vormsi (O_3vr) and Pirgu (O_3prg) stages; Silurian (S), Lower Silurian (S_1); Raikküla (S_1rk), Adavere (S_1ad), Jaani (S_1jn) and Paadla (S_2pd) stages; Tilžė (D_1tl), Ķemeri (D_1km), Rēzekne (D_1rz) and Pärnu (D_2pr) stages; Vadja (D_2nr^v), Leivu (D_2nr^l) and Kernavė (D_2nr^k) substages of the Narva Stage; Viljandi (D_2ar vl), Kureküla (D_2ar kr) and Tarvastu (D_2ar tr) beds of the Aruküla Stage; Härma (D_2br hm) and Koorküla (D_2br kr) beds of the Burtnieki Stage; Abava Substage (D_2br^{ab}) of the Burtnieki Stage and Sietiņi Member (D_2gj S) of the Gauja Stage. 1 - sandstone; 2 - siltstone; 3 - clay; 4 - dolomitic marl; 5 - sedimentary breccia; 6 - dolomite; 7 - dolomitic matrix; 8 - fish remains; 9 - Quaternary deposits.

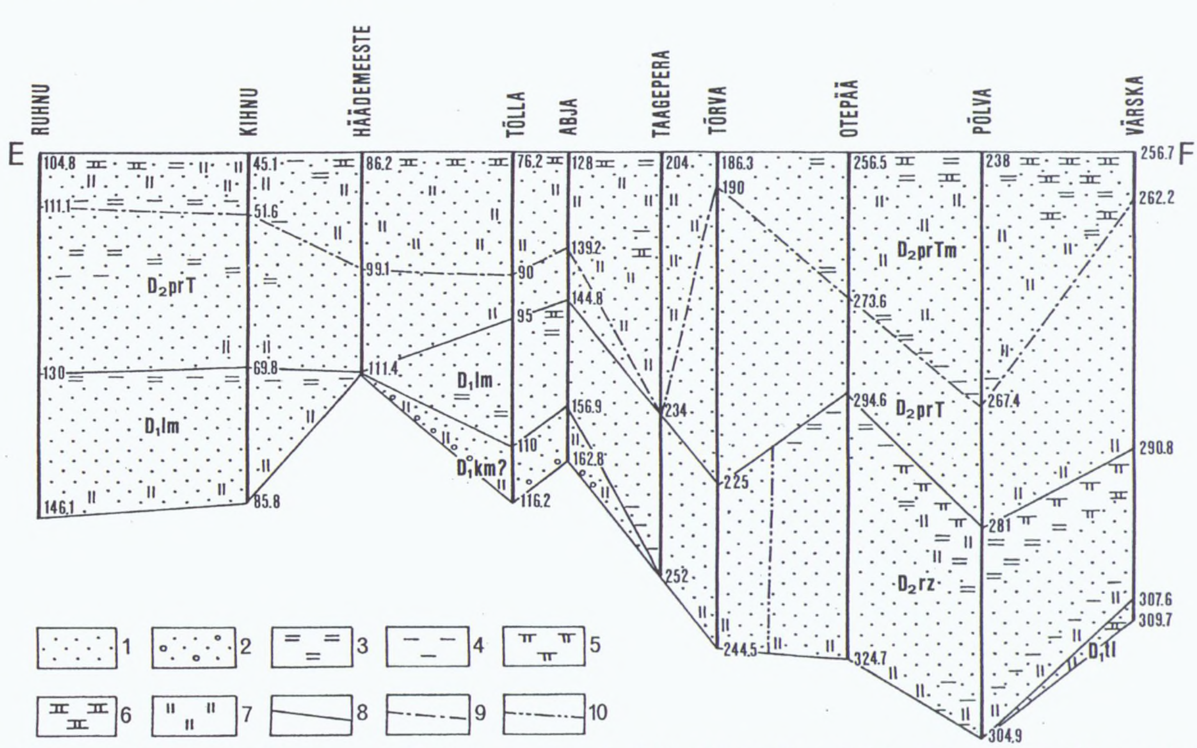


Fig. 77. Longitudinal stratigraphical cross-section (EF) comprising the Tilžė (D_1tl) and Ķemeri (D_1km) stages, Lemsi (D_1*lm*) and Rēzekne (D_1*rz*) formations of the Rēzekne Stage, and Tori (D_2prT) and Tamme (D_2prTm) members of the Pärnu Stage. For location of the section see Fig. 80. Boreholes: Ruhnu - 257, Kihnu - 224, Häädemeeste - 229, Tõlla - 232, Abja - 233, Taagepera - 236, Tõrva - 237, Otepää - 249, Põlva - 253 and Värska - 275. 1 - sandstone; 2 - conglomeratic sandstone; 3 - siltstone; 4 - clay; 5 - dolomitic marl; 6 - dolomite; 7 - dolomitic matrix; 8 - boundary between stages; 9 - boundary between the Rēzekne and Lemsi formations; 10 - boundary between the Tori and Tamme members.

the Ķemeri Formation (Table 10). In Estonia the formation consists of light-grey and pinkish poorly sorted horisontally thin-bedded sandstone and dolomite cemented with conglomeratic sandstone in the basal part. It includes interlayers of grey clay and dolomitic marl, seldom bluish-grey stabby siltstone.

Mineralogically, the rocks of the Ķemeri Formation are predominantly quartzose arenites with the quartz content reaching 80-98%. In the heavy fraction, ilmenite and transparent allothigenic minerals dominate, accounting for 17-57% and 37-53%, respectively. Among the latter group zircon is clearly prevailing, but garnet and tourmaline are also important (Fig. 78).

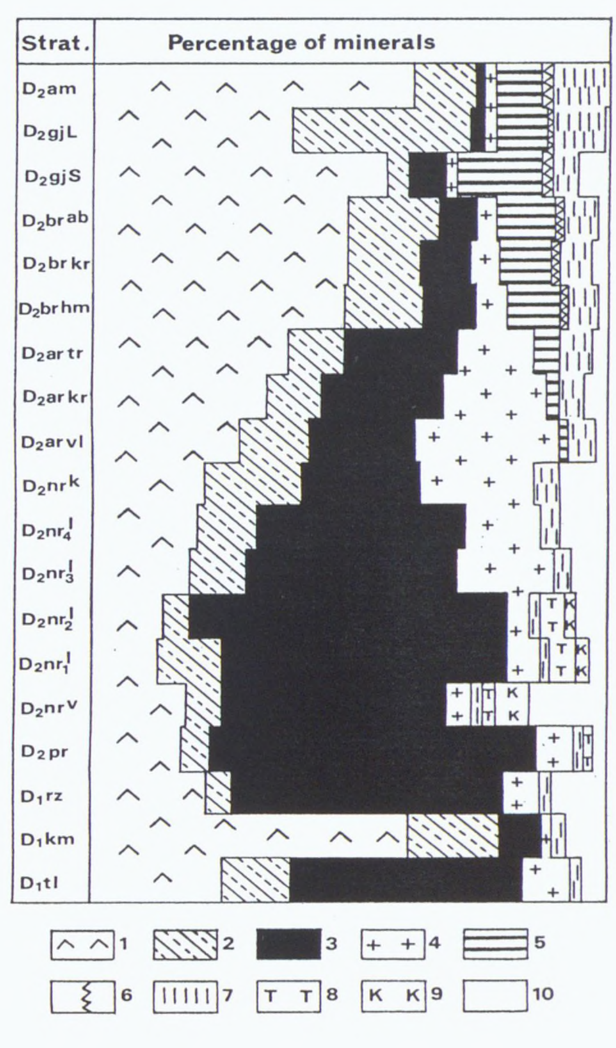


Fig. 78. Mineralogical changes of heavy detrital minerals in the vertical section from the Tilžė to Amata stages. Stratigraphical indices: D_1tl - Tilžė, D_1km - Ķemeri, D_1rz - Rēzekne and D_2pr - Pärnu stages; D_2nr^v - Vadja Substage; $D_2nr_1^l$ - bed one, $D_2nr_2^l$ - bed two, $D_2nr_3^l$ - bed three and $D_2nr_4^l$ bed four of the Leivu Substage; D_2nr^k - Ķernavė Substage; D_2ar vl - Viljandi, D_2ar kr - Kureküla and D_2ar tr - Tarvastu beds of the Aruküla Stage; D_2br hm - Härma and D_2br kr - Koorküla beds of the Burtnieki Stage; D_2br^{ab} - Abava Substage; D_2gjS - Sietini and D_2gjL - Lode members of the Gauja Stage; D_2am - Amata Stage. 1 - zircon; 2 - tourmaline; 3 - garnet; 4 - apatite; 5 - staurolite; 6 - kyanite; 7 - rutile; 8 - titanite; 9 - corundum; 10 - varia.

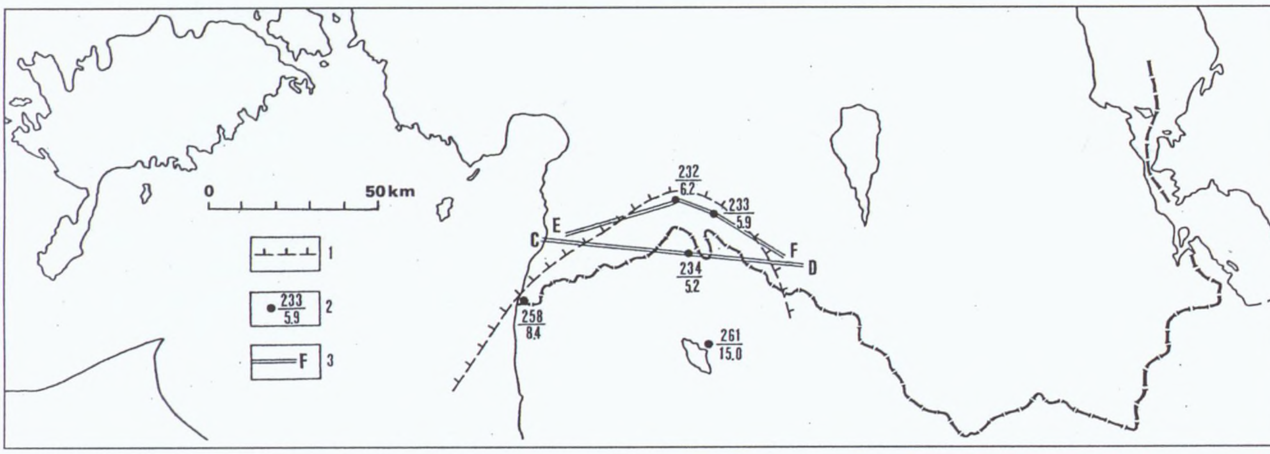


Fig. 79. Sketch-map showing the presumable present distribution and the thickness of the Ķemeri Stage in Estonia. CD and EF indicate part of transects of stratigraphical cross-sections presented in Figs. 76 and 77. Numerator at the borehole sign marks the number (see Fig. 3 for section names), denominator - the thickness of the stage in metres. 1 - contour of presumable rock distribution; 2 - borehole; 3 - transect of cross-section.

Rēzekne Stage

The Rēzekne Stage was established by Lyarskaya (Лярская 1974) with the stratotype in Akniste-5 drill core (interval 361.8 - 487.8 m) in southeastern Latvia. Earlier it was treated as the Kemeri (Лиепиньш 1952) or Viesite (Лиепиньш 1964) Formation. After destruction of Akniste-5 drill core, the interval of 427 - 446 m in Ludza-15 drill core in eastern Latvia was selected for the neostratotype of the Rēzekne Stage (Лярская 1978). In Estonia, the corresponding strata were earlier treated as the Pärnu Stage (Марк и Паасикиви 1960) and the Viesite Formation (Клеесмент 1966). As the Rēzekne Stage they were first mentioned in 1975 (Клеесмент и др. 1975). The age of the stage was palaeontologically determined by Mark-Kurik (1991a).

The Rēzekne Stage is spread in southern Estonia and covered with the rocks of the Pärnu Stage (Figs. 75, 76, 77). The best examined section is the Mehikoorma drill core (interval 220.3 - 246.2 m, Клеесмент и др. 1975). The total thickness of the stage varies from 0.7 m in the Asuküla and Kaagvere boreholes up to 51.5 m in the Kavastu borehole. The stage is at its thickest in eastern Estonia (Fig. 80).

The Rēzekne Stage is characterized by greenish-, purplish- and light-grey sandstone. In southeastern Estonia, the upper part of the section is represented by dolomitic marls. The stage lies with a stratigraphical unconformity on the different stages of Ordovician or Silurian age, in a few cases on the rocks of the Tilžė Stage (Värska, Laanemetsa, Väimela), in the Mõniste High it overlies the basement complex. The lower part of the section consists of sandstones with dolomitic matrix, often conglomeratic (Figs. 75, 76, 77).

The fossils known from the Rēzekne Stage are largely confined to the Rēzekne Formation. An equivalent of the latter unit, the Lemsi Formation contains a few unidentified fish remains (Сорокин 1981). Fossil fishes coming from the

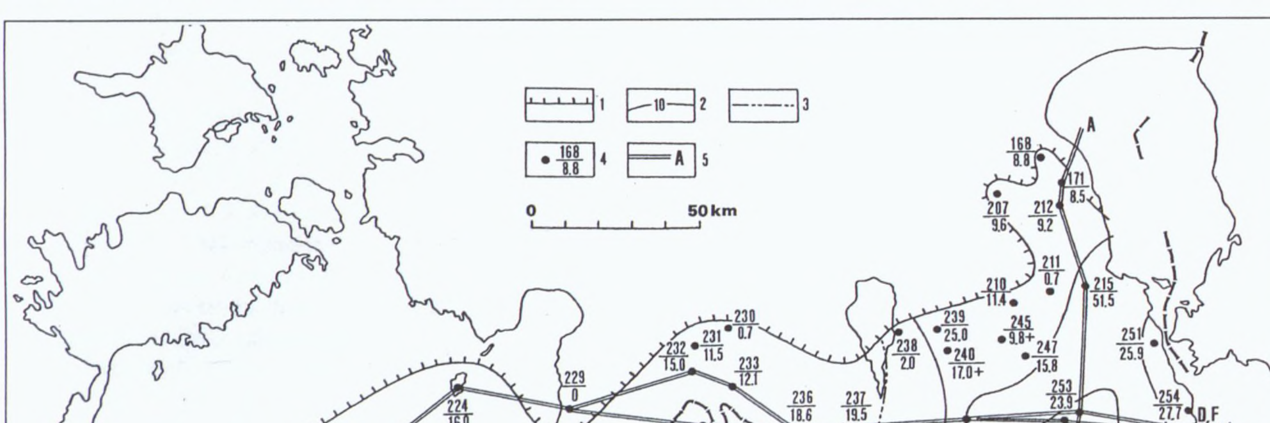


Fig. 80. Sketch-map showing the present distribution and thickness of the Rēzekne Stage in Estonia. AB, CD and EF indicate transects of stratigraphical cross-sections presented in Figs. 75, 76 and 77. Numerator at the borehole sign marks its number (see Fig.3 for section names), denominator - the thickness of the stage in metres. 1 - contour of rock distribution; 2 - isopach; 3 - boundary between the Rēzekne and Lemsi formations; 4 - borehole; 5 - transect of cross-section.

Rēzekne Formation include *Schizosteus* sp., Psammosteidae gen. indet., Cephalaspidida gen. indet., *Kartalaspis belarussica* Mark-Kurik in litt., Antiarchi gen. indet., *Laliacanthus singularis* Kar.-Tal., *Diplacanthus kleesmentae* Valiuk., *Acanthodes ?* sp. B Valiuk., *Acanthodes*? sp. C Valiuk., *Ptychodictyon ancestralis* Valiuk., *Cheiracanthus gibbosus* Valiuk., *Markacanthus parallelus* Valiuk., *Ectopacanthus flabellatus* Valiuk., *Rhadinacanthus primaris* Valiuk., *Nostolepis* sp., *Pruemolepis wellsi* Vieth-Schreiner, Crossopterygii gen. indet. The presence of otoliths is noteworthy.

Invertebrates comprise "*Lingula*" sp., Ostracoda, *Glyptasmussia*? sp., Gastropoda. Microfossils include a simple conodont and miospores: *Retusotriletes simplex* Naumova, *Leiotriletes microrugosus* (Ibr.) Naumova, *L. simplex* Naumova [rz_1], *Stenozonotriletes conformis* Naumova [rz_1], *Acanthotriletes perpusillus* Naumova [rz_1], *A. parvispinosus* Naumova [rz_1], *Archaeozonotriletes memorabilis* V. Umnova [rz_1], *Emphanisporites rotatus* McGregor [rz_1], *Dibolisporites* cf. *eifeliensis* (Lanninger) McGregor [rz_1], *Diatomozonotriletes devonicus* Naumova [rz_1], *Retusotriletes* cf. *priscus* V. Umnova [rz_2], *Leiotriletes* cf. *insuetus* V. Umnova [rz_2], *Granulatisporites* cf. *rudigranulatus* Staplin [rz_2], cf. *Calamospora pannucea* Richardson [rz_2]. The above list shows that the miospore content differs in the lower [rz_1] and upper [rz_2] parts of the Rēzekne Formation. The lists of fossils are by Kleesment *et al.* 1975 (Клеесмент и др. 1975) and Valiukevičius 1994 (miospores identified on generic level are not indicated).

In Estonia, the Rēzekne Stage consists of two formations, laterally replacing each other. In eastern Estonia, the Rēzekne Formation expands as far as Lake Võrtsjärv, west of it the Lemsi Formation occurs (Сорокин 1981, Figs. 77, 80).

The interval of 220.3 - 246.2 m of the Mehikoorma drill core has been selected as the parastratotype of the Rēzekne Formation. The thickness of the formation varies commonly from 10 to 30 m (Fig. 80). The lower part of the section is represented by grey, purplish- and pinkish-grey loose sandstone, with interlayers of brownish-black silty platy clay in its basal part. On contact with the underlying Ordovician or Silurian carbonate rocks (3-5 m) the sandstones are strongly cemented with dolomitic matrix. In southeastern Estonia, the upper part of the Rēzekne Formation is represented by grey silty dolomitic marl up to 10 m in thickness, in other regions by an up-to-1–2-m-thick layer of grey siltstone or silty sandstone (Figs. 75,76,77). The sandstone of the Rēzekne Formation is fine- and very fine-grained.

Mineralogically, the sandstone of the Rēzekne Formation is predominantly feldspatic arenite with the quartz content of 70-85%. The sand component of the dolomitic marl contains 60-75% of quartz. The heavy fraction is dominated by allothigenic transparent minerals. In sandstone its share is commonly 45-60% and in dolomitic marl it forms 30-50% of the fraction. This group is dominated by garnet (50-70%), accompanied by zircon (15-25%, Fig. 78). The content of garnet is relatively high in dolomitic marls.

The stratotype of the Lemsi Formation is the interval of 69.8 - 85.8 m of the Kihnu drill core (Сорокин 1981). The thickness of the formation is commonly 11-20 m (Fig. 77). It mostly consists of light grey, yellowish and brownish, most rarely of purplish loose sandstone, which in the basal contact part with Silurian carbonate rocks is strongly cemented with dolomitic matrix. The upper 0.6 - 2.9 m of the section consist of greenish and purplish-grey siltstones or very fine-grained sandstone, often strongly cemented with dolomitic matrix. The sandstone of the Lemsi Formation is predominantly fine- and medimum grained.

Mineralogically, the sandstones of the Lemsi and Rēzekne formations are similar. Only in the sandstone of the Lemsi Formation the content of zircon is higher, particularly in the Kanaküla - Tõlla - Ipiku area where it dominates over the garnet.

MIDDLE DEVONIAN

The Middle Devonian is the completest part of the Devonian section in Estonia with both the Eifelian (Pärnu, Narva, Aruküla) and Givetian (Burtnieki, Gauja, Amata) standard stages being present (Mark-Kurik 1993c). The total thickness of the Middle Devonian rocks reaches 400 m. The wide outcrop area contains excellent exposures (Figs. 81, 82, 84, 86, 87, 88).

Pärnu Stage

The Pärnu strata were established as an independent stratigraphical unit by Orviku (1930c,1932). The name "Pärnu" ("Pernu") was first used by Obruchev (Обручев 1933). Palaeontologically, it was distinguished as the *Schizosteus heterolepis* Zone by Gross (1942) and as a stage by Mark-Kurik (Марк 1958). The stratotype is the bank of the Pärnu River near the settlement of Tori. The exposures occur on the banks of the Pärnu and Navesti rivers in central and southwestern Estonia, including Oore dairy - the boundary outcrop with the Narva Stage (Fig. 81).

The Pärnu Stage is spread in southern Estonia. The outcrop forms two narrow wedgeform areas in the northwestern and northeastern parts of the distribution area. The total thickness of the stage ranges commonly from 15 to 47 m (Figs. 75, 76, 81). In the Võrtsjärv Depression, only the topmost part of the section, up to 8 m in thickness, is represented.

The Pärnu Stage is characterized by light-yellow fine-grained cross-bedded sandstone. In most of the distribution area it lies conformably on the Rēzekne Stage. The topmost layer of the Rēzekne Stage in the southeastern part of the area is represented by dolomitic marl which is overlain by sandstone of the Pärnu Stage. In the western part, the boundary between these stages is difficult to establish because of their similar composition. In the northern part of the of distribution area, the Pärnu Stage lies with a stratigraphical disconformity on the Silurian and Ordovician carbonate rocks.

The majority of the fossils of the Pärnu Stage are confined to the Tori Member. The Tamme Member has revealed gyrogonites (?) of charophyte algae and, probably, unidentified lamellibranchs and rare fish remains (Orviku 1930c). In the Tamme Member Valiukevičius (pers. comm.) has identified scales of acanthodians *Cheiracanthus gibbosus* Valiuk., *Rhadinacanthus primaris* Valiuk., *Cheiracanthus brevicostatus* Gross and *Acanthodes?* sp. D. Fossil fishes occurring in the Tori Member are: *Schizosteus heterolepis* (Preob.), *Psammolepis toriensis* (Mark-Kurik), *Tartuosteus*

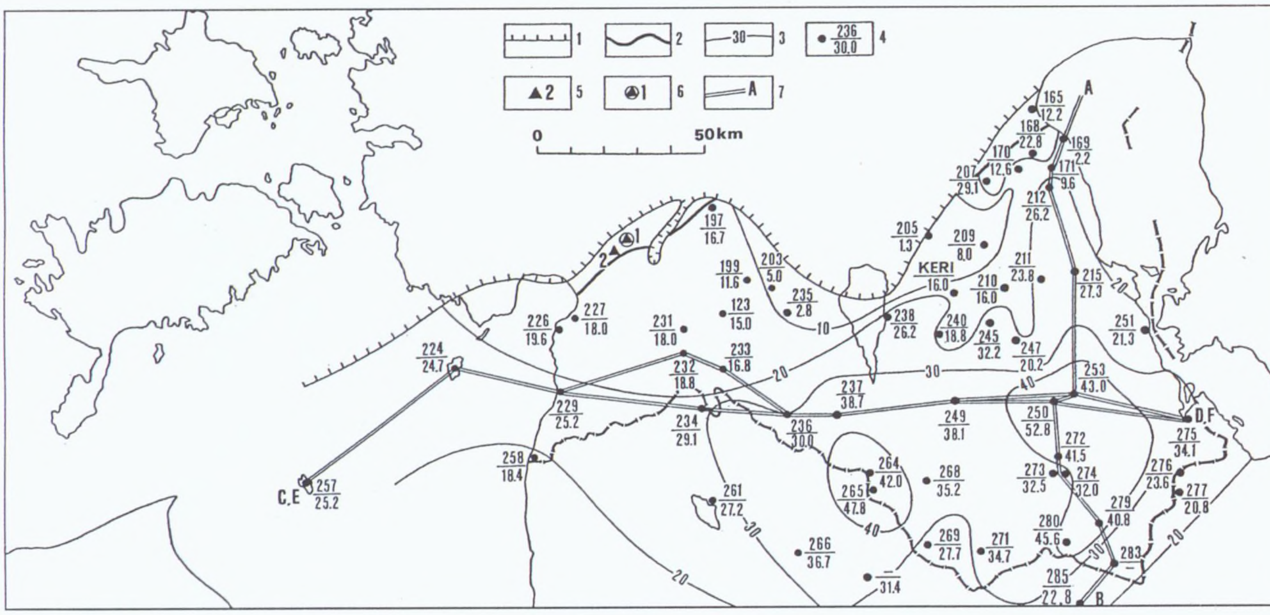


Fig. 81. Sketch-map showing the present distribution and the thickness of the Pärnu Stage in Estonia. AB, CD and EF indicate transects of stratigraphical cross-sections presented in Figs. 75, 76 and 77. Numerator at the borehole sign marks its number (see Fig.3 for section names), denominator - the thickness of the stage in metres. 1 - contour of rock distribution; 2 - limit of outcrop belt; 3 - isopach; 4 - borehole; 5 - outcrop (2 - Oore dairy); 6 - type section (1 - Tori); 7 - transect of cross-section.

sp., *Actinolepis tuberculata* Ag., *Homostius* sp., *Byssacanthus dilatatus* (Eichw.), *Archaeacanthus quadrisulcatus* Kade, *Diplacanthus kleesmentae* Valiuk., *Acanthodes* sp. B? Valiuk., *Porolepis* sp., *Glyptolepis* sp., Osteolepididae, Dipnoi?. Invertebrates (lingulates) are extremely rare. Common is fossil flora including macroremains of *Hostinella* sp., and *Psilophytites* sp., and miospores: *Periplecotriletes tortus* Egorova, *Emphanisporites rotatus* McGregor, *Retusotriletes raisae* Tchib., *R. devonicus* Naumova, *R. concinnus* Kedo, *R. incomptus* Kedo, R. *planituberculatus* Kedo, *Dibolisporites antiquus* (Kedo) Arkh., *Hymenozonotriletes marginodentatus* Kedo, *H. altus* Kedo, *H. ludzus* Kedo, *H. longus* Arkh., *Calyptosporites velatus* (Eisenack) Richardson, *C. tener* (Tchib.) Obukh. var. *concinnus* Tchib., *Camarozonotriletes apertus* Kedo, *Sinuosisporites sinuosus* (V. Umnova) Arkh., *Punctatisporites tortuosus* (Tchib.) Arkh. (data from Сорокин 1981, Валюкевичюс и др. 1986, modified according to Abukhovskaya (pers. comm.)).

In Estonia and adjacent areas, the Pärnu Stage is represented by the Pärnu Formation. In Estonia, the formation (Table 10) is divided into the Tori (below) and Tamme (above) members.

The Tori Member is dominantly represented by yellow, light-grey or purplish-grey loose cross-bedded sandstone. Strongly cemented sandstone with dolomitic matrix forms only a basal layer with a thickness of 0.03 to 2 m on the Silurian or Ordovician carbonate rocks. Commonly, it contains pebbles of the underlying sediments. The sandstone is dominantly fine-grained, in the bottommost part medium-grained sandstone is developed. The thickness of the Tori Member varies greatly and irregularly and is highest in the Tsiistre drill core. In some places it is absent (Taagepera drill core, Fig. 77).

The Tamme Member is represented by interbedding loose and dolomitic-cemented greenish-, pinkish- and purplish-grey sandstone containing thin interlayers of siltstone and clay. The complex is horizontally-bedded. Commonly, yellowish-grey sandy dolomite (dolostone) with a thickness of 0.5 to 1 m occurs in the topmost part of the section. Strongly cemented sandstones with dolomitic matrix contain irregular vugs with a diameter of 1 to 15 cm, fulfilled with loose sandstone. The sandstone of the Tamme Member is fine and very-fine grained. The thickness of the member varies irregularly from 2 m (Valga-324 borehole) to 30 m, which is the full thickness of the Pärnu Formation (Taagepera borehole, Fig.77).

Mineralogically, the sandstone of the Pärnu Formation is quartzose and feldspatic arenite with the quartz content of 75-85%. The heavy fraction is dominated by transparent allothigenic minerals (about 50%), among which garnet with a considerable supplement of zircon is prevailing (Fig.78). The sandstones of the Tori and Tamme members are quite similar; only in the Tamme Member the content of garnet is higher and that of zircon is lower.

Narva Stage

As an independent stratigraphic unit the Narva Stage was distinguished and termed by Obruchev (Обручев 1933). The left bank of the Narva River near Gorodenka, and the banks of the Gorodenka Brook and the Poruni River near the place where these watercourses flow into the Narva River in north-eastern Estonia, make up the stratotype area of the stage. The outcrops are concentrated in northeastern Estonia, with the Narva and Sirgala quarry sections (Fig. 82) being most noteworthy.

The Narva Stage is spread in southern and eastern Estonia. The outcrop area extends as a 10–30-km-wide belt from Ruhnu to Halliku. Besides, there is a separate area in north-eastern Estonia and a few isolated spots near the outcrop line. The total thickness ranges from 30 to 109 m, increasing from north to south (Fig. 82).

The lower boundary of the stage coincides with the base of a carbonate breccia or dolomitic marl, 0.2 to 10 m in thickness, which overlie sandy dolomite or sandstone of the Pärnu Stage (Figs. 75,76).

On the basis of palaeontological (acanthodians,

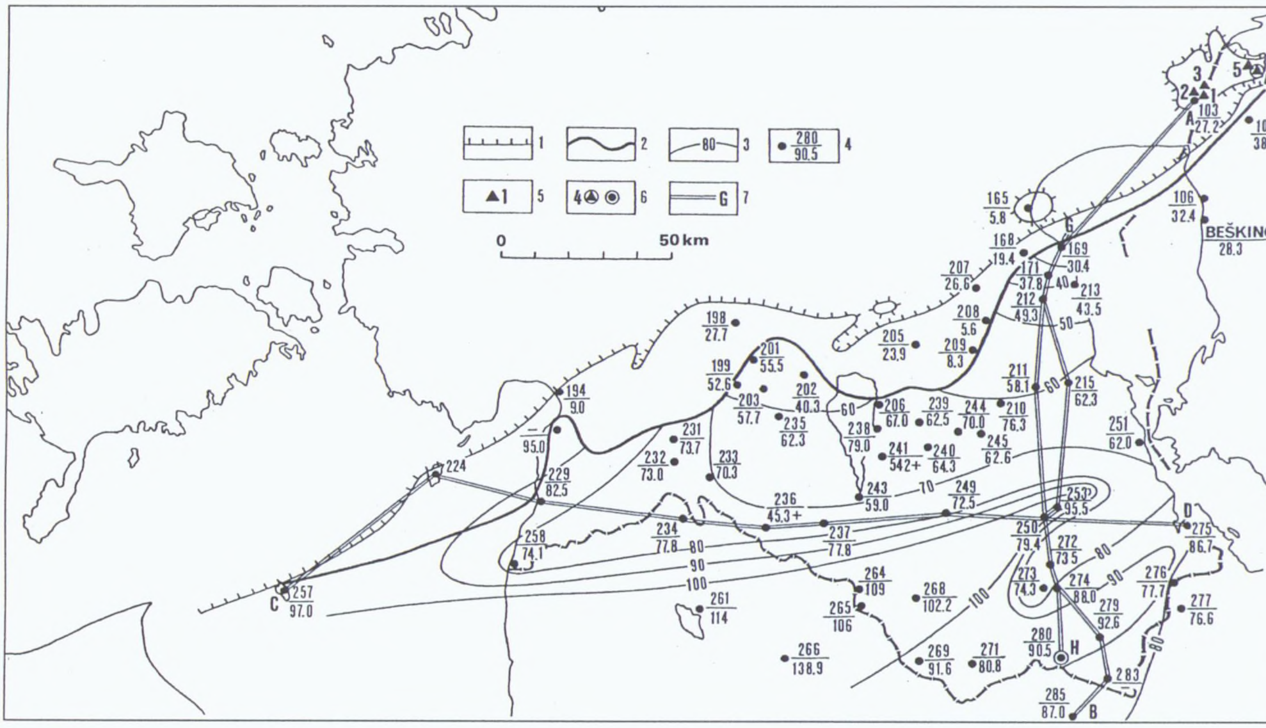


Fig. 82. Sketch-map showing the present distribution and the thickness of the Narva Stage in Estonia. AB, CD and GH indicate transects of stratigraphical cross-sections presented in Figs. 75, 76 and 83. Numerator at the borehole sign marks its number (see Fig. 3 for section names), denominator - the thickness of the stage in metres. 1- contour of rock distribution; 2 - limit of outcrop belt; 3 - isopach; 4 - borehole; 5 - outcrop (1 - Narva River; 2 - Gorodenka Brook; 3 - Poruni River; 5 - Sirgala quarry); 6 - type section (4 - Narva quarry); 7 - transect of cross-section.

inarticulates, brachiopodes, spores), lithological and mineralogical characteristics, the Narva Stage has been divided into three substages (Table 10) traceable from the stratotype area up to eastern Belarus (Валюкевичюс и др. 1986, Клеесмент и др. 1987).

The Narva oil shale quarry section, 10 km southeast of the Sirgala Settlement, has been selected for the stratotype of the lower, Vadja Substage. Its parastratotype is the outcrop on the left bank of the Narva River, 300 m downstream from the mouth of the Gorodenka Brook. The thickness of the substage in the stratotype area is about 16 m, in Estonia it varies from 10 to 31.9 m (Figs. 75, 76, 83).

The Luutsniku-451 drill core in the interval of 317-377.7 m has been selected as a stratotype for the Leivu (middle) Substage. In the stratotype area it is exposed at the Poruni and Narva rivers. The thickness of the substage in northeastern Estonia is about 5 m and it increases considerably in a southern direction reaching 60.7 m in the Luutsniku drill core (Figs. 75, 76, 83).

For the stratotype of the upper, Kernavė Substage the interval of the Ledai borehole in central Lithuania was proposed.

The Narva Stage is rich in fossil fishes, particularly its upper part - the Kernavė Substage [k] where the majority of the macroremains come from. In the Vadja Substage [v] and especially in the Leivu Substage [l] the fishes are more scarce. The fish fauna of the stage consists of *Schizosteus striatus* (Gross) [k], *Pycnolepis splendens* (Eichw.) [l?,k], *Cephalaspidida, Actinolepis tuberculata* Ag. [k], *Holonema* sp. A Mark-Kurik [v], *Holonema* sp. B Mark-Kurik [k], *Homostius latus* Asm. [k], *Coccosteus cuspidatus* Miller ex Ag. [k], *Protitanichthys*? sp.n. Mark-Kurik [v], *Byssacanthus dilatatus* (Eichw.), *Asterolepis estonica* Gross [k], *Cheiracanthoides estonicus* Valiuk. [v], *Acanthodes*? sp.C [v], *Cheiracanthus crassus* Valiuk. [v], *Rhadinacanthus balticus* Gross, *Acanthodes*? sp.B, *Acanthodes*? sp.D, *Cheiracanthus brevicostatus* Gross, *C. longicostatus* Gross, *Ptychodictyon distinctum* Valiuk. [l,k], *P. rimosum* Gross [l, k], *Cheiracanthus* sp. [l,k], *Diplacanthus* sp. [l,k], *Acanthodes* ? sp.A. [l,k], *Cheiracanthus intricatus* Valiuk. [k], *Nostolepis kernavensis* Valiuk. [k], *Cheiracanthoides proprius* Valiuk. [k], *Markacanthus costulatus* Valiuk. [k], *Minioracanthus laevis* Valiuk. [k], *Ptychodictyon sulcatum* Gross [k], *Diplacanthus carinatus* Gross [k], *Acanthodii* gen.n. Valiuk. [k], *Archaeacanthus quadrisulcatus* Kade, *Haplacanthus marginalis* Ag., *Homocanthus gracilis* (Eichw.), *Thursius fischeri* (Eichw.) [k], Osteolepididae, *Glyptolepis quadrata* Eichw. [k], *Glyptolepis* sp., *Onychodus* sp.[v,k], *Dipterus serratus* (Eichw.)[k], *Dipnoi* [v, l], *Cheirolepis* sp.sp., *Orvikuina* sp. [v, l], *O. vardiaensis* Gross [k].

Invertebrates are represented by lingulate brachiopods (*Bicarinatina sakalana* Rõõmusoks *et* Gravitis a.o.) which are especially numerous in the upper portion of the Leivu Substage. Ostracods (*Lepertitia tartuensis* Öpik var. *geographica* Hecker, Ostracoda inc. gen.) are less common and so are unidentified lamellibranchs in the Kernavė Substage. Conchostracans *Pseudestheria pogrebovi* Lutk., *Trigonestheria triangularis* Mir., *Glyptoasmussia quadrata* Mir., *G.* aff. *willweratica* Nov., *Concherisma eifelense* Raym., *Ulugkemia sinuata* Lutk., *U. mesodevonica* Mir., *Asmussia membranacea* Pacht, *Praeleaia quadricarinata* Lutk. are characteristic for the Vadja Substage and the lowermost part of the Leivu Substage. The lists of animal fossils are given after Sorokin *et al.* (Сорокин 1981), Valiukevičius *et al.* (Валюкевичюс и др. 1986) and Valiukevičius (1994).

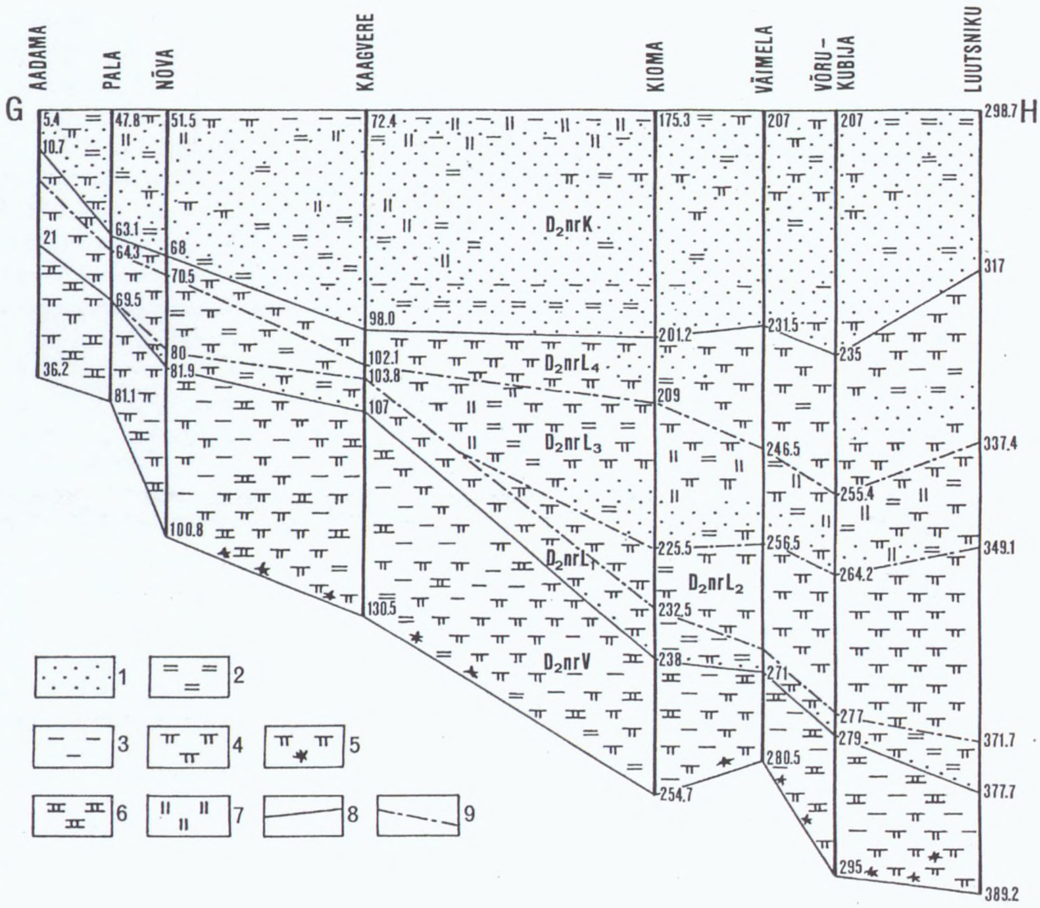


Fig. 83. Stratigraphical cross-section (GH) of Narva Stage from Aadama (north) to Luutsniku (south). For location of the section see Fig. 82. Boreholes: Aadama - 169, Pala - 171, Nõva - 212, Kaagvere - 211, Kioma - 250, Väimela - 272, Võru-Kubija - 274, Luutsniku - 280. Stratigraphical indices: D_2nrV - Vadja Member; D_2nrL_1, D_2nrL_2, D_2nrL_3 and D_2nrL_4 - beds of Leivu Member; D_2nrK - Kernavė Member. 1 - sandstone; 2 - siltstone; 3 - clay; 4 - dolomitic marl; 5 - sedimentary breccia; 6 - dolomite; 7 - dolomitic matrix; 8 - boundary between members; 9 - boundary between the beds of the Leivu Member.

Microspore assemblage from the Narva Stage is restricted to the Vadja Substage. It includes *Retusotriletes raisae* Tchib., *R. devonicus* Naumova, *R. concinnus* Kedo, *R. incomptus* Kedo, *R. planituberculatus* Kedo, *R.* cf. *brandtii* Streel, *R. fragosus* Arkh., *R. microsetosus* Kedo, *R. lanceolatus* Kedo, *R. luxispinus* Kedo, *R. engurensis* Kedo, *R. clivosiformis* Kedo, *Hymenozonotrietes* cf. *marginodentatus* Kedo, *H. altus* Kedo, *H.* cf. *echiniformis* Kedo, *H. ludzus* Kedo, *Camarozonotriletes apertus* Kedo, *Grandispora protea* (Naumova) Moreau-Benoit, *Phylotecotriletes triangulatus* Tiw. et Schaarschmidt. The Vadja Substage has also revealed acritarchs, not yet identified (data from Валюкевичюс и др. 1986, modified). Gyrognites (?) of charophyte algae (*Sycidium*) have been found from all members of the Narva Stage. The Kernavė Substage comprises poorly preserved plant macroremains.

In Estonia and adjacent regions, the Narva Stage is represented by the Narva Formation with a highly variable lithology. Its lower part consists mostly of dolomitic marl with interlayers of dolomite and dolomitic clay. Siltstone, very fine-grained and silty sandstone intercalating with interlayers of dolomitic marl and clay, forms the upper part of the Narva sequence. The sequence of the formation is divided into three members corresponding in volume to substages (Table 10, Fig. 83).

In the basal part of the lower, Vadja Member the breccia of dolomitic marl with unsorted irregular pebbles of dolomite, dolomitic marl and siltstone are common. In general this member is characterized by a thin-bedded complex of dolomitic marl, dark-grey to black silty clay and pale yellowish-grey dolomite which often includes crystalline dolomite, chalcedony, pyrite or sphalerite filled vugs. The detrital component of the rocks in the lower part of the unit is mineralogically relatively mature with the quartz content reaching 50-80%. In the upper part, the rocks are often micaceous containing quartz 20-50, feldspars 15-30 and micas 20-60%. The heavy mineral suite is prevailed by Fe hydroxides or pyrite, in some interlayers baryte is dominating. Nonopaque heavy minerals are dominated by garnet, followed by zircon (Fig. 78).

The middle, Leivu Member is prevailed by dolomitic marl. The section varies both lithologically and in thickness. Within the section four beds have been distinguished (Валюкевичюс и др. 1986, Kleesment 1995). Two lower beds are thinning towards the north-east (Fig.83). The lowermost bed of grey dolomitic marl contains a remarkable amount of silty-sandy particles with a diameter up to 1-2 mm. The next bed is a grey thin-bedded complex formed of intercalating dolomitic marl, dolomite and dolomitic clay. The third bed from the bottom comprises interlayers of grey silt- and sandstone. The topmost bed consists of reddish-brown, purplish-grey and grey mottled massive argillaceous dolomitic marl which serves as a good correlative level. Mineralogically, the detrital part of the rocks in the lowest bed belongs to the feldspatic arenites, in other beds – to the arcosic and micaceous arenites. The heavy fraction is dominated by Fe hydroxides, more rarely by pyrite or micas. The beds differ from one another by the nonopaque heavy mineral suite. In the two lower beds garnet is clearly dominating, while in the overlying beds it is accompanied by apatite, zircon and tourmaline. Significant is the presence of sphen (titanite) and corund in the two lower beds (Fig. 78).

The upper, Kernavė Member consists of brownish-red and grey loose and dolomite-cemented silty sandstone with intercalations of siltstone, dolomitic marl and clay. The complex is horisontal-, lenticular-, more rarely cross-bedded. Mineralogically, the rocks of the Kernavė Member belong to the arcosic arenites with the quartz content of 50-80%. In heavy fraction micas (30-60%), more rarely Fe hydroxides (up to 92%) are prevailing. The heavy nonopaque mineral suite is mostly dominated by apatite, followed by tourmaline, zircon and garnet (Fig. 78).

Aruküla Stage

The Aruküla strata were distinguished as an independent stratigraphical unit by Gross (1940,1942) and transfered to the rank of regional stage by Mark-Kurik (Марк 1958). The stratotype is the river bank near the Tartu Jaani Cemetery (Photo 24); not far from it are the Aruküla caves (Photo 13). The other important localities are Kallaste, Viljandi, Tamme and Õisu (Fig. 84). The rich material collected there during more than a hundred years contains the majority of the fauna characteristic of this unit.

The Aruküla Stage is spread in southern and southeastern Estonia. The outcrop area forms a 17– 55-km-wide belt extending from Ruhnu Island and Ikla Settlement in the west to Pala and Mehikoorma settlements in the east (Fig. 84). The total thickness of the stage in Estonia ranges from 66.0 to 97.2 m.

The Aruküla Stage consists of reddish-brown cross-bedded sandstone, interbedded with siltstone, clay, and dolomitic marlstone. It lies everywhere above the Narva Stage. The lower boundary of the Aruküla Stage coincides, in general, with the lower surface of the first significant uncemented reddish-brown sandstone layer above the dolomitic siltstone or marl of the Narva Stage (Figs. 75, 76). The topmost part of the Narva Stage often shows a greenish-grey siltstone layer; the overlying Aruküla sandstone is mostly inequigranular. In the Võrtsjärv Depression, this boundary is often difficult to establish (Марк и Тамме 1964).

The Aruküla Stage is rich in fossil fishes known from its all three subdivisions (Table 10). Characteristic are psammosteid heterostracans and several placoderms, both arthrodires and antiarchs. Fishes from the Viljandi [vl] and Kureküla [kr] beds are better known than those from the Tarvastu beds [tr]. The Aruküla fish fauna includes: *Tartuosteus giganteus* Gross, *Pycnosteus palaeformis* Preobr. [vl], *Ganosteus artus* Mark-Kurik, *Psammolepis proia* Mark-Kurik [vl, kr], cephalaspidids [vl], *Actinolepis tuberculata* Ag. [vl,kr], *Holonema obrutshevi* Mark [vl], *Homostius latus* Asm., *Heterostius ingens* Asm., *Coccosteus grossi* O.Obr. [vl,kr], *Millerosteus?* sp [tr], *Byssacanthus dilatatus* (Eichw.) [vl,kr], *Asterolepis estonica* Gross [vl,kr], *Archaeacanthus quadrisulcatus* Kade, *Haplacanthus marginalis* Ag., *Homacanthus gracilis* (Eichw.), *Rhadinacanthus balticus* Gross [vl], *R. multisulcatus* Valiuk., *Diplacanthus* sp. [vl], *D. carinatus* Gross, *D. gravis* Valiuk., *Markacanthus costulatus* Valiuk. [vl], *Minioracanthus laevis* Valiuk. [vl,kr], *M. alius* Valiuk., *Acanthodes?* sp. A, *Acanthodes?* sp.B, *Acanthodes?* sp. D, *Cheiracanthus brevicostatus* Gross, *C. longicostatus* Gross, *Ptychdictyon rimosum* Gross, *P. sulcatum* Gross, *Gyroptychius pauli* Vorob., *Glyptolepis* sp. sp., *Dipterus* sp. sp., *Conchodus* sp. [kr], *Orvikuina* sp.n., *Cheirolepis* sp., *Tartuosteus? luhai* Mark-Kurik [kr, tr], *Pycnosteus pauli* Mark [kr,tr], *Nodocosta pauli* Gross [kr], *Thursius estonicus* Vorob. [kr], *Hybosteus* sp. [kr], *Tartuosteus maximus?* Mark-Kurik [tr], *Nostolepis* sp.n. Valiuk. [tr], *Ptychodictyon distinctum* Valiuk. [tr], *Porolepis?* sp. [tr].

Invertebrates of Aruküla Age are mostly known from the Viljandi beds. These are: ostracods *Leperditia tartuensis* Öpik, *Aparchitellina taehtverensis* (Öpik), *Drepanellina orvikui* (Öpik), *Pontocypris rulescens* (Öpik), lingulates *Bicarinatina bicarinata* (Kut.), *B. ugalana* Rõõmusoks *et* Gravitis and unidentified lamellibranchs. Trace fossils are fairly numerous in the Kureküla beds. Lingulate brachiopods occur also in the Tarvastu beds.

Plant remains consist of gyrogonites? of charophyte algae (Viljandi beds) and some poorly preserved fragmental branches. The list of fossils is from Sorokin (Сорокин 1981) and Valiukevičius (1994, modified).

In Estonia, the Aruküla Stage is represented by the Aruküla Formation. On the basis of lithological and mineralogical evidence, three cyclic units are distinguished in the Aruküla Formation. These cycles occur in all sections of Estonia and adjacent areas and are defined as the Viljandi (lower), Kureküla (middle) and Tarvastu (upper) beds of the Aruküla Formation (Table 10). Each unit begins with relatively coarse and poorly sorted sandstones of a mature mineral composition, but ends with a clayey-silty complex (Figs. 75, 76, 85, Kleesment, 1994) .

The lower, Viljandi beds are dominated by very fine sandstones, often platy or slaty-bedded. The Kureküla beds are characterized by irregularly cemented interbeds of variegated siltstones, pockets of white sandstone, lenses of conglomeratic sandstone and interlayers with large clay pebbles. The section of the Tarvastu beds contains typically conglomeratic interbeds and surfaces and crusts of Fe hydroxide.

Mineralogically, the rocks of the Aruküla Formation are predominantly quartzose and feldspatic arenite with the quartz content of 60-90%. Micaceous arenites (content of micas 20-50%) occur as thin interbeds. The heavy fraction is dominated by ilmenite (30-60%) and transparent allothigenic minerals (15-40%). Among the latter, garnet and zircon are most significant. Tourmaline and rutile are also important. On the lower boundary of the formation the content of zircon and apatite increases significantly, and staurolite appears (Fig. 78).

Burtnieki Stage

As an independent stratigraphical unit the Burtnieki strata was distinguished by Gross (1940,1942). Into the rank of regional stage it was raised by Mark-Kurik (Марк 1958). The stratotype is the bank of the Salaca River, 12 km northwest of Lake Burtnieki in northern Latvia. In Estonia, main exposures are situated at Helme and on the banks of the Ahja (Photo 25) and Võhandu rivers. The outcrops of Karksi, Härma, Koorküla and Essi are known as localities of fossil fishes (Fig. 86).

The Burtnieki Stage is spread in southeastern Estonia. The outcrop area forms a 25–50-km-wide belt stretching from Ipiku and Valga in the west to Mehikoorma and Karisilla in the east. The total thickness ranges from 60.6 to 94.5 m (Fig. 86).

The Burtnieki Stage is mainly represented by light (white, yellowish, pinkish and greyish-brown) fine-grained medium- to weakly-cemented cross-bedded sandstones with interlayers of siltstone and clay. The stage lies everywhere above the

Photo 24. Outcrop near the Jaani Cemetery in Tartu is the stratotype of the Aruküla Stage. *Photo by H. Viiding.*

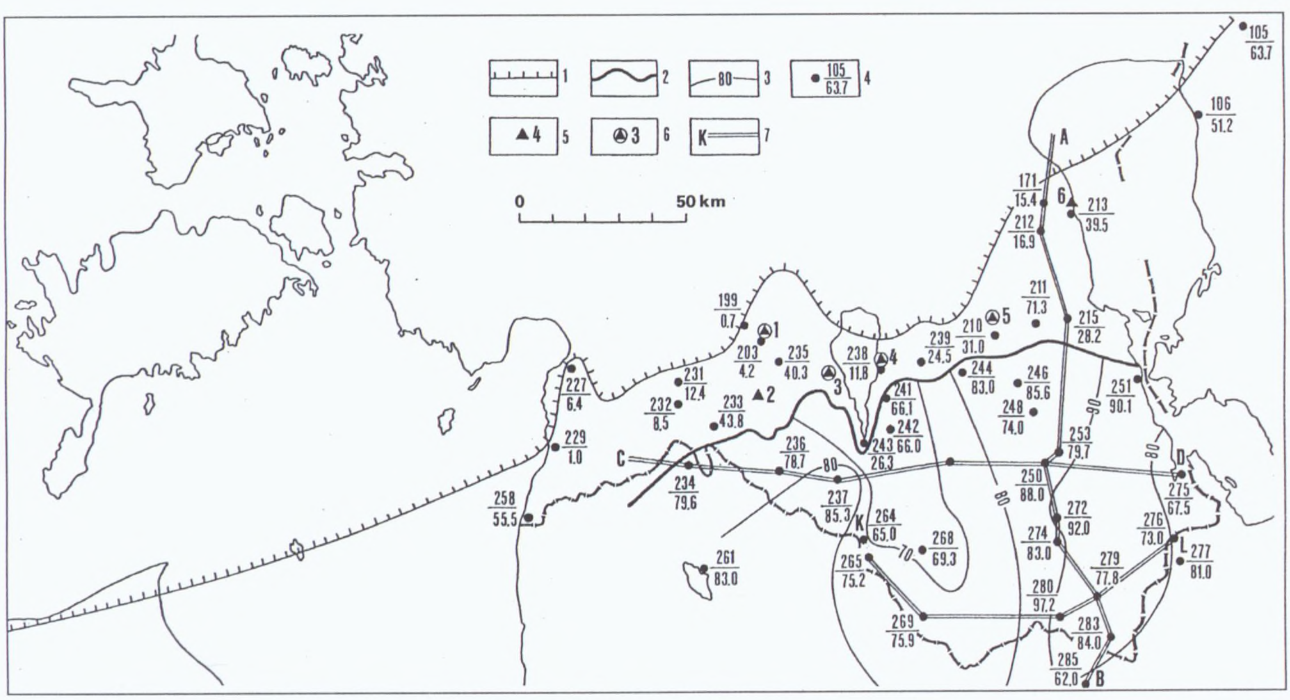


Fig. 84. Sketch-map showing the present distribution and the thickness of the Aruküla Stage in Estonia. AB, CD and KL indicate transects of stratigraphical cross-sections presented in Figs. 75, 76 and 85. Numerator at the borehole sign marks its number (see Fig.3 for section names), denominator - the thickness of the stage in metres. 1 - contour of rock distribution; 2 - limit of outcrop belt; 3 - isopach; 4 - borehole; 5 - outcrop (2 - Õisu); 6 - type section (1 - Viljandi; 3 - Tarvastu; 4 - Tamme; 5 - Tartu Jaani Cemetery/Aruküla caves); 7 - transect of the cross-section.

Tarvastu beds of the Aruküla Stage (Figs. 75, 76, 85). The topmost layer of the Tarvastu beds is, as a rule, represented by reddish or variegated (purplish-grey to reddish-brown) siltstones, which are overlain by white, yellowish-, brownish- or purplish-grey poorly sorted loose sandstones of the Burtnieki Stage.

The Burtnieki Stage is divided into the Salaca (below) and the Abava (above) substages (Table 10). Mark-Kurik (1993 a,b) has treated the latter as an independant stage.

Fossils coming from different parts of the Burtnieki Stage, the Härma [hm] and Koorküla [kr] beds and the Abava [ab] Substage belong mainly to fishes: *Tartuosteus maximus* Mark-

Kurik [hm], cephalaspidids [hm], *Pycnosteus tuberculatus* (Rohon) [hm,kr], *Ganosteus stellatus* Rohon, *Psammosteus bergi* (Obr.) [hm], *Actinolepis magna* Mark-Kurik [hm,ab], *Tropinema haermae* (Mark) [hm], *Homostius latus* Asm. [hm,kr], *Heterostius ingens* Asm. [hm,kr], *Coccosteus markae* O.Obr. [hm], *Asterolepis* sp.1 Kar.-Tal.[hm], *Homacanthus gracilis* (Eichw.) [hm], *Nodocosta* sp. [kr], *Acanthodes?* sp. A, *Acanthodes?* sp. B, *Acanthodes?* sp. D, *Acanthodes* sp. [ab], *Cheiracanthus brevicostatus* Gross [hm,ab], *C. longicostatus* Gross [hm], *Cheiracanthus* sp. [ab], *Ptychodictyon rimosum* Gross [hm], *P. sulcatum* Gross [hm], ? *Ptychodictyon* sp. [hm], *Diplacanthus carinatus* Gross [hm], *D. gravis* Valiuk. [hm], *Acanthodii* gen.n. Valiuk. [hm], *Markacanthus alius* Valiuk. [hm], *Rhadinacanthus multisulcatus* Valiuk. [hm], *Nostolepis* sp.n. Valiuk. [hm], *Gyroptychius elgae* Vorob. [hm], *Glyptolepis? karksiensis* (Vorob.) [hm], holoptychiids [hm], *Psammolepis* sp.sp., *Byssacanthus* sp.sp. [kr,ab], *Hamodus lutkevitshi* Obr. [kr,ab], *Panderichthys?* sp. [kr,ab], *Psammolepis abavica* Mark-Kurik [ab], *Psammosteus* sp.sp. [ab], *Watsonosteus* sp.n.? [ab], *Livosteus?* sp. [ab], *Plourdosteus? panderi* O. Obr. [ab], *Asterolepis essica* Lyarsk. [ab], *Microbrachius cf. dicki* Traq. [ab], Chondrichthyes? [ab], *Laccognathus* sp. [ab], *Osteolepididae* [kr, ab], *Onychodus?* sp. [ab], *Dipnoi* [ab], *Moythomasia?* sp. [ab], *Cheirolepis* sp. [ab] (Сорокин 1981, Valiukevičius 1994, modified).

Of other fossils, silicified wood has been found in the Härma beds and rare lingulates and various remains of the pteridophyte *Pseudosporochnus estonicus* Kalamees in the upper clayey part of the Abava Substage (Kalamees 1988).

In Estonia and adjacent areas, the Burtnieki Stage is represented by the Burtnieki Formation. On the basis of the lithological and mineralogical data, three cyclic units are distinguished in the Burtnieki Formation. These cycles are observable in all sections of Estonia and defined as the Härma (lower), Koorküla (middle) and Abava (upper) beds (Table 10). Each unit begins with relatively coarse-grained light, variegated (yellowish, pinkish, greyish and brownish) sandstones of a mature mineral composition and ends with clayey silt layers (Figs. 75, 76, 85, Kleesment 1995).

Lithologically and mineralogically (Fig.78), these three beds are rather similar. The sandstones are prevailed by fine-grained fraction which usually forms 50-70% of the rock. The rate of medium-grained and very fine-grained sand fractions is variable, forming 10-30 and 6-20% of the rock, respectively. The share of other fractions rarely exceeds 5%. The predominating thickness of the cross-bedded sandstone series is 20-30 cm. They are dipping to the south, southwest, and southeast. In the Härma beds, the southwest inclination is prevailing, while in the Koorküla and Abava beds the inclination directions are more variable. Siltstones are mostly medium-cemented, variegated, clays are strongly silty, grey, and reddish-brown.

Mineralogically, the rocks of the Burtnieki Formation are predominantly quartzose and feldspatic arenites with the quartz content of 70-90%. Micaceous arenites (content of micas up to 50%) occur as rare thin interbeds. The heavy fraction is dominated by ilmenite (45-65%). The share of allothigenic transparent minerals in the heavy fraction is in general 15-30%. This group is dominated by zircon (40-70%). Of other

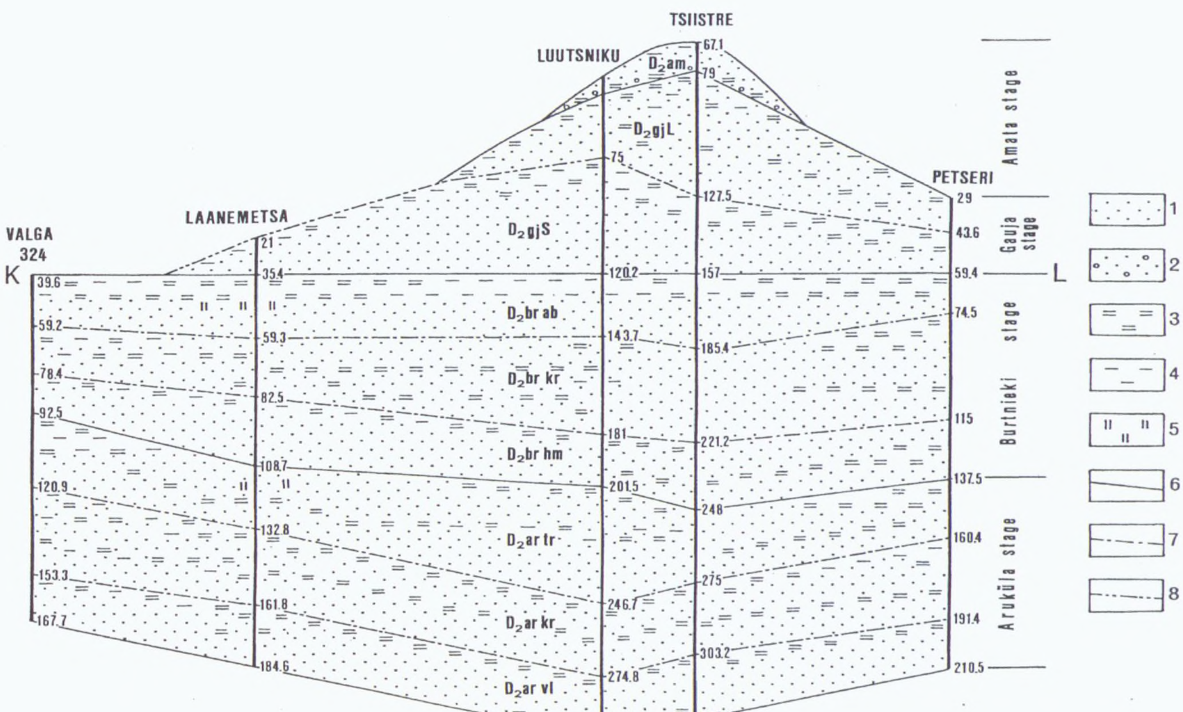


Fig. 85. Stratigraphical cross-section (KL) from Valga-324 (west) to Petseri (east), comprising the Aruküla, Burtnieki, Gauja and Amata stages. For location of the section see Fig. 84. Boreholes: Valga - 265, Laanemetsa - 269, Luutsniku - 280, Tsiistre - 279 and Petseri - 276. Stratigraphical indices: D_2ar vl - Viljandi, D_2ar kr - Kureküla, D_2ar tr - Tarvastu, D_2br hm - Härma and D_2br kr - Koorküla beds; D_2br^{ab} - Abava Substage; D_2gj S - Sietiņi and D_2gj L - Lode members; D_2am - Amata Stage. 1 - sandstone; 2 - conglomeratic sandstone; 3 - siltstone; 4 - clay; 5 - dolomitic matrix; 6 - boundary between stages; 7 - boundary between beds (substages); 8 - boundary between members.

Photo 25. Sandstones of the Burtnieki Stage in the bank of the Ahja River. *Photo by A. Miidel.*

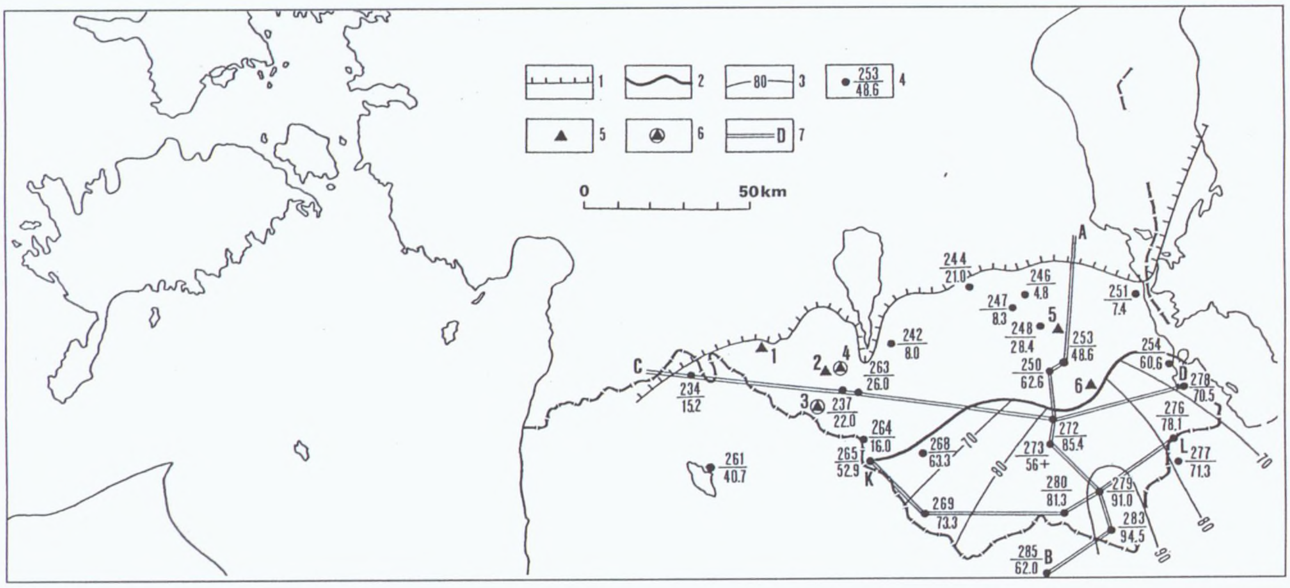


Fig. 86. Sketch-map showing the present distribution and the thickness of the Burtnieki Stage in Estonia. AB, CD and KL indicate transects of stratigraphical cross-sections presented in Figs. 75, 76 and 85. Numerator at the borehole sign marks its number (see Fig.3 for section names), denominator - the thickness of the stage in metres. 1 - contour of rock distribution; 2 - limit of outcrop belt; 3 - isopach; 4 - borehole; 5 - outcrop (1 - Karksi; 2 - Helme; 5 - Ahja; 6 - Essi); 6 - type section (3 - Koorküla; 4 - Härma); 7 - transect of the cross-section.

accessory minerals, tourmaline (7-20%) and staurolite (3-15%) are more important, noteworthy is the appearance of kyanite (Fig. 78). The share of tourmaline is greatest in the Abava beds where it makes up 10-30% of the group of transparent allothigenic minerals.

Gauja Stage

The Gauja Stage was formally established by Liepinš (1951), although the corresponding stratigraphical unit as a Stage already existed in the scheme of Krauș (1934) and had been distinguished by Gross (1942) as "Oredesch-Stufe". In different periods it has been treated as a separate stage or as the lower part of the Šventoji Stage (Сорокин 1981). The stratotype of the Gauja Stage is the bank of the Gauja River between Cēsis and Sigulda in northern Latvia. In Estonia more important localities are the banks of the Piusa, Pärlijõgi and Mustjõe rivers, Tuhkvitsa Brook and sand quarries near the railway station at Piusa (Fig. 87).

The Gauja Stage is spread in a restricted area in south-eastern Estonia. The outcrop area forms a 14–30-km-wide belt which extends from Valga and Luutsniku in the west to Karisilla and Petseri in the east. The total thickness of the stage in Estonia ranges from 78 to 79.8 m (Fig. 87).

The Gauja Stage consists mostly of weakly- to medium-cemented white and light- to yellowish-grey cross-bedded sandstones. It lies everywhere on the topmost clayey-silty complex of the Abava beds of the Burtnieki Stage (Figs. 75, 76, 85). On the contact level the sandstone is often rich in carbonate cement.

The lower and upper parts of the Gauja Stage differ in fossils. The lower, Sietiņi Member has yielded fossil fishes: *Psammolepis venyukovi* Obr., *P. paradoxa Ag., P. heteraster* Gross, *P.alata* Mark-Kurik, *Plourdosteus livonicus* (Eastm.), *Asterolepis ornata* Eichw. *sensu* Ag., *Bothriolepis?* sp., *Glyptolepis baltica* Gross, *Laccognathus panderi* Gross and *Megadonichthys kurikae* Vorob. in litt. In the Sietiņi Member also some large fragments or stems, or both, of silicified and ferriferous wood have been found (Сорокин 1981, modified).

In the Lode Member, only plant macroremains (*Hostinella* sp., *Archaeopteris* sp., *A. fissilis* Schalh.) and miospores are known. The miospore assemblage includes: *Retusotriletes rugulatus* Riegel, *Samarisporites triangulatus* Allen, *S. eximius* (Allen) Loboziak et Streel, *Geminospora micromanifesta* (Naumova) Arkh., *G. lemurata* Balme, emend. Playford, *Ancyrospora* sp. cf. *A. incisa* (Naumova) M. Rask. et Obukh., *Dictyotriletes* sp. cf. *Reticulatisporites perlotus* (Naumova) Obukh., *Perotriletes* sp. cf. *Rugospora? impolita* (Naumova) Tchib. (Blieck *et al.* 1996).

In Estonia and adjacent areas, the Gauja Stage is represented by the Gauja Formation. In the latter, two cyclic complexes can be distiguished, corresponding to the Sietiņi (lower) and Lode (upper) (Table 10) members, established by Kuršs (Куршс 1992). The Sietiņi Member consists mostly of sandstones, with variegated siltstone in the topmost part. The lower part of the Lode Member is represented by light, mainly white sandstones, its upper part is dominated by siltstones and clays (Figs. 75, 85).

The sandstones of the Gauja Formation are fine-grained. The share of fine-grained particles is usually 55-65%, the content of very fine-grained particles is 22.5%, on an average. The cross-bedded series are 5-40, mostly 15-30 cm thick, predominately inclined to the southwest, south and southeast. Characteristic are brown iron-rich surfaces, pebbles of purplish-brown and grey clay, quartz and Fe hydroxide.

The siltstones, which form ca 20 % of the section on an average, are usually clayey, represented by variegated, grey and brownish varieties. Clays (average 15%) are strongly silty, grey, and purplish-grey, often dolomitic.

Mineralogically, the rocks of the Gauja Formation are predominantly quartzose arenites with the quartz content of 80-94%. The heavy fraction is dominated by ilmenite, transparent allothigenic minerals make up 20-30%. In the latter group, the leading mineral is zircon, although in the Lode Member tourmaline often dominates (Fig. 78). The Sietiņi and Lode members differ notably in the composition of clay minerals. In the Sietiņi Member, the average share of hydromicas is 75% and kaolinite 25%. In the Lode Member, these values are 45 and 55%, respectively. The Lode Member is the most kaolinite-rich level in the Devonian sequence of Estonia.

Amata Stage

The Amata Stage was formally established by Liepinš (1951), although the corresponding stratigraphical unit as Stage existed in the scheme of Kraus (1934) and had also been distinguished as the "Podsnetogor-Stufe" by Gross (1942). The stratotype of the stage is situated at the lower course of the Amata River in Latvia. In Estonia, the main outcrops are the banks of the Piusa (Loosi) and Peetri rivers (Karisöödi), and in the vicinity of Vastseliina (Fig. 88).

The Amata Stage is spread in a restricted area in southeastern Estonia. The outcrop area forms a 5–10-km-wide belt from Mõniste and Ape in the west to Petseri and Dekshino in the east. The total thickness in boreholes ranges from 12 - 21 m (Fig. 88), but in the outcrops on the banks of the Piusa River it reaches 30 m.

In Estonia, the Amata Stage is represented by sandy-silty sediments alternating with frequent clay interbeds. The stage lies everywhere on the grey clay of the Gauja Stage and starts with a layer of breccia-like sandstone (Figs. 75,85). According to Kuršs (Куршс 1992), in the lower Amata layers the cross-bedded series are inclined to the north which is not typical of this part of the Devonian.

The Amata Stage contains *Psammolepis undulata* (Ag.), *Psammosteus praecursor* Obr., *P. maeandrinus* Ag., *Asterolepis radiata* Rohon, *Bothriolepis prima?* Gross, *B. cellulosa?* Pand., *Panderichthys rhombolepis* Gross and *Laccognathus panderi* Gross (Сорокин 1981).

In Estonia and adjacent areas, the Amata Stage is represented by the Amata Formation. According to borehole data,

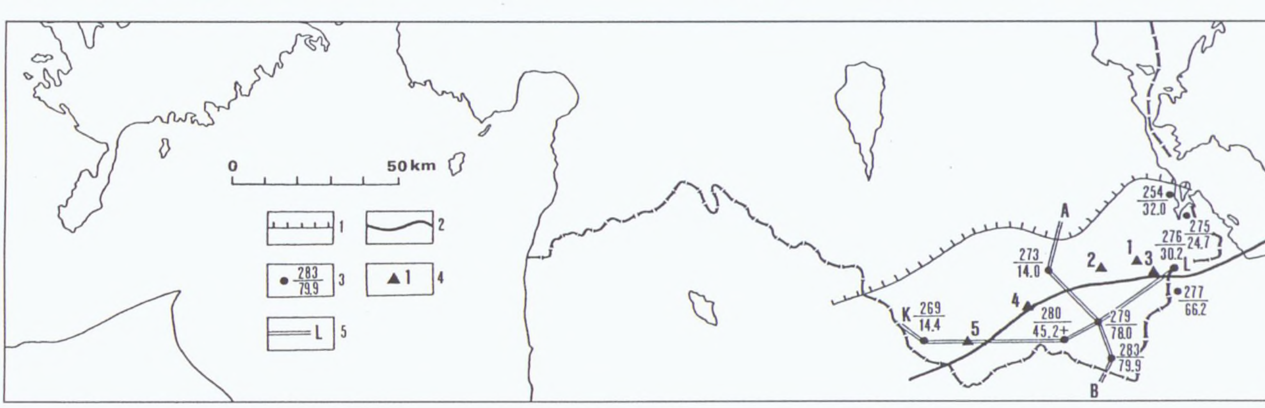


Fig. 87. Sketch-map showing the present distribution and the thickness of the Gauja Stage in Estonia. AB and KL indicate part of the transects of stratigraphical cross-sections presented in Figs 75 and 85. Numerator at the borehole sign marks its number (see Fig. 3 for section names), denominator - the thickness of the stage in metres. 1 - contour of rock distribution; 2 - limit of outcrop belt; 3 - borehole; 4 - outcrop (1 - Piusa glass sand quarries; 2 - Piusa River; 3 - Tuhkvitsa Brook; 4 - Pärlijõgi River; 5 - Mustjõgi River); 5 - transect of the cross-section.

the predominating rock type in the Amata Formation is the greenish- and purplish-grey siltstone which forms on average of 45% of the section. In outcrops, however, the sandstones are prevailing. The sandstones of the Amata Formation are light to yellowish-grey, more rarely reddish-brown, fine-grained, medium- to strongly-cemented, with indistinct cross-bedding, the inclination of which varies in wide limits. The sandstones often contain pebbles and lens-shaped interlayers of clay, more rarely quartz pebbles. Clay interlayers are usually purplish-grey and -brown and form 30% of the section as an average.

Mineralogically, the sandy-silty rocks of the Amata Formation are predominantly quartzose arenites with the quartz content being 80-90%. The heavy mineral suite is dominated by ilmenite, the share of transparent allothigenic minerals is relatively great, varying from 26 to 40%. Among this group zircon is predominating. It is followed by tourmaline, staurolite and rutile in almost equal amounts (Fig. 78, Kleesment 1995). The assamblage of clay minerals is dominated by hydromicas with the average content of 95%.

UPPER DEVONIAN

In southeastern Estonia, the Upper Devonian is represented by carbonate rocks, the thickness of which reaches 47 m in the Parmu borehole. The outcrop belt of rocks of the Upper Devonian Pļaviņas, Dubniki and Daugava stages has a complicated configuration. In the area where the clayey-sandy sediments of the Middle Devonian Gauja and Amata stages crop out, single outlier-islands of carbonate rocks are encountered, the largest being Saarlase and Loosi (Fig. 89).

Pļaviņas Stage

The Pļaviņas Stage and the Pļaviņas Formation have been named after the exposures in the vicinity of the Town of Pļaviņas in Latvia (Liepinš 1951). Currently, these exposures are under the waters of the Pļaviņas reservoir and, therefore, the outcrops near Izborsk (Irboska) have been selected as the neostratotype for the Pļaviņas Stage.

In Estonia and adjacent areas, the Pļaviņas Stage has a thickness of 27-32 m. In the vicinity of Izborsk it is 37 m thick (Fig. 90). The lower boundary of the stage is lithologically clear - the clayey sandy deposits of the Amata Stage are overlain by carbonate rocks of the Pļaviņas Stage. Based on the palaeontological and lithological characteristics, the Pļaviņas Stage has been subdivided into the Snetnaya Gora, Pskov and Chudovo substages.

The Snetnaya Gora Substage has been named after a type section near the Snetnaya Gora Monastery in the vicinity of the Town of Pskov in Russia. In Estonia, the rocks of the substage crop out at the Peetri River upstream of Karisöödi, at the Pärli River near the Saarlase Mill, in the Rõuge Ööbikuorg, in the environs of Loosi and Vastseliina (Fig. 89). According to the borehole data, the thickness of the substage is ranges from 5.5 to 12 m, and increases from west to east (Fig. 90).

The substage is represented by rhythmically alternating yellowish- and greenish-grey micro- to cryptocrystalline argillaceous silty dolomite (MgO 16%, CaO 24%) and dolomitic marl (insoluble residue 30%, MgO 13%, CaO 19%), less frequently by clay. Dolomites and dolomitic marls contain silty interlayers. The complex is micro- and thin-laminated. Imprints of cubical salt crystals are found in dolomite. In northern regions, thin sand interlayers occur. At the Peetri River, the lower part of the section is composed of clay, and the upper part of dolomite.

The fossils, occasionally found in the section, are represented by the brachiopods *Camarotoechia aldoga* Nal., conchostracans *Asmussia vulgaris* Lutk. and the fishes *Psammosteus meandrinus Ag., Ctenurella pskovensis* (Obr.) and *Bothriolepis cellulosa* Pand., *Grossilepis tuberculata* (Gross), *Moythomasia perforata* (Gross).

The Pskov Substage has been named after the type section on the bank of the Velikaya River near Pskov in Russia. The exposures occur in the same area where those of the Snetnaya Gora Substage are situated. According to the borehole data, the Pskov Substage is 7-13 m thick, on the base of the exposures in the vicinity of the Izborsk Castle (Russia) it is about 17 m thick (Fig. 90).

The Pskov Substage is represented by grey, in the lower part by pale purplish limestone. The rate of dolomitization grows to the west. In the Karisöödi area, the lower part of the substage consists of dolomite (MgO 20%, insoluble residue 6%) with 3–10-cm-thick clay interlayers. In the east (Tsiistre, Hino, Vungi), the substage is mainly represented by thin-layered, often cavernous dolomite (MgO 20%), partly silty-

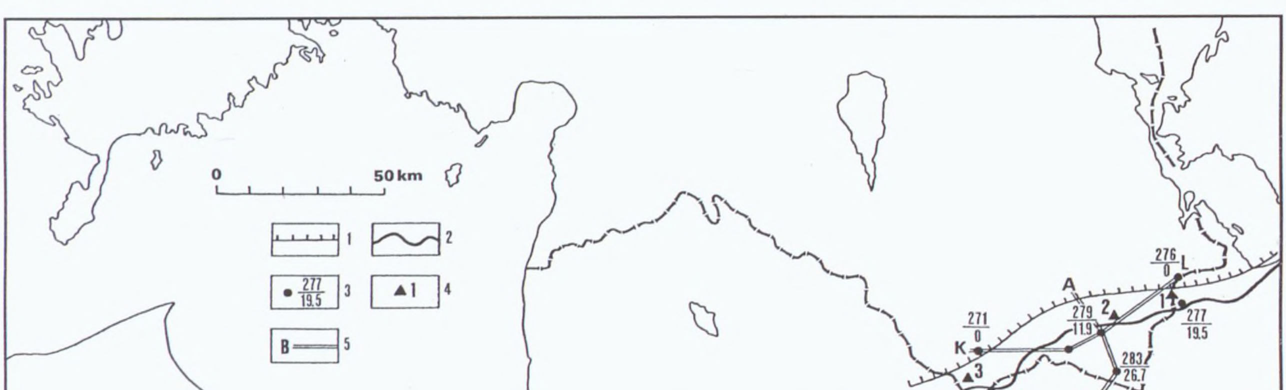


Fig. 88. Sketch-map showing the present distribution and the thickness of the Amata Stage in Estonia. AB and KL indicate part of the transects of stratigraphical cross-sections presented in Figs 75 and 85. Numerator at the borehole sign marks its number (see Fig. 3 for section names), denominator - the thickness of the stage in metres. 1 - contour of rock distribution; 2 - limit if outcrop belt; 3 - borehole; 4 - outcrop (1 - Loosi; 2 - Vastseliina; 3 - Karisöödi); 5 - transect of the cross-section.

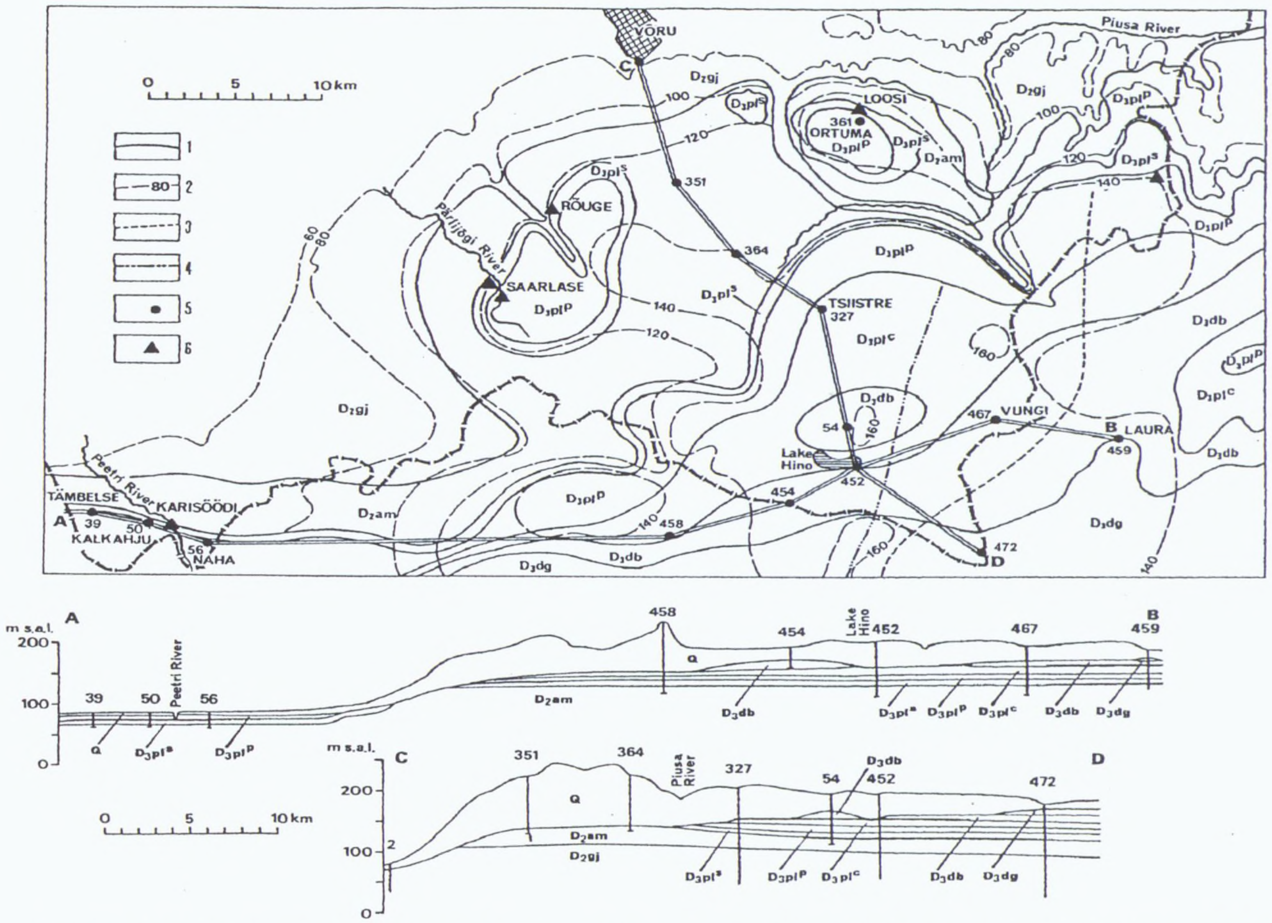


Fig. 89. Bedrock map of southeastern Estonia with geological cross-sections AB (latitudinal) and CD (meridional). Stratigraphical indices: D_2gj - Gauja Stage, D_2am - Amata Stage, D_3pl^s - Snetnaya Gora Substage, D_3pl^p - Pskov Substage, D_3pl^c - Chudovo Substage, D_3db - Dubniki Stage, D_3dg - Daugava Stage. 1- boundary of stages and substages; 2 - isoline of bedrock relief; 3 - boundary of dolomitic (west) and limestone (east) facies of the Pskov Substage; 4 - boundary of dolomitic (west) and limestone (east) facies of the Chudovo Substage; 5 - borehole; 6 - outcrop.

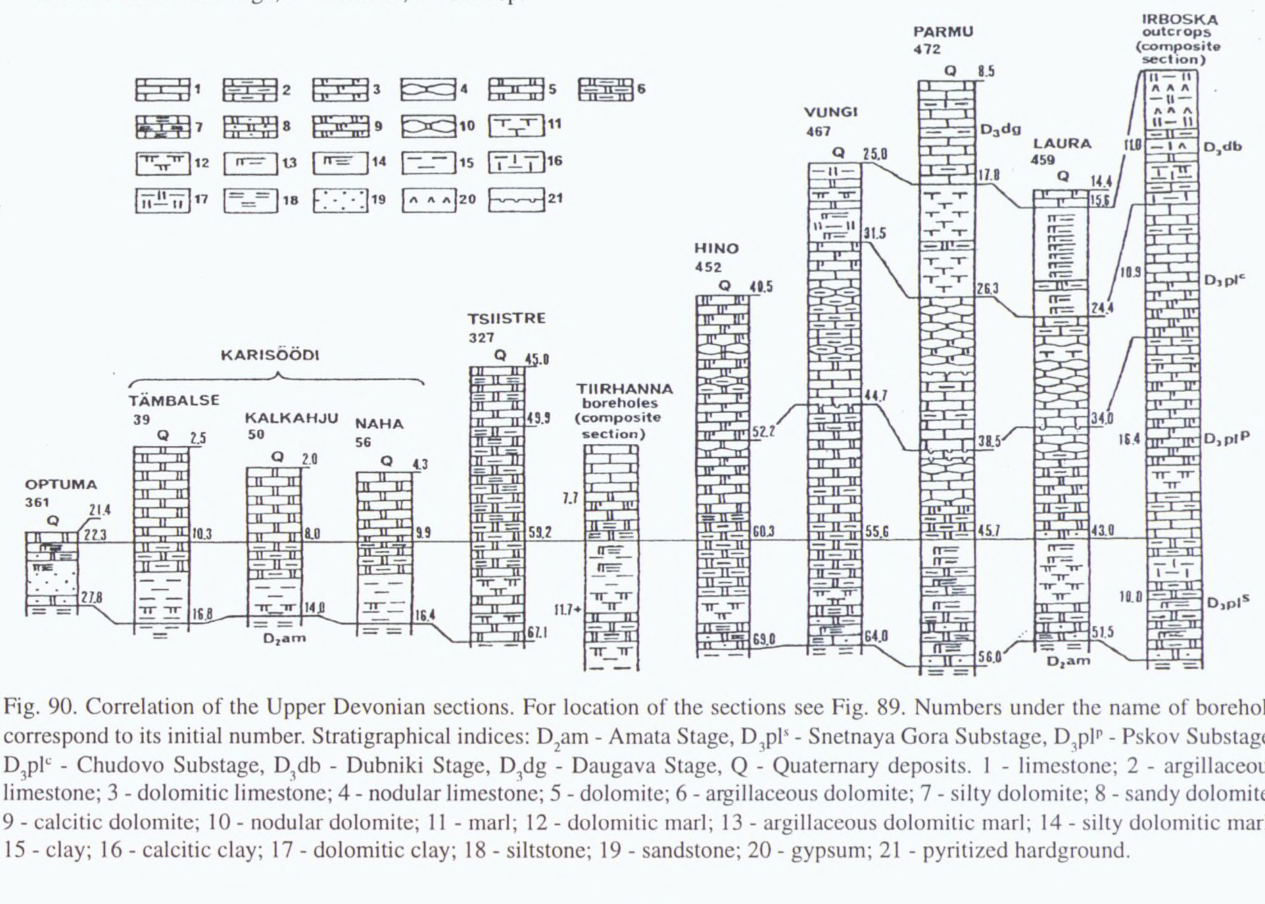


Fig. 90. Correlation of the Upper Devonian sections. For location of the sections see Fig. 89. Numbers under the name of borehole correspond to its initial number. Stratigraphical indices: D_2am - Amata Stage, D_3pl^s - Snetnaya Gora Substage, D_3pl^p - Pskov Substage, D_3pl^c - Chudovo Substage, D_3db - Dubniki Stage, D_3dg - Daugava Stage, Q - Quaternary deposits. 1 - limestone; 2 - argillaceous limestone; 3 - dolomitic limestone; 4 - nodular limestone; 5 - dolomite; 6 - argillaceous dolomite; 7 - silty dolomite; 8 - sandy dolomite; 9 - calcitic dolomite; 10 - nodular dolomite; 11 - marl; 12 - dolomitic marl; 13 - argillaceous dolomitic marl; 14 - silty dolomitic marl; 15 - clay; 16 - calcitic clay; 17 - dolomitic clay; 18 - siltstone; 19 - sandstone; 20 - gypsum; 21 - pyritized hardground.

argillaceous (insoluble residue 12-20%), in the upper part of the section it is calcareous in places (CaO 30-35%). On the east margin of the Haanja Heights (Tiirhanna, Parmu), the lower beds are represented by dolomite, the upper ones by limestone.

The Pskov Substage is rich in fossils. The brachiopods *Anatrypa micans* (Buch), *Atrypa velikaya* Nal., *Ladogia meyendorfii* (Vern.), *Ripidiorhynchus pskovensis* (Nal.) dominate. Calcareous algae have also been found.

The Chudovo Substage was differentiated on the basis of the exposures in the vicinity of the Town of Chudovo, Russia. In Estonia, the Quaternary cover is thick and the rocks of the Chudovo Substage are not exposed. The substage crops out near the Pskov - Riga highway. Based on the key fossils *Ripidiorhynchus tschudovi* (Nal.) and *Anatrypa heckeri* Nal., the age of the substage has been established in the Izborsk outcrops in Russia. The thickness of the substage in the boreholes reaches 13 m. In places (Laura, Vungi, Parmu), the lower boundary of the substage is marked by a pyritized discontinuity surface.

In the easternmost part of its distribution area (Vungi, Parmu), the Chudovo Substage is represented by micro- and cryptocrystalline limestones (CaO 44-49%, insoluble residue 5-10%). Dolomitization of rocks increases westwards and the substage consists of micro- and fine-crystalline dolomitic limestones (CaO 32-34%, MgO 16-17%, insoluble residue 3-7%) to coarse-crystalline cavernous dolomites (CaO 28-29%, MgO 20%, insoluble residue 3-7%). Dolomitic facies is spread west of Misso.

The Pskov and Chudovo substages are lithologically very similar and sometimes it is expedient to treat them together as the Izborsk Member (Table 10).

Dubniki Stage

The Dubniki Stage and the Dubniki Formation have been named after the former gypsum quarry which is situated east of Izborsk Town in Russia (Bekker 1924a). In the walls of the quarry up to 12.5 m thick bed of greenish-grey clay with gypsum and dolomite interlayers was exposed. Fossils are represented by the brachiopods *Comiotoechia bifera* (Phill.), *Ripidiorhynchus strugi* (Nal.) and the ostracod *Acratia benevaensis* Zasp.

The Formation covers a limited area in the southeasternmost part of Estonia, but the exposures do not occur there because of the thick Quaternary cover (Fig. 89). The thickness of the stage reaches 10 m (Fig. 90) in the boreholes . The section consists of bluish-grey marl (CaO 33%, MgO 3% and insoluble residue 27%) and argillaceous dolomitic marl (CaO 7%, MgO 6%, insoluble residue 58%) with clay and dolomite interlayers.

Daugava Stage

The Daugava Stage and the Daugava Formation have derived the name from the exposures on the banks of the Daugava River, Latvia (Liepinš 1951). In Estonia, the uppermost 8.5 m of the Devonian section in the Parmu borehole belong to the Daugava Stage (Fig. 90). In the borehole, the stage is represented by argillaceous micro- and cryptocrystalline limestones (CaO 49%, MgO 2%, insoluble residue 27%).

V QUATERNARY COVER

Structure of the Quaternary cover

Estonia belongs to the zone of glacial erosion or moderate accumulation and, therefore, the Quaternary cover is rather thin. In northern Estonia, on the outcrops of the Ordovician and Silurian carbonate rocks it is usually less than 5 metres. Occasionally, on the so-called alvars, it is even lacking (Photo 26). The Quaternary cover is at its thickest (Fig. 91) in the Haanja and Otepää heights (often more than 100 m) and in the buried valleys of southern Estonia (at Keskküla 207 m). Thick Quaternary deposits are also encountered on lee-sides of heights and elevations (*e.g.* the Saadjärve Drumlin Field in the "shadow" of the Pandivere Upland) and in front of escarpments oriented against the movement of glaciers which favoured the accumulation of a considerable amount of sediments. In the fore-klint area, the deposits are more than 100 m and reach 143 m in the ancient Harku Valley (in the vicinity of Tallinn). In the Gulf of Finland, the greatest thicknesses have been established in megadrumlins to the north of Tallinn, for example, on the Island of Prangli, where the Pleistocene deposits are up to 123 m thick. Here the Quaternary cover comprises several till layers (Kajak 1961, 1995; Каяк 1965a, Раукас 1978).

About 95% of the Pleistocene cover is formed of glacial and aqueoglacial deposits. Glacial sediments, 70% by volume and 47.7% by surface area, dominate (Раукас 1978). Of wide distribution are also glaciolacustrine (6.8% by area) and glaciofluvial (3.1%) deposits (Fig. 92). Five till beds, often of great thickness, are to be noticed more or less distinctly. Only in a few cases they are separated from one another by deposits containing spores and pollen of interglacial or interstadial origin which considerably aggrevates the correlation and dating of the glacial strata.

Photo 26. An alvar at Kõmsi, western Estonia. *Photo by A. Raukas.*

In terms of glacial stratigraphy and lithogenesis, the bedrock valleys and interlobate "insular heights" are the main objects of interest. Deep ancient river valleys, which were further eroded by glaciers and their meltwaters, vary in morphology and sediment facies infill. They may be filled with glaciofluvial deposits of the last glaciation (Pada Valley), glaciolacustrine deposits of the last glaciation (Selja Valley) or one till bed with under- or overlying glacio-aquatic deposits (Kunda Valley). There are also many valleys with a complicated structure comprising several till beds and accompanying glacio-aquatic deposits. Valleys of the first three types are characteristic of northern Estonia, whereas valleys with a complicated structure prevail in the southern part of the Republic (Таваст и Раукас 1982). The deposits of radial and marginal valleys also have certain areal differences. The de-

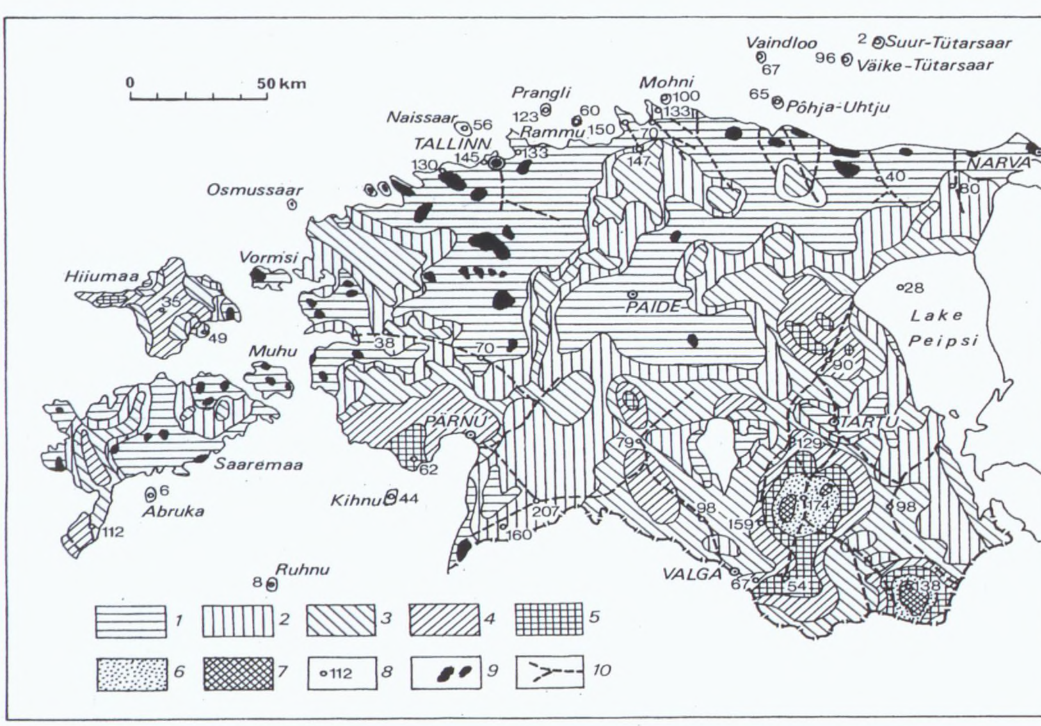


Fig. 91. Thickness of the Quaternary deposits: 1 - below 5 m; 2 - from 5 to 10 m; 3 - from 10 to 20 m; 4 - from 20 to 40 m; 5 - from 40 to 60 m; 6 - from 60 to 80 m; 7 - over 80 m; 8 - boreholes with the thickness of the Quaternary deposits; 9 - alvars; 10 - buried valleys.

Fig. 92. The map of Quaternary deposits: g - till, f - glaciofluvial, lg - glaciolacustrine, m - marine, a - alluvial, l - lacustrine, b - boggy, t - technogenous; black - Pre-Quaternary.

posits of radial valleys have to a large extent been reworked by glaciers and contain less older deposits. With regard to their age, the Upper Pleistocene and Holocene deposits predominate in the buried valleys of northern Estonia and Middle Pleistocene deposits in the valleys of southern Estonia (Raukas & Tavast 1987). The deposits of ancient valleys are the most suitable objects for stratigraphical studies because they contain less erratics than uplands. In the latter, the blocks of bedrock and older Quaternary sediments have been displaced not only horizontally, but occasionally also a considerable up-thrusting or folding, or both have taken place. Older beds are thus found standing more or less vertically in a position tens and even hundreds of metres above their normal stratigraphic position (Raukas & Gaigalas 1993).

The so-called "accumulative insular heights" (the name is derived from the "island-like" position in the topography) form three belts with a N-S orientation in the East-European Lowland (Аболтиньш и др. 1989). Estonian insular heights-Haanja and Otepää - belong to the Latgale Zone. These heights are characterized by a hummocky topography and by a considerable thickness of Quaternary deposits (up to several hundred metres). They have a rather great altitude, distinct slopes, plateau-like forms of glaciolacustrine origin in water-divide areas, and usually a bedrock core. During their development they have undergone the following four morphogenetical stages: (1) subglacial, (2) englacial, (3) marginal accumulation, and (4) a dead ice stage (Аболтиньш и др. 1989). The heights have formed between rather large ice lobes as a result of frequent redeposition of older deposits, accounting for the mosaic pattern of sediments. Representative outcrops on heights have served as the main areas of stratigraphic investigations for over a century. As a result of redeposition of interglacial and interstadial sediments, the number of supposed interglacials could be erroneously increased. This, in turn, may lead to an older age being assigned to the tills separating them and to misleading palaeogeographical conclusions (cf. *e.g.* discussion in Liivrand 1991). A precise correlation of Pleistocene deposits assumes great knowledge of glaciosedimentation processes and elaboration of special dating methods.

Classification and composition of deposits

The classification of the Estonian Quaternary deposits is based on the study of genetical types of deposits resulting from the development of a certain dynamic form of accumulation, and playing qualitatively different role in the structure and history of the formation of the sedimentary cover (Раукас 1978). Genetical types can be merged into paragenetical series, groups and subgroups, and, at need, into smaller taxonomical units (facies and subfacies).

Among the Estonian Pleistocene deposits six paragenetical series occur: eluvial, organogenous, colluvial and deluvial, aqueous, glacial and subaerial.

In the paragenetical series of eluvial deposits, areal and linear (along the crevasses and zones of tectonic faults) crusts of weathering, soil horizons on the boundaries of stadial or phasial beds of different age, and the deposits and relief forms of permafrost (cryogenous eluvium) occur in some places. Cryogenous phenomena (occurrences of cryoturbation, ice wedges, bedding disturbances and structural grounds) are most widespread in the Younger Dryas sediments (Photo 27).

Deposits of organogenous paragenetical series occur as interglacial peat and gyttja in few places (Rõngu, Karuküla, *etc.*). Organogenous deposits did not accumulate in late-glacial times and the thin interlayers of peat are most probably redeposited.

Colluvial and deluvial deposits are relatively frequent in front of Ordovician and Silurian escarpments and in the hilly topography of southeastern Estonia where in late-glacial times soliflunctional processes resulting from the melting of buried ice played a significant role.

Deposits of aqueous paragenetical series are represented by several genetical types of different lithological composition. Among those, proluvial and subsurface aquatic deposits are rare. Little is also known about the interglacial and interstadial alluvial deposits represented by intermorainic silty-sandy sediments (Valguta, Peedu) and sandy gravelly deposits in the ancient buried valleys. Late-glacial alluvial deposits of the last glaciation occur in the terraces of a great many river valleys in southern Estonia where their thickness usually ranges between 3 and 10 m.

Interglacial and interstadial lacustrine deposits are represented by gyttja (Karuküla, Rõngu) and silty-sandy sediments (Otepää, Sudiste). Late-glacial terrigenous lacustrine sediments of the last glaciation are to be found under many contemporary bogs. The boundary between the Pleistocene and Holocene lacustrine deposits is rather clear and easy to notice due to an abrupt increase of organic matter in the Holocene deposits or carbonates in the form of lacustrine lime.

Marine interglacial and interstadial deposits are distributed in the fore-klint area (Prangli Island) and on the islands (Kihnu) of the Gulf of Liivi (Riga). The deposits of the Baltic Ice Lake, occurring as bottom and coastal formations, are conventionally also regarded as marine deposits.

As mentioned above, deposits of glacial paragenetical series form a great part of the Pleistocene cover in the Republic (Fig. 2). All the deposits of the glacial paragenetic series can be divided into two paragenetic groups: glacial drift deposited by glaciers on ground (subaerial tills) and underneath ice shelves (subaqueous tills). Among the subaerial varieties the lodgement, superglacial (ablation) and frontal (margin) tills and among the subaqueous varieties iceberg and submarine ablation tills can be distinguished as genetic types (Раукас 1978). On grounds of detailed micropalaeontological, geochemical, geomorphological, structural and other studies, it has been proved that the tills in Estonia are mainly of continental subaerial genesis (Раукас 1973).

Marine microorganisms (diatoms, foraminifers, ostracodes, *etc.*) are very rare in Estonian tills. Microfauna and -flora has been discovered in greater amount only in a narrow fore-klint area which in the past, and most likely during the interglacial stages as well, was occupied by the sea or big glaciolacustrine basins. The borehole at Vääna-Jõesuu provides an excellent illustration of the above. However, also there the content of foraminifers and diatoms in till is remarkably lower than in intermorainic silty clays (Раукас и Лийвранд 1971). The occurrence of foraminifers and diatoms predominantly in the intermorainic layer evidences of the fact that the microfauna and -flora, at least partially, had redeposited here from the bottom of the Gulf of Finland where marine conditions were of repeated existence in the past.

Geochemical and most of the lithological evidence support the theory of subaerial genesis of Estonian glacial deposits. For instance, this is indicated by the great similarity between the lithological composition of tills and underlying rocks, poorly sorted sediments, lack of new authigenous formations of marine origin, poor rounding of clasts and the increase in roundness towards the south and south-east, *i.e.* in the direction of the supposable movement of continental ice, but also the distribution of indicator (index) boulders, a relatively high content of clayey particles and a high degree of compaction in tills, the orientation of clasts in the direction

Photo 27. Different permafrost phenomena in Estonia are most widespread in Younger Dryas sediments. Late-glacial silts on the bank of the Purtse River. *Photo by Karl Orviku.*

of advance of ice masses, parallel to glacial striae and almost horizontal in position, the occurrence of glaciodislocations (Photo 28) *etc.* (Раукас 1973).

Lodgement tills are of the widest distribution. The formation of subaerial tills is thought to have taken place both beneath an advancing glacier (lodgement till) and as a result of bottom melting of a passive glacier during the degradation of glacial covers (basal melt-out till). In Estonia, deformation tills are quite frequent. They are formed by subsole drag underneath the moving glacier (Раукас 1978).

Due to the flatness of the topography, the content of supraglacial material in tills is quite low in Estonia. By contrast, the importance of englacial and basal debris is much greater and quite different. In most cases, the material of basal debris predominates, whereas in places, lodgement tills composed of the material of englacial debris (the so-called erratic tills) are also found. Besides, local tills consisting entirely of local sedimentary material are spread (Photo 29). The vast majority of lodgement tills belong to the intermediate group between local and erratic varieties containing local and allochthonous (far-transported) material in different ratios. According to Gaigalas (Гайгалас 1969), it would be expedient to call them transitional lodgement tills. In transitional lodgement tills the local sedimentary material is prevailing.

Ablation tills resulting from surficial melting and gravity flowing of superglacial and englacial debris are represented by flow tills and melt-out tills. Usually they are difficult to distinguish from lodgement tills. Exceptionally, lodgement tills are immediately overlain by ablation tills.

Frontal (margin) tills are present in end moraines which fall into push and dump moraines. Among dump moraines stationary and recessional forms are distinguished. There occur also end moraines of complicated structure which bear traces of ice pressure and are overlain by glaciofluvial deposits, or *vice versa*, as well as interior peripheral moraines of push character developed between dead and active ice (Раукас и др. 1971). Frontal tills consist of various squeeze lodgement, deformation and flow-till facies with injections of aqueoglacial deposits and bedrock erratics.

The formation of lithological and mineral composition of tills depends on a number of factors, such as the composition and topography of the underlying bedrock, the dynamics of the movement of glacier, the location of material in the body of the glacier, the character of accumulation, the nature and intensity of weathering of material, *etc.*

Numerous investigations have enabled to elucidate that on its way the glacier accumulated in its body great quantities of local bedrock material (Raukas 1969, a.o.). At that, maximum content of rock particles (about 60-80%) from a certain stratigraphic unit is usually traceable near the distal boundary of the outcrop of the unit. Already at a distance of 6-8 km from the bedrock boundary, the amount of rock particles from the corresponding unit does not exceed 20-30% of the total (Раукас 1978). The content of erratic material in a typical lodgement till does not usually exceed 5-10%, but in englacially formed till it amounts to 100%. The transport distance of clasts is greatly dependent upon the resistance of rocks. Claystones and weakly cemented sandstones have travelled no more than 15 km, fine-grained limestones 120 km, dolomites 300 km and resistant varieties of crystalline rocks 800-900 km. During transport the till becomes enriched with more resistant clasts. For example, as crystalline rocks are more durable than carbonaceous ones, the South-Estonian tills on sandstones are enriched with crystalline clasts. At the same time, the fragments of carbonaceous rocks become enriched with dolomites as the tougher ones (Раукас 1978).

The local bedrock exerts also a remarkable influence on the mineral composition of tills (Raukas 1974, a.o.). For example, territorial differences are easy to trace on the outcrops of Cambrian clays and siltstones, Ordovician and Silurian carbonaceous rocks and Devonian sandstones. However, even smaller dependences can be traced from territorial aspect and with respect to the composition of different minerals. Thus, the content of carbonaceous minerals decreases abruptly and that of quartz increases southwards from the outcrops of carbonate rocks. The content of amphiboles, pyroxenes and other minerals of heavy subfraction, typical of the outcrops of Precambrian rocks of Finland, gradually decreases in the southern and southeastern direction. Correspondingly, the quantity of weathering-resistant minerals typical of the underlying Palaeozoic rocks, such as garnet, zircon, tourmaline, rutile, *etc.*, increases. Great variations in the proportions of these minerals occur depending upon local conditions. In favour of the above speaks the content of garnet and zircon in the tills of southern Estonia where the underlying Devonian rocks display distinct regularities with respect to those minerals (Raukas 1974).

During the various glaciations, the movements of glaciers have differed (Таваст и Раукас 1982). This enables correlation of till beds on the basis of lithological (Table 11) and

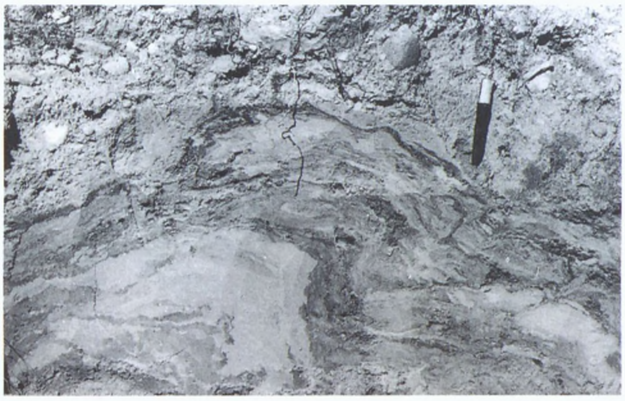


Photo 28. Subaerial tills in Estonia often display glaciotectonic structures of various sizes that indicate ice sheet activity. Järva till at Lelle, central Estonia. *Photo by A. Miidel.*

Photo 29. Grey stony till, rich in local carbonate material at Ruila, northern Estonia. *Photo by A. Raukas.*

mineralogical (Table 12) data.

Of lithological methods, most promising for solving the problems of the Pleistocene stratigraphy and palaeogeography seems the study of crystalline indicator (index) boulders, the content of which in deposits was only slightly influenced by the differences of the local bedrock, and has remained almost stable over vast areas (Раукас 1963б).

The paragenetical subgroup of glaciofluvial deposits is divided into englacial and periglacial genetical types with frequent transitions between them. The deposits of radial eskers and fluviokames are conventionally regarded as englacial glaciofluvial deposits. The deposits of glaciofluvial deltas (Photo 30), sandurs and marginal eskers are identified as periglacial. Glaciofluvial deposits are mostly characterized by a highly variable grain-size composition and structure, and the great variation in lithological and mineralogical composition, everywhere closely connected with the composition of the adjacent till and the bedrock (Раукас 1978).

The maximum distance of transport of pebbles in glaciofluvial streams extends to 16-20 km, being naturally controlled by a great many additional factors, *e.g.* the hardness of rocks, the width of the outcrop of regional stages, the bedrock topography, *etc.* (Раукас и др. 1971). In the course of the formation of glaciofluvial deposits, the content of resistant rocks and minerals increases on account of less stable fractions that are crushed or destroyed during their transportation by water streams. Usually the content of crystalline rocks in gravel and pebble fractions of glaciofluvial deposits is 10-15% higher than in tills, whereas the content of carbonate rocks in them is accordingly lower. The content of metamorphic (predominantly gneisses) and coarse-grained magmatic rocks (predominantly rapakivi and pegmatites) is 5-10% lower than in the initial tills (Раукас 1978).

Glaciolacustrine deposits are also divided into englacial and periglacial genetical types. Englacial deposits form plateau-like limnoglacial kames (Раукас и др. 1971) and are included in superposed kames (Каяк 1963). They are of the widest distribution in northeastern Estonia, in the vicinity of Iisaku-Illuka, while periglacial deposits are most common in

Photo 30. Glaciofluvial delta deposits at Männiku (Tallinn) with the inclination of laminae in the distal direction. *Photo by A. Raukas.*

the Otepää and Haanja heights.

Periglacial glaciolacustrine deposits, predominantly silts and varved clays (Photo 31), are more frequent in western Estonia, in the fore-klint area and on the Narva Lowland (Fig. 2), but also in the numerous river valleys of southern and northern Estonia (Пиррус 1968). The thickness of varved clays reaches 26-27 m. The similar mineral composition of tills and varved clays, and the high content of weathering-resistant minerals in clays points to the insignificant role of chemical changes in the transformation of the initial morainic material into glaciolacustrine clay (Пиррус и Раукас 1963).

Deposits of subaerial paragenetical series are located along ancient shorelines of the Baltic Ice Lake and the larger local ice lakes where they form coastal dunes, up to 12 m high, or in the form of a thin mantle covering the sandy beach ridges. More seldom, they are represented by hillocky plains. Fine-grained sand prevails in the deposits.

Stratigraphical subdivision

Several local and regional stratigraphical schemes have been compiled for Estonia (in 1956, 1957, 1961, 1963, 1970, 1976). These were mainly correlative parts of the schemes of the European portion of the former Soviet Union or the Baltic

Table 11. Average lithologic composition of coarse gravel fraction (5-10 mm) of South-Estonian tills of different age according to Raukas (Раукас **1978), %**

Region	Rocks	Age				
		$Q_{II}sn$	$Q_{II}ug_1$	$Q_{II}ug_3$	$Q_{III}jr_{vl}$	$Q_{III}jr_{vr}$
SW Estonia	Limestones	2.7	28.0	45.2	34.1	30.8
	Dolomites	2.1	21.4	29.1	22.1	25.4
	Crystalline	91.2	41.3	24.4	40.6	35.6
	Sandstones	4.0	9.3	1.3	3.2	8.2
Otepää Heights	Limestones	23.7	31.4	46.5	41.9	35.6
	Dolomites	18.4	25.0	30.1	28.0	30.2
	Crystalline	55.5	36.2	21.8	25.2	27.4
	Sandstones	2.4	7.4	1.6	4.9	6.8
Haanja Heights	Limestones	32.6	28.8	41.1	38.8	27.2
	Dolomites	36.5	27.1	31.0	29.2	31.2
	Crystalline	27.3	38.6	26.1	28.9	34.4
	Sandstones	3.6	5.5	1.8	3.1	7.2

Table 12. Average mineral composition of coarse silt (0.05-0.1 mm) fraction of different-aged tills in key sections of southeastern Estonia according to Raukas (Раукас **1978**), %

Minerals		Age				
		$Q_{II}sn$	$Q_{II}ug_1$	$Q_{II}ug_3$	$Q_{III}jr_{vl}$	$Q_{III}jr_{vr}$
Light subfraction (<2.89 g/cm³)	Quartz	86.0	76.9	68.2	74.5	75.3
	Feldspars	12.3	13.3	14.4	13.6	16.6
	Micas	0.2	0.7	0.2	0.6	0.9
	Carbonates	1.5	9.1	17.2	11.3	7.2
Heavy subfraction (>2.89 g/cm³)	Hematite, limonite	5.6	1.9	1.0	1.2	2.2
	Magnetite, ilmenite	34.4	36.6	27.6	25.7	31.2
	Pyrite	0.2	2.5	2.0	2.2	1.8
	Leucoxene	0.6	1.1	0.9	0.9	1.2
	Garnet	21.2	22.6	35.0	34.6	27.6
	Amphiboles	18.1	13.6	16.8	15.2	16.6
	Pyroxenes	2.9	2.1	3.2	2.4	2.3
	Zircon, monazite	9.4	9.2	5.3	5.1	6.9
	Tourmaline	0.6	2.2	0.8	1.0	1.5
	Epidote, zoisite	3.4	2.9	2.6	2.2	2.9
	Staurolite, disthen	0.4	0.6	0.2	0.2	0.5
	Apatite	1.1	2.1	1.7	1.6	1.5
	Dolomite	1.1	3.9	8.8	6.7	2.8
	Rutile, sphene, anatase	1.6	0.8	0.6	1.2	0.5

States and Belarus (Orviku 1960, Орвику 1956, 1960г, Раукас 1978). In the scheme compiled by Kajak *et al.* (Каяк и др. 1976) local geographical names were for the first time used to denote stratigraphic units. Over a period of more than 15 years, the scheme served as a basis for large-scale geological mapping and applied works in the Republic.

On May 6, 1993, a new official stratigraphical chart of Quaternary deposits of Estonia (Table 13) was accepted by the Estonian Stratigraphic Commission (Raukas & Kajak 1995). The scheme was approved as a correlative part of the stratigraphical chart of the Baltic States at the II Stratigraphic Conference in Vilnius (May 9-14, 1993).

Photo 31. Varved clays in the Tsirgulinna quarry, southern Estonia. Bedding disturbances result from recurrent moistening and drying up of clays. *Photo by E. Pirrus.*

In the Quaternary stratigraphy, the age of tills is of special interest as it enables correlation of lithologically similar formations over vast areas (Раукас 1978). The age of tills is generally determined by bedding conditions, by their position with respect to interglacial or interstadial deposits. Unfortunately, the latter are rather uncommon. Besides, the advancing glaciers of the succeeding glaciations crushed most of unconsolidated intermorainic organic deposits which today are often found as erratics embedded in younger sediments. The deposits of the Prangli (Eemian, Mikulinan) interglacial, both continental (Rõngu) and marine (Prangli), serve as key sediments in stratigraphic subdivision and correlation of the Pleistocene cover. The Karuküla (Holsteinian, Likhvian) deposits are more complete in the Karuküla section in southwestern Estonia. The spore and pollen spectrum of all other intermorainic sections is not clear, as these sediments often contain reworked pollen. The most important type sites are shown in Figure 93.

In the Estonian stratigraphical chart (Table 13), lithostratigraphic terms have been used as basic units. As a fundamental unit, formation is used in a meaning of glacial and interglacial episodes in the event stratigraphy. Formations are the three-dimensional sedimentary bodies which have been formed by a specific geological process in the time span of one clear geological event. Big stadial episodes in a meaning of event stratigraphy are comparable with subformations. Using close in the meaning but not synonymous chronostratigraphical (*e.g.* the Prangli Stage), climatostratigraphical (Prangli Interglacial), lithostratigraphical (Prangli Formation) and event stratigraphical (Prangli Interglacial Episode) terms has been avoid. Although interglacial sediments are differentiated on the basis of spore and pollen and other fossil evidence, and pollen assemblage zones underlie their description, for the unity of the scheme, even here lithostratigraphical terms (Prangli and Karuküla formations) were preferred.

Lower Pleistocene deposits are absent in Estonia, and the oldest Middle Pleistocene deposits identified so far in the official chart belong to the Upper Sangaste Subformation. Taking into account some similarity of pollen spectrum of sandy clayey sediments in the Otepää buried valley (Harimägi borehole No. 323 at a depth of 143.3-169.2 m) with the Turgeliai and Beloveže subformations in Lithuania and Belarus (Каяк и Лийвранд 1967), Kajak (1995) includes those beds to the Middle Sangaste Subformation and underlying tills and glaciofluvial deposits (Otepää borehole No. 2 at the depth 123.7-173.7 m) to the Lower Sangaste Subformation. Liivrand (1991) includes all the mentioned sediments to the Järva Formation.

MIDDLE PLEISTOCENE

Sangaste Formation

The Sangaste Formation is correlated with the Dainava Stage in the southern Baltic, the Oka Stage in the European part of Russia and with the Elsterian Stage in Western Europe.

The lowest diamicton unit in Estonia termed as the Upper Sangaste Subformation (named after the Sangaste Parish north-east of the Town of Valga) till, is very compact, brownish, sometimes greenish in colour with indications of shearing. It rests directly upon the bedrock and is found only in the bottommost part of ancient valleys. The thickness of the till bed is small - 15 m at Puiestee, 10.7 m at Sudiste and 5.4 m at Mägiste. Borehole 177 (Puiestee) at a depth of 169.0-207.0 m was chosen for the stratotype section

Table 13. Stratigraphic scheme of Quaternary deposits in Estonia (Raukas & Kajak 1995)

General units			Local units		Palaeontological and lithological characterization	Most important sites
System	Division	Sub-division	Forma-tions	Subfor-mations		
QUATERNARY	HOLOCENE	Separate stratigraphic scheme of Holocene (Flandrian) deposits has been accepted			Variegated continental and marine deposits, 10 assemblage zones	Continental deposits all over Estonia, marine deposits in Lower Estonia
	PLEISTOCENE	Upper Pleistocene	Järva III_{jr}	Võrtsjärve III_{vr}	Grey till in North Estonia, reddish-brown till in South Estonia, aqueoglacial deposits	All over Estonia
				Savala III_{sv}	Dry periglacial vegetation	Savala, Väänа-Jõesuu
				Valgjärve III_{vl}	Grey till in North Estonia, purplish-grey till in South Estonia, aqueoglacial deposits	Valgjärve (Kitse), Kaagjärve, Prangli
				Kelnase III_{kl}	Cryo- and hydrophilous vegetation	Prangli
			Prangli (Rõngu) III_{pr}		Forest vegetation, pollen zones P_2-P_8, marine and continental deposits	Prangli, Kihnu, Rõngu, Küti, Kitse
		Middle Pleistocene	Ugandi II_{ug}	Upper Ugandi II_{ug3}	Brown till in North Estonia, grey till in South Estonia, aqueoglacial deposits	Prangli, Rõngu, Saadjärv, Juminda, Suur-Munamägi
				Middle Ugandi II_{ug2}	Periglacial vegetation	Prangli, Keskküla, Valguta
				Lower Ugandi II_{ug1}	Brown till both in North and South Estonia, aqueoglacial deposits	Prangli, Naissaar, Keskküla, Mägiste, Lanksaare
			Karuküla II_{kr}		Forest vegetation, pollen zones K_I-K_{IV}	Karuküla, Kõrveküla
			Sangaste II_{sn}	Upper Sangaste II_{sn3}	Shaly brownish till in Central and South Estonia	Saadjärv, Keskküla

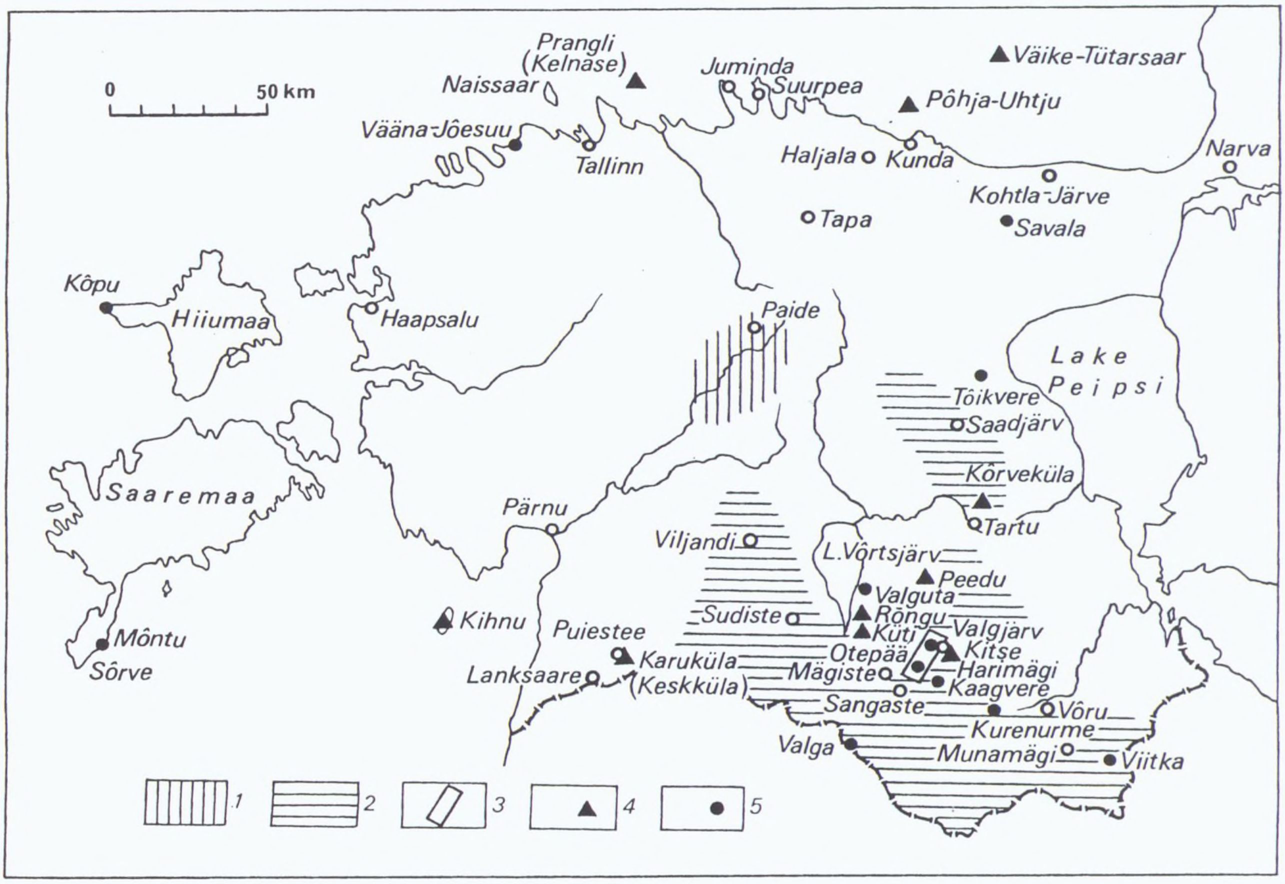


Fig. 93. Areal stratotypes of the Järva (1), Ugandi (2) and Sangaste (3) glacial formations and location of the main interglacial (4) and interstadial (5) sequences.

(Раукас и др. 1993). The clast composition is different: in southwestern Estonia crystalline rocks are clearly prevailing (up to 95%), but in southeastern Estonia their amount is only 25-60%.

The high content of Vyborg rapakivi and Suursaari quartz porphyries in southeastern Estonia and the absence or a very low content of rapakivi from southwestern Finland suggest that the deposition of till was due to the southward flowing ice. The poorly sorted diamicton is richer in clay particles than the uppermost till units. The latter are high in kaolinite (up to 30-35%) derived from the weathered bedrock. Due to the influence of Devonian sand- and siltstones, the sand and silt fractions of till are richer in quartz and contain less feldspars and carbonates than other till units (Раукас 1978).

According to bedding conditions (Kajak 1995), to the Upper Sangaste Subformation belong grey and brownish till beds and glaciofluvial deposits below organogenous bog and lacustrine deposits in the Karuküla and Kõrveküla sections, up to 23 m in thickness.

Karuküla Formation

The Karuküla Formation (interglacial) is palynologically correlated with the Butenai Stage (interglacial) in the southern Baltic, the Likhvinan Stage in the European part of Russia and the Holsteinian of Western Europe.

The Karuküla type site is situated in southwestern Estonia, in the Pärnu County, about seven kilometres south of the Town of Kilingi-Nõmme (Fig. 93). It displays continental deposits and was first described by Orviku (1944). The name of the stratotype proposed by Kajak *et al.* (Каяк и др. 1976) is inaccurate because the section is actually situated in the Keskküla Village. Due to the rather long history of investigations and wide recognition of the site, changing of the stratotype's name was considered unpurposeful.

The Middle Pleistocene (Likhvinan, Holsteinian) age of the section was first suggested by Danilans (Данилас 1966) and Voznyachuk (Вознячук 1966) and later established by Liivrand (1984, Величкевич и Лийвранд 1976, 1984).

The information currently available on the Karuküla site has been derived through the study of about 70 boreholes and excavations. The interglacial deposits are probably of allochtonous bedding (Левков и Лийвранд 1988). There seem to be three large and two small erratics and two lumps of Holsteinian deposits within one stratigraphic level measuring 105 m horizontally and 3.25 m vertically (Liivrand 1991). The Karuküla section and its palaeobotanical characteristics have been described in detail in several publications (Лийвранд 1972, 1990, Liivrand 1984, 1991). Another well investigated site of the Karuküla Formation is at Kõrveküla near Tartu (Лийвранд и Саарсе 1983).

Ugandi Formation

The Ugandi Formation, called after an ancient South-Estonian and North-Latvian area, where those deposits are most widespread, is correlated with the Žeiminiai Formation in Lithuania, Kurzeme in Latvia, Middle Russian in Russia and Saale in Western Europe. In some places Middle Ugandi interstadial beds have been described. Borehole 6 on Prangli Island (depth 75.5-123.0 m) and borehole 268 at Valguta (13.1-35.0 m) have been established as the unit and boundary stratotypes for northern and southern Estonia, respectively (Раукас и др.1993).

The till of the L o w e r U g a n d i S u b f o r m a t i o n, which is correlated with the Dniepr till in Russia and the Žemaitia till in Lithuania, is reddish-brown both in northern (Prangli, Naissaar, Suurpea) and southern Estonia (Mägiste, Lanksaare, Sudiste) and up to 50 m thick (Mägiste). The till is compact and occurs in uplands mainly in sheltered position or rests in ancient valleys. The clast lithology (high content of Vyborg rapakivi in eastern Estonia) indicates that the Lower Ugandi till was deposited by the southward flowing ice (Раукас 1978). Clasts in northern Estonia are almost completely fragments of crystalline rocks, whereas in southern Estonia their composition reflects both the local provenance (up to 10% of local Devonian sand- and siltstones) and the influence of the outcropping carbonaceous rocks on the way of the moving ice (50-60% carbonaceous clasts). Of all Estonian tills, it has the highest content of sandy fraction. In clay fraction illite (50-70%) prevails, but also the content of kaolinite is rather high (20-45%).

M i d d l e U g a n d i sands, silts, loams and sandy loams often contain rebedded pollen and their stratigraphic position is not clear (Liivrand 1991).

U p p e r U g a n d i till is massive to slightly stratified, reddish-brown in the fore-klint area (Prangli, Juminda) and grey in northern (Sõrve, Saadjärv) and southern (Rõngu, Suur-Munamägi) Estonia. The till unit is up to 70 m thick (Suur-Munamägi) and often cemented with carbonates. According to its composition (absence of Vyborg rapakivi and quartz-porphyries from the Island of Suursaari), the till entrained by southeastward flowing ice. Practically all the clasts in the fore-klint area originate in the crystalline basement, but in other areas carbonate rocks prevail (65-80%). In southern Estonia, this till has the highest content of silt particles and the lowest content of Devonian material. Illite (65-80%) prevails and the content of kaolinite (10-20%) is low in the clay fraction.

The aqueoglacial deposits of the Lower (Puiestee 60 m) and Upper (Vääna-Jõesuu 60 m) Ugandi subformations are rather thick and variable in composition (Раукас 1978).

UPPER PLEISTOCENE

Prangli (Rõngu) Formation

The Upper Pleistocene in Estonia begins with the well-known Eemian interglacial deposits in Western Europe and Mikulinan in Eastern Europe. In the Regional Scheme of the Baltic area this interglacial is called the Merkine Interglacial after the town in southeastern Lithuania. The Eemian (Mikulinan) deposits, both continental (Rõngu) and marine (Prangli), correlated on the basis of the pollen assemblage zones, are in good stratigraphical agreement (Liivrand 1991).

The continental Eemian deposits at Rõngu were investigated in particular detail about half a century ago (Orviku 1939, Thomson 1939a, 1941). Later, complementary investigations were carried out in several other sections (Küti, Kitse, Peedu) by Liivrand (Лийвранд 1977) and Kajak (1995).

In the sixties, marine Eemian deposits were found on Prangli Island in the Gulf of Finland (Kajak 1961) and subject to palynological (Лийвранд и Вальт 1966, Liivrand 1974, 1991, Лийвранд 1987, 1990) and diatom (Черемисинова 1961) studies. Later, marine deposits of the Prangli Formation were found in several places (Põhja-Lehtju, Väike Tütarsaar, Kihnu a.o.).

A stratotype section at a depth of 67.6-75.5 m in borehole 6 on Prangli Island and a parastratotype for the continental deposits in borehole 264 (2.3-7.8 m) and excavation II (2.0-5.8 m) on the lands of the Vaeva Farm, 2 km west of Rõngu, were established for the Prangli Formation (Раукас и др. 1993). The name for the formation was proposed by Kajak *et al.* (Каяк и др. 1976).

Järva Formation

The name of the formation was proposed by Kajak *et al.* in 1976 (Каяк и др. 1976) after the Järva County in central Estonia where a typical grey till of the last glaciation is widespread in the drumlins and lowland near the Town of Paide. The Järva Formation is correlated with the Nemunas Formation in Lithuania, the Baltia in Latvia, the Valdaian Stage in Russia and the Weichselian in Western Europe. The Vääna-Jõesuu (13-70 m) and Kitse boreholes (0-31.1 m) were chosen for stratotype sections in northern and southern Estonia, respectively (Раукас и др. 1993).

The Kelnase beds were named after the village on Prangli Island. In the Prangli section, they are represented by clayey silts with the pollen spectra characterized by an increasing quantity of *Betula nana* (40-80%) and herbs (tundra species). *Gramineae* and *Cyperaceae* are common. *Selaginella selaginoides, Lycopodium alpinum* and *Artemisia arctica* are present. A cryophilous and hydrophilous vegetation refers to the approaching glacial advance (Каяк и др. 1976, Liivrand 1991).

The Valgjärve beds, named after the lake in southern Estonia, are represented by grey till in northern and purplish-grey till and related aqueoglacial deposits in southern Estonia. The purplish-grey till was proposed for a specific stratigraphical unit by Orviku (1939) and described lithologically by Orviku (Орвику 1958a) and Raukas (Раукас 1963a, 1978). In the Kitse borehole No. 19 near Lake Valgjärv at a depth of 4.2-31.1 m, the till of the Valgjärve bed covers the organogenous deposits of the Prangli (Rõngu) Formation (Kajak 1995).

The Savala beds named after the village in northeastern Estonia belong to the M i d d l e J ä r v a S u b f o r m a t i o n (Каяк и др. 1976). The type section (borehole 7854, depth 25.8-30.2 m) is situated in the Savala ancient valley about 120 km east of Tallinn. It is mainly filled with grey-coloured varved clays. The pollen and spore composition of the intermorainic layer suggests dry periglacial conditions (Лийвранд 1986, Liivrand 1991). The Savala interstadial warming was not accompanied by any substantial development of forests.

The Võrtsjärve beds, named after Lake Võrtsjärv, are represented mainly by tills of different colour of the last glaciation and aqueoglacial deposits above and beneath the till. In several places some till layers with thin intermorainic interstadial or interphasial sediments occur (Orviku 1939, Раукас 1963a). Tills of the last glaciation on the Cambrian blue clays, sand- and siltstones in the fore-klint area are bluish-grey, mostly clayey and contain mainly clasts from Finland and the bottom of the Gulf of Finland. On the crystalline basement, the till is brown or reddish-brown. Stony tills on the

Ordovician and Silurian bedrock are enriched with the local carbonaceous material (Photo 29). The constituent clasts are mainly angular. Tills on the Devonian sand- and siltstones are reddish-brown. The rather well-rounded local carbonaceous and erratic crystalline material occurs in various ratios in the cobble and pebble fractions and are under the influence of the Devonian bedrock, comparatively rich in sand and silt fractions (Орвику 1958а, Раукас 1978). In the stratotype area - the basin of Lake Võrtsjärv, both grey carbonaceous (Valma) and reddish-brown (Tamme) tills are widespread.

Stratigraphy of Late-glacial deposits

The Upper Järva Late-glacial deposits are divided into Arctic (Oldest Dryas, Bølling, Older Dryas) and Subarctic (Allerød, Younger Dryas) chronozones (Table 14). According to the decision of the INQUA Congress in Paris in 1969, the Holocene/Pleistocene boundary is accepted as 10,000 ^{14}C years.

Traditionally, the Late-glacial interval in Estonia starts from the accumulation of Rauna interstadial deposits in central Latvia (Каяк и др. 1976). In the Raunis section, interstadial sands with alternating layers of silt and clay, which contain peat and plant remains, lie between two layers of till to the southeast from the Town of Cēsis, on the right bank of the Raunis River. Organic remains from the Raunis section have been dated in several laboratories (13,390±500: Mo-196; 13,250±160: TA-177; 13,320±250: RI-39) and the results obtained are in good agreement (Пуннинг и др. 1968).

In mainland Estonia and on the islands of the Gulf of Finland, Eemian (Prangli) deposits or pre-Weichselian tills are in some places (*e.g.* Prangli Island) overlain by four till beds, the exact age of which is uncertain. The upper till beds are regarded as stadial ones of Late Weichselian age representing secondary oscillations of the ice sheet. It is also possible that a two- or three-layered till beds may consist of contemporaneous basal and ablation tills from a single glacial event (Раукас 1963а). Locally, Haanja/Otepää and Pandivere/Palivere tills are separated by terrigenous layers containing subfossil molluscs (Kameri, in Orviku 1939) and pollen assemblages of a cool character. Tills of the Haanja, Otepää, Pandivere and Palivere stadials have specific colour and lithological composition (Раукас 1963а, 1978) and can be regarded as lithostratigraphical units of the lowest taxonomical rank (beds).

In some places intermorainic layers have been dated by the ^{14}C method, but the results are contradictory. In the Kurenurme section, southeastern Estonia, remains of *Salix* wood were taken from sandy loam overlying Haanja till (Ильвес и др. 1974). Quite reliable radiocarbon datings of the wood (12,650±520: TA-57) and organic detritus (12,420±100: Tln-35) indicate that these deposits accumulated at the beginning of the Bølling Interstadial. Unfortunately, the process of the deposition of the organic material is not clear (Карукяпп 1985) and this hampers the usage of the section in the till stratigraphy. In the Kaagvere section southeast of Otepää, the dates obtained on the reddish-brown till (15,150±575: TA-50, >30,000: TA-36) suggest redeposition of older interglacial material. The palynological characteristics of interstadial layers between the stadial till beds are not clear either. Probably, these layers contain a lot of material redeposited from older interglacials (Liivrand 1991).

Palynological studies of pre-Allerød deposits above the till beds in Estonia give evidence of severe climatic conditions throughout the Arctic period. They do not permit the layers related to the Bølling amelioration to be distinguished. Such deposits may be present in southern but hardly in northern Estonia. In the section of Haljala (Мянниль и Пиррус 1963), a pollen assemblage suggesting a brief interval of warming, possibly Bølling, has been reported from a sandy interlayer at a depth of 10.5-11.2 m. However, as its redeposition in the sandy sediment is not excluded, the kind of pollen data interpretation must be taken with great caution (Пиррус и Раукас 1969), and the more that no corresponding warm interval is known from any other site in Estonia.

Deposits of Older Dryas age occur both in northern and in southern Estonia. The lower boundary of the Older Dryas is undefined in Estonia (Каяк и др. 1976), but probably it is the boundary between the Otepää and Kurenurme beds.

Table 14. Stratigraphical scheme of Estonian Late-glacial deposits (Pirrus & Raukas, 1996)

<table>
<tr><th>Chronological scale, yr BP</th><th>Main subdivisions</th><th>Sub-formations</th><th colspan="2">Chrono-zones</th><th colspan="2">Index</th><th>Definition of boundaries, yr BP</th><th>Palynozone</th><th>Index</th><th>Baltic Sea stages</th><th>Ice marginal formations</th></tr>
<tr><td rowspan="8">10 000 —
10 500 —
11 000 —
11 500 —
12 000 —
12 500 —
13 000 —
13 500 —</td><td rowspan="2">Holo-cene</td><td rowspan="2">Lower Holo-cene</td><td colspan="2" rowspan="2">Preboreal</td><td rowspan="2">PB</td><td>PB_2</td><td rowspan="8">9 500 —
10 000 —
10 800 —
11 300 —
11 800 —
12 200 —
13 200 —</td><td>*Betula*</td><td>B</td><td>Ancylus Lake</td><td rowspan="8">Palivere zone
(11 200+430 yr BP)
Pandivere zone
(12 050+430 yr BP)
Sakala zone
(12 250+430 yr BP)
Otepää zone
(12 600+430 yr BP)
Haanja zone
(13 000-13 200
+430 yr BP)</td></tr>
<tr><td>PB_1</td><td>*Betula-Pinus*</td><td>B-P</td><td rowspan="2">Yoldia "sea"</td></tr>
<tr><td rowspan="6">Pleistocene</td><td rowspan="6">Võrtsjärve</td><td rowspan="3">Subarctic</td><td>Younger Dryas</td><td colspan="2">DR_3</td><td>*Artemisia - Betula nana*</td><td>Ar-Bn</td></tr>
<tr><td rowspan="2">Allerød</td><td rowspan="2">AL</td><td>AL_b</td><td>*Pinus*</td><td>P</td><td rowspan="3">Baltic Ice Lake</td></tr>
<tr><td>AL_a</td><td>*Pinus-Betula*</td><td>P-B</td></tr>
<tr><td rowspan="3">Arctic</td><td>Older Dryas</td><td colspan="2">DR_2</td><td>*Artemisia Chenepodiaceae*</td><td>Ar-Ch</td></tr>
<tr><td>Bølling</td><td colspan="2">BØ</td><td>*Betula-Cyperaceae*</td><td>B-Cy</td><td rowspan="2">Local ice Lakes</td></tr>
<tr><td>Oldest Dryas</td><td colspan="2">DR_1</td><td></td><td></td></tr>
</table>

In the light of the pollen evidence, the retreat of the ice margin from the Haanja position started during the transitional from Oldest Dryas to Bølling time and the deglaciation of the Estonian territory was completed during the second half of the Allerød (Пиррус и Раукас 1969).

According to Reet Pirrus (Pirrus & Raukas 1996), some more or less clear trends in the vegetation history could be given (Table 15).

OLDER DRYAS CHRONOZONE (DR_2)

***Artemisia - Chenopodiaceae* palynozone**

The Older Dryas Chronozone is represented by glaciolacustrial varved clays or rhythmically laminated silts and sands and overlain by lacustrine silts and clays. In the southern part of Estonia, minerogenic lacustrine sediments may contain minor amounts of plant remains. The thickness of deposits ranges from 1.3 to 11.3 m. In the Older Dryas about 12,000 years ago the Baltic Ice Lake formed and corresponding deposits started to accumulate.

This zone is characterized by high herb pollen percentages (*Artemisia, Chenopodiaceae, Helianthemum, Cyperaceae, Gramineae,* and several other species of primary vegetation) along with *Betula nana* L.

ALLERØD CHRONOZONE (AL)

The Allerød Chronozone is represented by lacustrine clays and silts (0.15-1.85 m in thickness) with blackish-grey interlayers and the Baltic Ice Lake sediments. Scattered plant remains, mostly leaves and stalks of *Bryales* moss are common in lake sediments.

The Allerød Chronozone is subdivided into two pollen zones (Pirrus & Raukas 1996): a) *Pinus-Betula* Zone (ALa), b) *Pinus* Zone (ALb).

The lower boundary of AL Chronozone is fixed with a rather distinct increase of AP pollen and decrease of herbs (*Artemisia, Chenopodiaceae*) and *Betula nana* L.

Characteristic of AL Chronozone is the prevalence of tree pollen. *Betula* shows a rapid increase and towards the uppermost part of the zone *Pinus* increases distinctly and has its Late-glacial culmination. At the same time, herb pollen is at its minimum. *Betula nana* L. is constantly present in low percentages. The variety of bog and meadow species of terrestrial herbs and water plants has increased. Xerophytes, halophytes, heliophytes and tundra plants are continuously present, but in low values. Fine preservation and abundance of pollen as well as the regularity of pollen curves indicate their bedding *in situ*.

YOUNGER DRYAS CHRONOZONE (DR_3)

***Artemisia - Betula nana* Zone**

The younger Dryas Chronozone is represented with the Baltic Ice Lake and Yoldia Sea sediments and by lacustrine silts and clays, often with fine-grained sand interlayers. *Bryales* remains are scattered or occur as thin layers, occasionally abounding in hydrotroilite. The thickness of lacustrine deposits ranges from 0.2 to 4.0 m.

The zone boundary AL/DR_3 is placed at the strong and rapid increase of the content of herb pollen (particularly *Artemisia*) and *Betula nana* L. This zone is characterized by remarkably high frequency of herb pollen ranging from 40-60%. Maximum values of *Betula nana* L. pollen in different profiles range from 20 to 25%. The Late-glacial culmination

Table 15. Pollen stratigraphy of Estonian Late-glacial deposits (compiled by R. Pirrus)

Sub-stage	Chronozone		Pollen Zone	Index		Palynological characterization	Characteristic deposits	Type sections
Upper Järva (Võrtsjärve)	Sub-Arctic	Younger Dryas	*Artemisia—Betula nana*	DR_3		Herbs dominate (40—60%). *Artemisia* content in S Estonia is more than 50%, in N Estonia 70% (up to 170%), Chenopodiaceae amount to 19 and 25%, respectively. *Betula nana* L. reaches 25%. *Betula* dominates in S Estonia, *Pinus* in N Estonia. Slight culmination of *Picea* occurs either in the lower or upper part of the zone.	Silts and clays, often with interlayers of fine-grained sand and *Bryales* remains.	Remmeski Visusti Haljala Kunda
		Allerød	*Pinus*	b	AL	Trees (*Pinus, Betula,* less *Picea*) dominate. b) *Pinus* has Late-glacial culmination (80—90%). Herbs amount to 1—5%, in N Estonia to 15%, *Betula nana* L., some %.	Clays and silts comprising plant remains and interlayers rich in troilite	Remmeski Visusti Haljala
			Pinus—Betula	a		a) *Betula* frequency is increased and it may prevail over *Pinus*. Herbs form 18—20%, *Artemisia* 7—23%, *Betula nana* L. 4—7%.		
	Arctic	Older Dryas	*Artemisia*—Chenopodiaceae	DR_2		High values of herbs (20—30%). In S Estonia up to 37—43%. Chenopodiaceae form 5—10% (up to 29%), *Artemisia* amounts to 44%. Xerophilous, halophilous, and tundra species of pioneer plants are common. Among trees *Pinus* dominates, in lower part — *Betula*. High values of pollen (*Alnus, Corylus, Picea, Q. mix.*, partly *Betula* and *Pinus*) rebedded from till, have been identified.	Varved clays, silts and sands	Remmeski Visusti

of *Picea* is either in the lowermost (Võru, Visusti, Haljala) or uppermost (Remmeski, Loobu) part of the pollen zone.

The boundary DR_3/PB is placed at a rapid increase of tree pollen, prevailingly *Betula* (about 80%, in SE Estonia up to 90%) and *Pinus* (about 20%).

Holocene deposits and their stratigraphical subdivision

The Holocene continental deposits, occasionally rather thick, occur practically everywhere above the Pleistocene deposits. Unfortunately, the offshore and nearshore marine deposits are characterized by numerous unconformities and rapid facies changes and in many sequences gaps cover longer time spans than the preserved strata. The main stages in the Baltic Sea history are known from the very beginning of the century, but they have never been properly defined as stratigraphical units (Хюваринен и Раукас 1992). Therefore, the stratigraphical scheme of the Holocene deposits (Table 16) is mainly based on the continental deposits. The existence of the four major phases in the postglacial history of the Baltic – the Preboreal Yoldia "Sea", the Ancylus Lake, the Litorina Sea and the Limnea Sea – is recognised.

In Estonia, there are 9836 peat bogs and about 1150 lakes greater than 1 ha in area (Orru 1992, Mäemets 1976). The peat is at its thickest (16.8 m) in the small Vällamäe kettle hole. The peat deposits are usually 8-10 m thick. The greatest thickness of organic lacustrine deposits is 18 m (Väimela-Alajärv), lake marl 6-7 m (Tapa, Kulina), travertine 5-6 m (Loosi, Rõuge), alluvial deposits 15 m (Väike-Emajõgi) and aeolian deposits 15-20 m (Sininõmme, Kõpu, Rannametsa). In the first half of the century, the palynological approach (Thomson 1925) was applied to the stratigraphical studies in Estonia, at the end of the fifties, physical dating methods were taken into use (Ильвес и др. 1974). P. Thomson investigated lake and mire deposits in about 20 localities and modified his first (Thomson 1925) Estonian Holocene stratigraphical scheme in several high standard publications (Thomson 1926, 1929, 1930, 1933). Some 30 years later K. and L. Orviku published the next Holocene stratigraphical scheme (Орвику 1956, Орвику Л. Ф. 1960).

The first official stratigraphical chart of the Estonian Holocene deposits was compiled under the leadership of Prof. K. Orviku and accepted in May 1976 (Каяк и др. 1976). The second official stratigraphical chart presented in this book (Table 16), was compiled by R. Pirrus, A. Raukas and S. Veski (Raukas *et al.* 1995b). The part of the scheme dealing with continental deposits is based on the studies by H. Kessel, R. Pirrus, A. Sarv, L. Saarse, K. Kimmel, T. Koff, L.-K. Königsson, S. Veski and A. Poska. The investigations of J. Lutt, H. Kessel and A. Raukas underlie the subdivision of marine deposits. The scheme was accepted at the session of the Estonian Stratigraphical Commission on May 6, 1993 and a week later it was approved at the Stratigraphical Conference of the Baltic States in Vilnius. The regional chart for the Baltic States was approved at the same time. They both followed the Scandinavian scheme (Mangerud *et al.* 1974). The charts have parallel subdivisions for the continental and marine deposits.

According to the international rules, stratigraphical charts should be based on the unit and boundary stratotypes. Unfortunately, up to now there are no officially accepted stratotype

Table 16. The stratigraphical scheme of the Holocene deposits in Estonia (Raukas *et al.* 1995b)

Chronological scale 10^3 years	Stage	Substage	Chronozone	Index	Index	Definition of boundaries in years BP	Pollen assemblage zones	Index	Index (after Lennart von Post)	Baltic stages	Definition of boundaries years BP
1	Holocene	Upper	Sub-Atlantic	SA	SA3	1000	*Pinus-Betula*	*P-B*	I	Limnea	
2	Holocene	Upper	Sub-Atlantic	SA	SA2	2000	*Betula-Pinus-Picea*	*B-P-Pc*	IIa	Limnea	
	Holocene	Upper	Sub-Atlantic	SA	SA1	2500	*Betula-Alnus*	*B-A*	IIb	Limnea	
3, 4	Holocene	Middle	Sub-Boreal	SB	SB2	4000	*Picea*	*Pc*	III	Limnea	4000
5	Holocene	Middle	Sub-Boreal	SB	SB1	5000	*Quercus*	*Q*	IV	Litorina	
6	Holocene	Middle	Atlantic	AT	AT2	6500	*Tilia-Ulmus-Fraxinus*	*T-U-Fr*	V	Litorina	
7, 8	Holocene	Middle	Atlantic	AT	AT1	8000	*Ulmus-Corylus*	*U-Co*	VI	Litorina	8000
	Holocene	Lower	Boreal	BO	BO2	8500	*Pinus-Alnus*	*P-A*	VII	Ancylus	
9	Holocene	Lower	Boreal	BO	BO1	9000	*Pinus*	*P*	VIII	Ancylus	
	Holocene	Lower	Pre-Boreal	PB	PB2	9500	*Betula*	*B*	IXa	Ancylus	9300
10	Holocene	Lower	Pre-Boreal	PB	PB1	10000	*Betula-Pinus*	*B-P*	IXb	Yoldia	
11	Pleistocene	Upper Järva	Subarctic: Y Dryas	DR3	DR3	10800	*Artemisia-Betula nana*	*Ar-Bn*	X	Yoldia	10300
	Pleistocene	Upper Järva	Subarctic: Allerød	AL	ALb	11300	*Pinus*	*P*	XIa	Baltic Ice Lake	
	Pleistocene	Upper Järva	Subarctic: Allerød	AL	ALa	11800	*Pinus-Betula*	*P-B*	XIb	Baltic Ice Lake	
12	Pleistocene	Upper Järva	Arctic: OD		DR2	12200	*Artemisia*-Chenopodiaceae	*Ar*-Ch	XIIa	Baltic Ice Lake	12000
	Pleistocene	Upper Järva	Arctic: Bölling	BÖ	BÖ		*Betula*-Cyperaceae	*B*-Cyp	XIIb	local ice lakes	

sections for Holocene deposits in Estonia or in the other Baltic States. This work is in progress.

Each site with its own local pollen assemblage biozones is effectively its own stratotype, but no stratotype can exist for the regional pollen assemblage biozones, which are artificial synthesis (Turner 1989). The same type of artificial synthesis is the proposed local stratigraphical chart (Table 16), based on the multiple sections throughout Estonia, all having their own characteristics. As the pollen zones are time transgressive, the boundaries between palynological chronozones have not been drawn and this makes the chart useful and applicable in Estonia as a whole.

VI HYDROGEOLOGY

WATER-BEARING FORMATION

Basic data

In terms of groundwater formation and circulation, the groundwater system in Estonia can be divided into three principal hydrostratigraphical units.

1. T h e Q u a t e r n a r y d e p o s i t s. The sandy and clayey Quaternary deposits and peat form porous aquifers with mainly unconfined groundwater which are directly influenced by meteorological conditions. The whole infiltration water percolates into the Quaternary cover and the greater part of groundwater discharge flows through it. The upper portion of the Quaternary cover or sporadically all Quaternary deposits belong to the aeration zone where a lot of water circulates by the agency of capillary force or evaporates, in addition to the filtration flow.

2. T h e b e d r o c k. The terrigeneous and carbonate Palaeozoic and Proterozoic rocks form porous, fissured and karstified, mostly confined aquifers, which are isolated from each other with aquitards of different isolation capacity. In the karst cavities near the ground, shallow groundwater flows very fast and its chemical composition is close to that of the surface water. However, the deeper strata contain pre-Quaternary groundwater, which is high in total dissolved solids (TDS) and moves very slowly under natural conditions.

3. T h e c r y s t a l l i n e b a s e m e n t. Predominantly pre-Quaternary groundwater in the fissures of igneous and metamorphic rocks contains a high rate of TDS and under natural conditions it is sporadically almost stagnant. The lower portion of the crystalline basement serves as an aqifuge for the whole overlying water-bearing formation in Estonia (to the exclusion of the hypothesis about water originating and arising from the depths of the Earth's crust).

A q u i f e r and a q u i t a r d are the units of detailed hydrogeological stratification of the water-bearing formation. Aquifer is a relatively homogeneous water-bearing layer or rock with similar water conductivity and storage capacity yielding water in a useable quantity to a well.

According to the hydraulic conductivity value K*, the degree of the permeability of water-bearing strata is the following:

$10^2 \leq K < 1$ very low
$1 \leq K < 3$ low
$3 \leq K < 10$ medium
$10 \leq K < 30$ high
$30 \leq K < 70$ very high
$K > 70$ extremely high

Permeability in a lateral direction can be up to 100 times higher than in a transversal direction.

Aquitards are the strata, the transversal conductivity K_t of which is generally less than 10^{-2} m/d. The following degrees of impermeability can be distinguished:

$10^{-2} > K > 10^{-4}$ weak
$10^{-4} \geq K_t > 10^{-6}$ medium
$10^{-6} \geq K_t > 10^{-8}$ strong
$K_t < 10^{-8}$ very strong

Not a single aquitard with the above-mentioned filtration characteristics has an absolute isolation ability. According to this gradation, even strong aquitards are permeable to up- or downward groundwater flows, the total amount of which in large areas can extend up to 10^4 m³/d.

Aquifers which lie one over another are not necessarily isolated with aquitards. The rocky complex consisting of aquifers and aquitards with similar hydraulic characteristics but with different rock types is termed aquifer system.

In terms of the real water supply, the aquifers and aquifer systems can be subdivided into sufficiently water yielding aquifers and aquifer systems (with specific capacity of wells correspondingly $q > 0.1$ l/(s×m) ≈ 10 m³/(d×m), $K > 1$ m/d) and weakly water yielding aquifers and aquifer systems ($q < 0.1$ l/(s×m), $K < 1$ m/d). Aquifers and aquifer systems can be sufficiently or weakly water yielding either sporadically (locally) or in the whole distribution area. According to the aforenamed criteria, at least 23 aquifers and 4 regional aquitards can be distinguished in the water-bearing formation of Estonia (Table 17, Figs. 94, 95).

Water in the Quaternary cover

The technogeneous deposits (tQ_{IV}) in settlements mostly consist of stuff and building waste. Besides, there are 495 dumps of different size in Estonia. In the mining region extensive areas are under spoil heaps and oil shale plateaus. The water in technogeneous deposits is usually highly polluted. The water leaking through ash hills and dumps is dangerous to the environment.

The boggy deposits (bQ_{IV}) are mostly represented by peat. Under natural conditions, the water generally lies at a depth of 0.1...0.5 m, being deeper only in the fields of milled peat. The conductivity of peat is 0.3...1 m/d. The inflow of water into experimental pits amounts to 1...10 m³/d per 0.5...1 m of drawdown. Bogs recharge from precipitation, while the replenishment to fens is also from lateral flows of unconfined groundwater and vertical flows of confined groundwater from deeper strata. The water of boggy deposits has a nasty taste and smell and is practically not used for the water supply.

The aeolian deposits (vQ_{IV}) are mainly represented by fine-grained and well- sorted sands of dunes on the northern coast of Lake Peipsi and on ancient and present beaches of the Baltic Sea. Due to its chain-like morphology, the upper and greater portion of a dune is usually dry; water occurs in the lower part at a depth of 10...15 m from the surface. The yield of wells does not usually exceed 1 to 5 m³/d. At the sea coast the occasional intrusion of brackish sea water can take place.

The lacustrine deposits (lQ_{IV}) occur in limited areas in intermittent strata of loamy sand, loam and sapropel mostly in the Alutaguse and Võrtsjärve lowlands. The strata are poorly permeable and not suitable for water supply.

The alluvial deposits (aQ_{IV}) are represented by gravel, sand, loamy sand and loam of river valleys with a total thickness of up to 15 m. Due to the limited distribution, they do not have any practical importance.

The marine deposits (mQ_{IV}) are up to six, occasionally even more metres thick and consist mostly of sand and coastal gravel which are found in Lower Estonia. In places, the water level can lie close to the ground. Water can sometimes be

* The dimension of K is m/d [LT^{-1}].

Table 17. Subdivision of water-bearing formation of Estonia

Standard scheme: System	Local units: Formation, Stage	Unit of hydrogeological stratigraphy: Aquifer system	Aquifer		Aquitard		Thickness, m	Depth to the groundwater table or piezometric surface[1], m	Well discharge, l/s	Drawdown, m	Specific capacity, l/(s×m)
Quaternary Q	Holocene	Quaternary (Q)	technogeneous deposits (tQ_{IV})								
			boggy deposits (bQ_{IV})				6-8	0.1-1.5	0.01-0.1	0.5-1.0	< 0.1
			aeolian deposits (vQ_{IV})				5-10	2-10	0.01-0.06	0.3-1.0	< 0.1
			lacustrine deposits (lQ_{IV})				0.5-5	0.5-2	0.01-0.2	0.2-1.2	< 0.1
			alluvial deposits (aQ_{IV})				2-6	0.1-2	0.02-0.1	0.4-0.8	< 0.2
			marine deposits (mQ_{IV})				3-5	1-3	0.1-0.5	0.3-0.7	0.3-0.6
	Järva		glaciolacustrine deposits (lgQ_{III})				5-15	2-4	0.02-0.3	0.5-1.0	< 0.2
	Ugandi				glaciolacustrine varved clay (lgQ_{III})		5-10	★	★	★	★
	Sangaste		glaciofluvial deposits (fQ_{III-II})				5-50	+5-7	0.1-8	0.2-10	0.4-10.0
			sporadically waterbearing glaciogenous deposits (gQ_{III})		loamy glaciogenous deposits (gQ_{III})		2-10	1-4	0.01-0.2	0.5-1	< 0.1
Devonian D	Daugava, Dubniki, Pļaviņas	Upper Devonian (D_3)	undistinguished				17-25	3-8	2-15	5-10	0.1-6
					Snetnaja Gora - Amata (D_3sn-D_2am)		8-10	★	★	★	★
	Amata, Gauja, Burtnieki, Aruküla	Middle Devonian (D_2)	undistinguished				50-250	+1-20	3-8	3-7	0.4-1.0
	Narva	Middle-Lower Devonian (D_{2-1})	undistinguished		Narva regional (D_2nr)		40-100	★	★	★	★
	Pärnu, Rēzekne, Tilžė						30-80	+1-30	5-10	4-8	0.5-2.0
Silurian S		Silurian-Ordovician (S-O)	undistinguished				50-100	0-10	5-10	2-3	3-5
Ordovician O					Silurian-Ordovician regional (S-O)		10-300	★	★	★	< 0.05

Table 17 continued:

Standard scheme: System	Local units: Formation, Stage	Unit of hydrogeological stratigraphy: Aquifer system	Aquifer		Aquitard		Thickness, m	Depth to the groundwater table or piezometric surface[1], m	Well discharge, l/s	Drawdown, m	Specific capacity, l/(s×m)
Cambrian Є	Vērgale	Ordovician-Cambrian (O-Є)	Ruhnu ($Є_2rh$); Soela-Tiskre ($Є_1sl$-ts)	Kallavere-Tiskre (O_1kl-$Є_1ts$)	Irben ($Є_1ir$)		20-60	+2.5-50	5-7	10-15	0.2-0.4
	Lükati				Lükati-Sõru ($Є_1lk$-sr)	Lükati-Lontova regional ($Є_1lk$-ln)	40-50	★	★	★	★
	Lontova	Cambrian-Vendian (Є-V)	Voosi ($Є_1vs$)								
Vendian V	Voronka		Voronka-Gdov (V_2vr-gd)	Voronka (V_2vr)			20-40	5-80	3-6	1-10	0.2-5.8
	Kotlin					Kotlin (V_2kt)	30-50				
	Gdov			Gdov (V_2gd)			40-65	50-100	1-5	2-10	1.5-2.5
Crystalline basement PR_1		The fissured zone of crystalline basement (PR_1)					20-50	+14-56	0.1-1.0	10-60	0.01-0.1

[1]the sign + means a piezometric surface above the ground surface

* undetermined

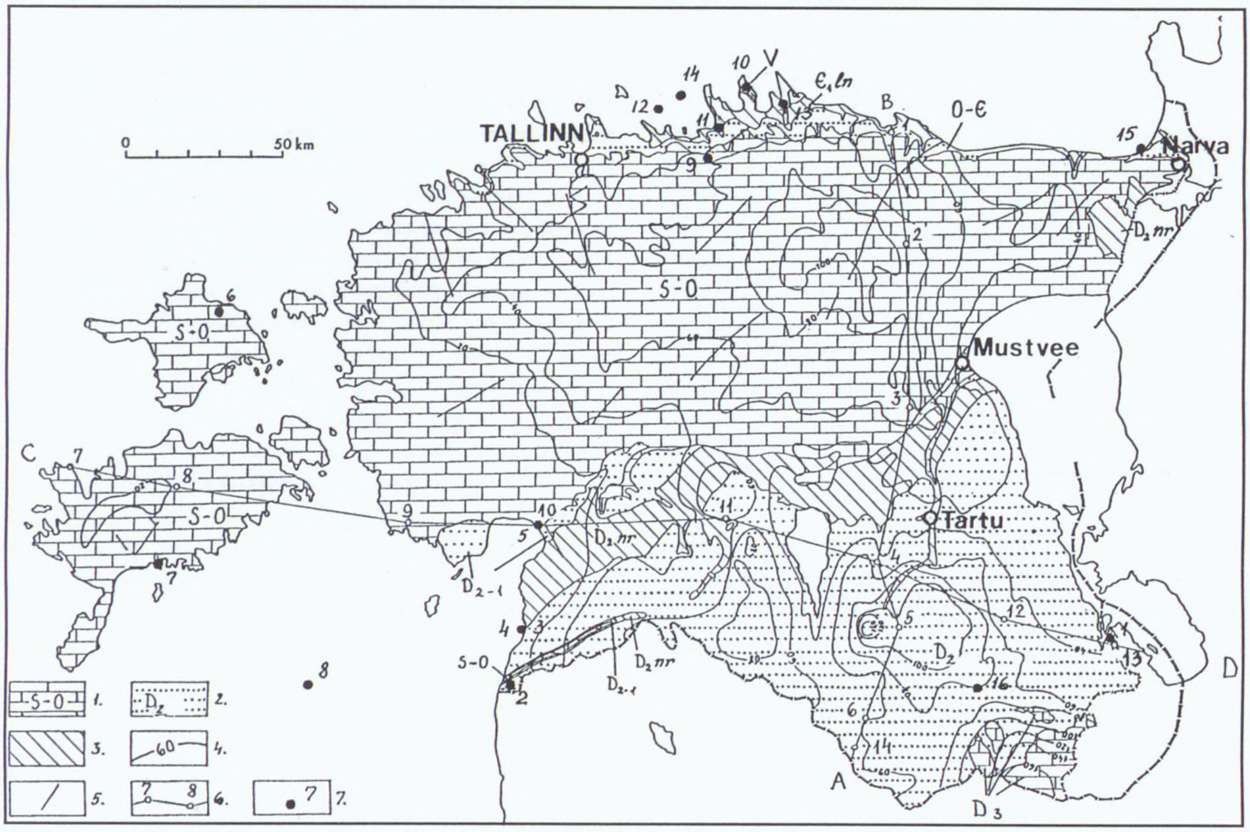


Fig. 94. Hydrogeological sketch-map of Estonian bedrock, distribution of aquifers and aquitards: 1 - carbonaceous (limestone, dolomite, marl), and 2 - terrigenous (sandstone, siltstone) water-bearing rocks; 3 - sporadically water-bearing and impermeable rocks; 4 - contours of the groundwater table and potentiometric surface a.s.l., m; 5 - tectonical fault; 6 - section line; 7 - mineral water deposit (see also Table 20): 1 - Värska, 2 - Ikla, 3 - Arumetsa, 4 - Häädemeeste, 5 - Pärnu, 6 - Kärdla, 7 - Kuressaare, 8 - Ruhnu Island, 9 - Hirvli, 10 - Pärispea, 11 - Pudisoo, 12 - Rammu Island, 13 - Käsmu, 14 - Põhja-Uhtju Island, 15 - Meriküla, 16 - Võru. For indices see Table 17. *Compiled by R. Perens.*

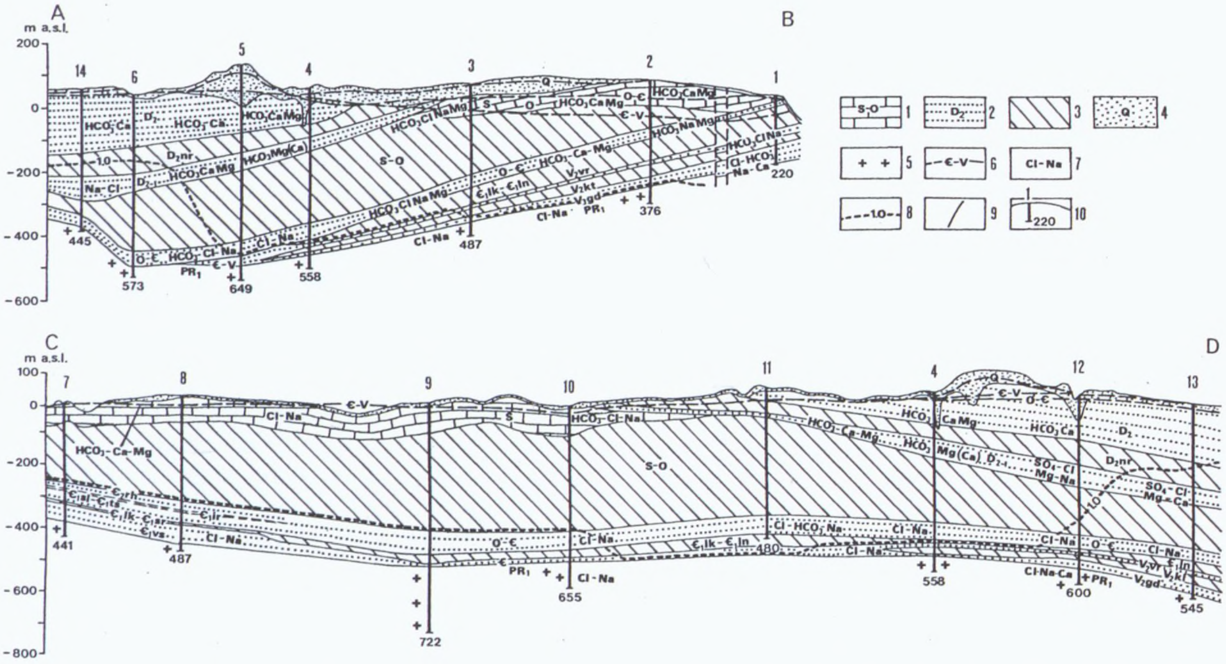


Fig. 95. Hydrogeological cross-sections of Estonian bedrock: 1 - carbonaceous (limestone, dolomite, marl), and 2 - terrigenous (sandstone, siltstone) water-bearing rocks; 3 - sporadically water-bearing and impermeable rocks; 4 - Quaternary deposits; 5 - crystalline basement; 6 - contours of the groundwater table and potentiometric surface; 7 - hydrochemical type of water; 8 - distribution limit of fresh groundwater; 9 - tectonical fault; 10 - boring, upper figure shows the number on the map, lower figure - depth, m. For indices see Table 17. *Compiled by R. Perens.*

weakly confined due to the occurrence of clayey interlayers in sands. The discharge of 2...4-m-deep wells ranges from 10 to 60 m³/d and this water is used in many households.

The glaciolacustrine deposits (lgQ_{III}), with a total thickness of 5...10 m, cover a large area and are represented by fine-grained sand, loamy sand and varved clay. Sands and light loamy sands are weakly or sufficiently water yielding with their conductivity varying from 0.1 to 5 m/d. Many wells with the discharge ranging from 0.5 to 20 m³/d have been sunk into these deposits.

Varved clay (lgQ_{III}) with a thickness of up to 22 m and transversal conductivity less than 10^{-4} m/d, forms medium and strong local aquitards all over Estonia. The largest aquitard (30 km²) occurs in the catchment of the Kasari River in western Estonia. Varved clay effectively protects deeper aquifers from pollution.

The glaciofluvial deposits (fQ_{III-II}) form frontal aprons, eskers and deltas and occur in some buried valleys. They consist mostly of sand and gravel, the conductivity of which is generally 5...10 m/d, in some places even up to 100 m/d. Due to this, the wells tapping the glaciofluvial deposits are generally high yielding. Glaciofluvial sediments in buried valleys are usually confined by clayey glaciolacustrine deposits and till. Public water intakes with a pumping rate of up to 10,000 m³/d tap glaciofluvial aquifers in the buried valleys at Vasavere near Jõhvi, at Raadi-Maarjamõisa in Tartu, and at Männiku-Pelguranna in Tallinn.

The glaciogeneous sediments (gQ_{III}) cover almost 2/3 of Estonian territory. Weakly or sufficiently water yielding are the loamy-sandy varieties of till and sporadically spread lenses of sand and gravel in till with a thickness of a couple of metres. The conductivity of loamy-sandy till ranges from 0.01 to 1.0 m/d. The majority of up-to-10-m-deep dug wells all over Estonia get water from till. The discharge of these wells is predominantly 0.2...2 m³/d. Usually the water table is at a depth of 1.5...3 m from the surface, quite often it is at a depth of 8...12 m, in the Otepää and Haanja heights occasionally even 20 m below the ground. In late summer, shallow (2 - 5 m) wells in loamy-sandy till often run dry. The loamy-sandy till with the conductivity of $10^{-3}...10^{-4}$ m/d is considered a weak or medium aquitard.

Water in the bedrock

The Upper Devonian aquifer system (D_3) consists of karstified and fissured dolomites and dolomitized limestones of the Dubniki and Pļaviņas stages. The total thickness of this aquifer system is 17...25 m and it covers some 500 km² in southeastern Estonia (Fig. 94). The siltstone of the Snetnaja Gora Member with interlayers of clay in the lower portion of the Pļaviņas Stage forms an aquitard with medium isolation ability. The aquifer system is overlain by the Quaternary cover with a thickness of 40...80 m. Groundwater is mostly confined and its potentiometric surface lies at a depth of 3...8 m from the ground. Big sink-holes through which melt- and rainwater percolates fast into the karstified bedrock strata occur at Rõuge, Meremäe, Meeksi and some other places. Conductivity of karstified carbonate rocks varies between 1...50 m/d. According to this, the specific capacity of wells ranges from 0.1 to 6.0 l/(s×m), predominantly it is 1 l/(s×m). Due to its limited occurrence, the Upper Devonian aquifer system is used for the public water supply in a few places only.

The Middle Devonian aquifer system (D_2) is extending in southern Estonia (Fig. 94) between the Gulf of Liivi (Riga) and Lake Peipsi. It consists of sand- and siltstones with interlayers and lenses of clay of the Amata, Gauja, Burtnieki and Aruküla stages. Clayey material prevails in the Amata Stage, forming with the Snetnaja Gora Member a weak or medium aquitard between the Upper and Middle Devonian. One third of the volume of this aquifer system includes clayey rocks which serve as weak or medium aquitards and, for this reason, probably form several confined aquifers of local distribution (Верте 1965). The occurrence of the latter has not been sufficiently proved yet.

The northern boundary of the distribution of the Middle Devonian aquifer system lies approximately on the Häädemeeste - Mustvee line. To the south from this line, the thickness of the aquifer system increases up to 250 m in the Haanja Heights. The aquifer system outcrops only occasionally in deeper river valleys, elsewhere it is covered with Quaternary deposits, ranging 5...80 m in thickness. In uplands the potentiometric surface lies at a depth of 10...15 m from the surface, while in lower areas a lot of flowing wells are encountered (Tõrva, Valga, Antsla, Võru, *etc.*). The absolute height of the potentiometric surface ranges from 80 to 130 m in the Otepää and Haanja heights, in the Sakala Upland it is between 50...80 m.

The lateral conductivity of aquifer system is rather equable: predominantly 1...3 m/d. Transmissivity reaches 200...500 m²/d in the Sakala Upland, Otepää and Haanja heights, elsewhere it is usually less than 100 m²/d. The storage coefficient amounts to $5\times10^{-5}...10^{-3}$. The discharge of wells changes between 200...700 m³/d per 3...7 m of drawdown. The specific capacity of wells is predominantly 0.4...1 l/(s×m). The Middle Devonian aquifer system is used for the public water supply mainly in the areas south of the Häädemeeste - Põlva line, but also in the towns of Tartu, Viljandi, Elva and Kallaste.

The Narva regional aquitard (D_2nr) consists of layers of siltstone, dolomite, marl and clay of the Narva Stage with a total thickness of up to 90 m. In southern Estonia, these layers form the uppermost effective bedrock aquitard, the transversal conductivity of which is $10^{-5}...10^{-4}$ m/d, in places 10^{-6} m/d or even less. The clayey layers of the Narva Stage serve as a regional aquitard for the whole Baltic Artesian Basin (Иодказис 1989). The rocks in the upper portion of the stage supply water for the area between Tartu and Mustvee and for Ruhnu Island. The specific capacity of wells is 0.06...0.2 l/(s×m). The Narva aquitard separates the Middle Devonian aquifer system from the underlying water-bearing strata.

The Middle-Lower-Devonian aquifer system (D_{2-1}). The Narva aquitard is underlain by the water-bearing layers of the Pärnu Stage (Middle Devonian) and Rēzekne and Tilžė stages (Lower Devonian) which consist of fine-grained weakly cemented sand- and siltstones with interlayers of clayey and dolomitized sandstone. Together with the underlying Silurian strata the layers are used for the public water supply in Pärnu, Viljandi and Tartu. The association of water-bearing strata has been named the Middle-Devonian-Silurian aquifer system and the united account has been kept of its water extraction and water resources (Savitski *et al.*1996). However, in view of the collector characteristics of the rocks, it would be more correct to treat the complexes of terrigeneous and carbonate rocks separately.

In southern Estonia, the Middle and Lower Devonian aquifer system with a thickness of up to 100 m lies at a depth of more than 200 m below sea level (Fig. 95). The water is predominantly confined. In lowlands, where the potentiometric surface extends above the ground, flowing wells occur. In the uplands, the potentiometric surface is at a depth of 10...20 m below the ground.

Due to its good collector characteristics, the water yielding capacity of sandstone is relatively high. The discharge of wells is predominantly between 260...700 m^3/d by drawdown of 4...10 m. The specific capacity of wells ranges from 0.25 to 1.0 l/(s×m). Conductivity of sandstones is mostly 2...6 m/d, rarely 8...10 m/d. Transmissivity of the aquifer system is 50...500 m^2/d, the storage coefficient ranges from 0.001 to 0.15.

The Silurian-Ordovician aquifer system (S-O) is an important source of water supply in the regions north of the Pärnu - Põlva line and on the islands of the West-Estonian Archipelago (Fig. 96). It consists of diverse limestones and dolomites with clayey interlayers. The upper portion of the rocky complex with a thickness of 30 m is extremely cavernous, with numerous cracks and fissures (Heinsalu 1977). Karst cavities form some half-a-metre-high canals trending in the direction of bedrock fissures. Caverns are especially abundant in dolomites and dolomitized limestones. Close to the ground, bigger karst cavities, a couple of metres high and some twenty or thirty metres long, occur in some places. Karst phenomena and fissures are most abundant in the carbonate rocks forming the upper part of the bedrock (Photo 32) - the weathering zone, usually 1...3 m, rarely 5...10 m in thickness. The deeper the lying depth of carbonate rock, the less fissures and cavities; such rocks generally turn into an aquitard. Southward from the Ikla - Elva - Mehikoorma line the Silurian-Ordovician rocks practically yield no water.

Photo 32. The carbonate bedrock is dissected by joints promoting karst processes and pollution of the upper aquifer systems. Lasnamägi in Tallinn. *Photo by A. Raukas.*

Besides traditional aquifer pumping tests, the impeller method (flowmeter logging) has been widely used in studying the filtration characteristics of Estonia's bedrock (Перенс 1984, Перенс и Палтанавичюс 1989, Perens *et al.* 1994). The results indicate (Fig. 97) that the Silurian-Ordovician carbonate rocks have fragmentary water conducting zones with parallel lamination and an abundance of fissures. In these 1...2-m-thick zones groundwater flows in a lateral direction (wells included). Water conductivity zones are separated from each other by 5...10-m-thick layers in which groundwater flows predominantly in vertical fissures. Only about 13% of the whole length of the rock complex is covered by lateral water zones (Перенс 1984). As an average, there are about 5 water conducting zones per 100 m of vertical cross-section. In a lateral direction the water conducting zones are fragmentary and their stratigraphical level may be more or less the same

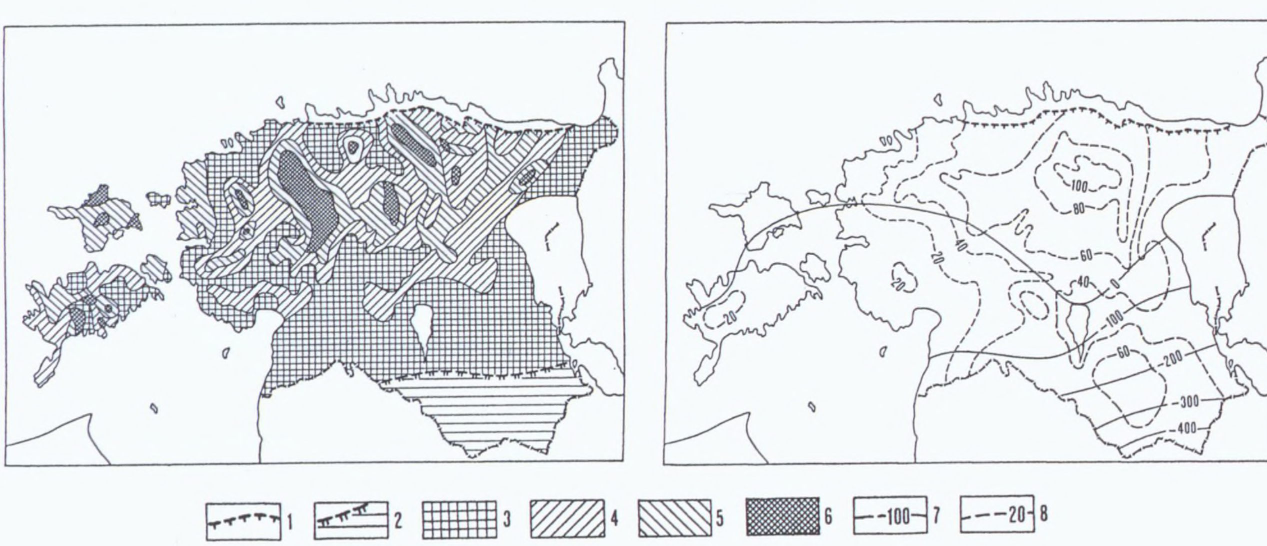
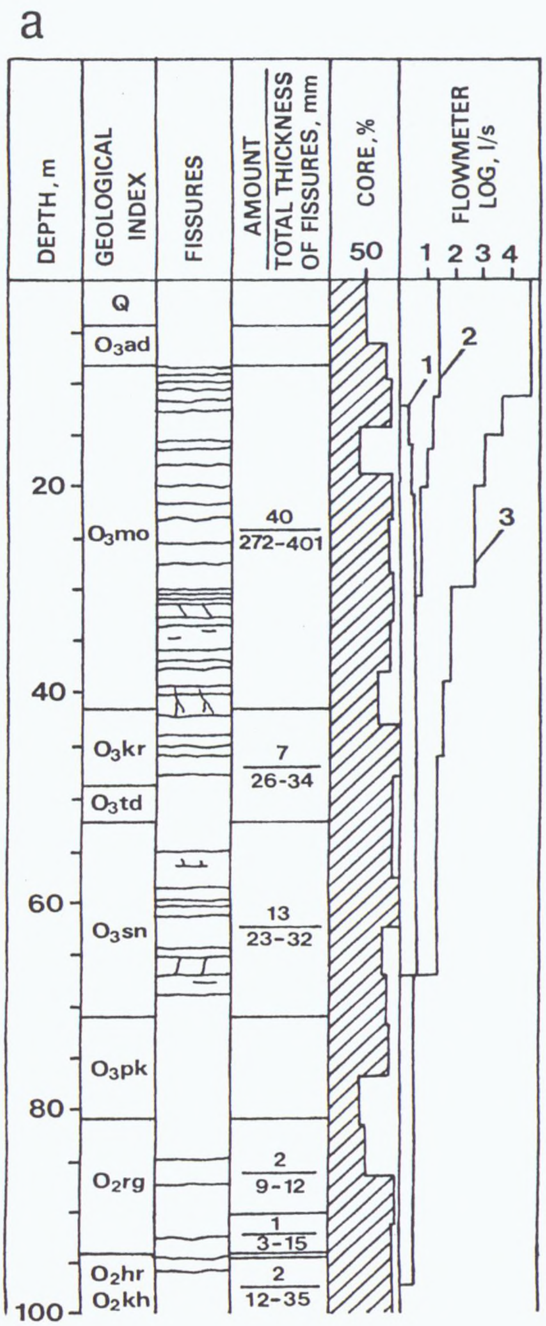


Fig. 96. Silurian-Ordovician aquifer system: 1 - boundary of the aquifer system; 2 - boundary of water-bearing rocks; 3 - 6 - transmissivity, m^2/d: 3 - up to 100; 4 -100...200; 5 - 200...500; 6 - more than 500; 7 - isoline of the height of aquifer roof a.s.l.; 8 - contour of the groundwater table and potentiometric surface, a.s.l., m. *Compiled by R. Perens.*



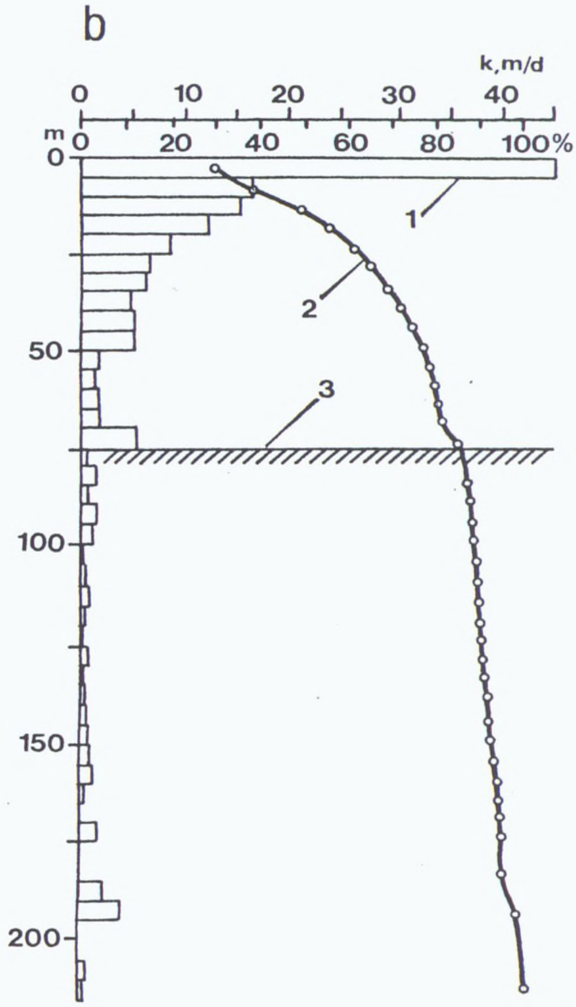


Fig. 97. Results of flowmeter logging: a) Geological section of Paasvere boring (southern portion of the Pandivere Upland), fissurization by television device and groundwater discharge into boring: 1 - flowmeter log under natural conditions; 2 - 3 - the same during pumping. Q - Quaternary cover; O_3 - Upper Ordovician; O_2 - Middle Ordovician; formations: ad - Adila, mo - Moe, kr - Kõrgessaare, td - Tudulinna, sn - Saunja, pk - Paekna, rg - Rägavere, hr - Hirmuse, kh - Kahula. b) Conductivity of carbonate rocks and the impact of bedding depth on total transmissivity: 1 - mean conductivity of weighed depth intervals (k), m/d; 2 - cumulative graph of transmissivity, %; 3 - the boundary of the most water-abundant portion of the section. *Compiled by R. Perens.*

only within a couple of kilometres. The water zones of quite different stratigraphical levels can be found in wells lying only a few hundred metres from each other.

According to the logging data gathered in more than 300 wells, about half of water in these wells is provided by the upper portion of cross-section with a thickness of 15 m and average transmissivity of 400 m²/d. Downwards the total transmissivity of carbonate rocks is evenly increasing and at depths of 50 and 75 m it is 630 and 700 m²/d, respectively. As new water conducting zones are very rare in deeper layers, the depth of 75 m can be considered the lowest border of sufficiently water yielding layers of the whole Silurian-Ordovician aquifer system. In western Estonia and on islands, the thickness of the main water yielding portion of the Silurian-Ordovician aquifer system is only 30...40 m. In most of Estonia, the total transmissivity of the Silurian-Ordovician aquifer system usually varies between 100...500 m²/d, depending essentially on the distribution of zones of tectonic disturbances. In those areas transmissivity increases and often exceeds 1000 m²/d.

The lateral conductivity in the carbonate bedrock is variable: 10...50m/d in the topmost 20 m, 5...8 m/d at a depth of 20...50 m, and only 1...2 m/d at a depth of 50...100 m. The lateral conductivity of deeper strata does not usually exceed 1 m/d, although occasionally strata with considerably higher conductivity can be found even as deep as 200 m from the surface (Fig. 97). The data of 235 pumping tests in the oil shale region of northeastern Estonia have shown that the lateral conductivity of Ordovician carbonate rocks near the ground is predominantly between 5...300 m/d, while at a depth of 80...100 m it is only 0.1 m/d (Рнет 1974, Riet 1976).

According to the water budget calculations, the transversal conductivity of the layers between the lateral water conducting zones is $10^{-5}...10^{-2}$ m/d (Валлнер 1980, Йыгар 1983). These interlayers serve as weak or medium aquitards confining the local aquifers of different range, the distribution of which is not yet completely clear. In northeastern Estonia (Верте 1965, Газизов 1971, Йыгар 1977, Savitski *et al.* 1996 a.o.), including the region of oil-shale mines with plentiful experimental data, the following aquifers and aquitards can be distinguished (from top downwards): Nabala-Rakvere aquifer (O_{3-2}nb-O_2rk), Oandu aquitard (O_2on), Keila-Jõhvi aquifer (O_2kl-O_2jh), Jõhvi-Idavere aquitard (O_2jh-O_2id), Idavere-Kukruse aquifer (O_2id-O_2kk), Uhaku aquitard (O_2uh) and Lasnamäe-Kunda aquifer (O_2ls-O_1kn). The occurrence of aquitards and aquifers in the Silurian-Ordovician aquifer system has also been proved by hydrogeological modelling. The satisfactory calibration of the more extensive filtration models is impossible without this distinction (Vallner 1996b).

The average lateral conductivity of the Silurian-Ordovician aquifer system is 8.1 m/d. Higher values of conductivity have been recorded in the limestones of the Raikküla Stage (24.2 m/d) and in the Adila Formation (17.0 m/d). The average conductivity of the Idavere and Uhaku stages is only 0.3 and 0.7 m/d, respectively (Perens 1989).

Water in the fissure systems and karst cavities of the carbonate bedrock flows relatively fast. In the outcrop area of aquifer system it recharges from Quaternary deposits and, for this reason, can easily become polluted in the areas with a thin Quaternary cover (Photo 26). The upper portion of aquifer system enfolds an aeration zone which in North-Estonian uplands is up to 20 m and in alvars only a couple of metres thick. Elsewhere, the Silurian-Ordovician aquifer system is more or less confined by the Quaternary cover and the uppermost aquitards of the carbonate bedrock. On the foot of uplands and in river valleys the potentiometric surface can often be 0.5...2 m above the ground which is the reason for many springs and flowing wells (Heinsalu & Vallner 1995). The North-Estonian watershed serves as the main area of head generating. In the Pandivere Upland, the groundwater level is at a height of up to 110...120 m above sea level.

The specific capacity of wells tapping the upper portion of the aquifer system ranges from 1 to 3 l/(s×m), being 3...5 l/(s×m) on average. The specific capacity of wells deriving water from deeper strata does not usually exceed 1 l/(s×m). The average yield of wells by drawdown of 5...10 m is usually 400...900 m^3/d. The average specific yield of unconfined aquifer system is 0.02...0.06, depending on the degree of fissuration and karstification. The storage coefficient changes between $10^{-6}...10^{-3}$.

The Silurian-Ordovician regional aquitard (S-O) consists of limestones, marls, siltstones, clays and argillites of the Toila, Leetse, Varangu and Türisalu formations (Lower Ordovician), extending at a length of 30 km southward from the North-Estonian Klint. Farther in the south, the aquitard includes all Silurian and Ordovician rocks. Its thickness increases from a couple of metres in the vicinity of the klint up to 200...350 m near the southern border of Estonia. Conductivity is very variable: the lateral conductivity changes between 0.001...1 m/d, extending sometimes up to 2...5 m/d (Fig. 97), the transversal conductivity ranges from 10^{-7} to 10^{-5} m/d (Vallner 1996a, Валлнер 1980).

The Ordovician-Cambrian aquifer system (O-€) underlies the Silurian-Ordovician regional aquitard extending under most of Estonia, except the North-Estonian coastal region, Mõniste-Lokno uplift area and the islands of the West-Estonian Archipelago (Fig. 98). The aquifer system includes the Kallavere-Tiskre aquifer (O_1*kl*-$€_1$*ts*) in mainland Estonia which consists of fine-grained sandstone and siltstone of the Lower Ordovician Kallavere Formation and the Lower Cambrian Tiskre Formation. The thickness of the aquifer system is 20...60 m, it increases from north to south. The depth from the ground increases from 10...20 m at the North-Estonian Klint up to 500 m on Estonia's southern border. The main recharge area is the Pandivere Upland where water from Ordovician strata leaks downward through the Silurian-Ordovician regional aquitard and disperses in radial directions as confined filtration flows. There the absolute altitude of potentiometric surface is up to 70 m under natural conditions. In Lower Estonia, Võrtsjärve, Alutaguse, Valga and Varnja-Värska lowlands and in the Väike-Emajõgi Valley, the potentiometric surface of the Ordovician-Cambrian aquifer system extends above the ground. The aquifer system is an important source of public water supply in northern Estonia and it is also intensively used in the towns of Pärnu, Viljandi and Tartu. This has caused several depressions of potentiometric surface (Fig. 98).

The lateral conductivity of the Ordovician-Cambrian aquifer system mostly ranges from 1 to 3 m/d. Transmissivity tends to decrease southwards. However, due to the thickness of water-bearing strata it is 80...130 m^2/d in central and southeastern and only 25...50 m^2/d in northern Estonia. The yields of wells are predominantly 430...600 m^3/d per 10...15 m of drawdown. The specific capacity of wells changes between 0.2...0.4 l/(s×m). The storage coefficient is $2.5\times10^{-5}...6\times10^{-3}$; the specific yield of the aquifer drained is 0.12...0.14.

On the islands of the West-Estonian Archipelago, the Ruhnu ($€_1$*rh*) and Soela-Tiskre ($€_1$*sl-ts*) aquifers, isolated by the Irben ($€_1$*ir*)aquitard, belong to the Ordovician-Cambrian aquifer system. They consist of the Lower Cambrian sand- and siltstone (Ruhnu and Soela formations).

The Lükati-Lontova regional aquitard ($€_1$*lk*-$€_1$*ln*), spread in most of mainland Estonia, is represented by siltstones and clays of the Lower Cambrian Lükati and Lontova formations. The total thickness of layers decrease from 90...100 m in northern Estonia to pinching out on the Mõisaküla - Vastseliina line in southern Estonia and in western Estonia (Fig. 99). This aquitard has a strong isolation capacity; the transversal conductivity is predominantly between $10^{-7}...10^{-5}$ m/d (Vallner 1996, unpublished report, Валлнер 1980). In the West-Estonian Archipelago, the Ordovician-Cambrian aquifer system is isolated from the underlying Cambrian-Vendian aquifer system with clays and clayey siltstones of the Lower Cambrian Lükati and Sõru formations.

The Cambrian-Vendian aquifer system (€-V). Cambrian-Vendian terrigeneous rocks occur all over Estonia, except the Mõniste-Lokno uplift area. The water yielding portion consists of sand- and siltstones with interlayers of clay.

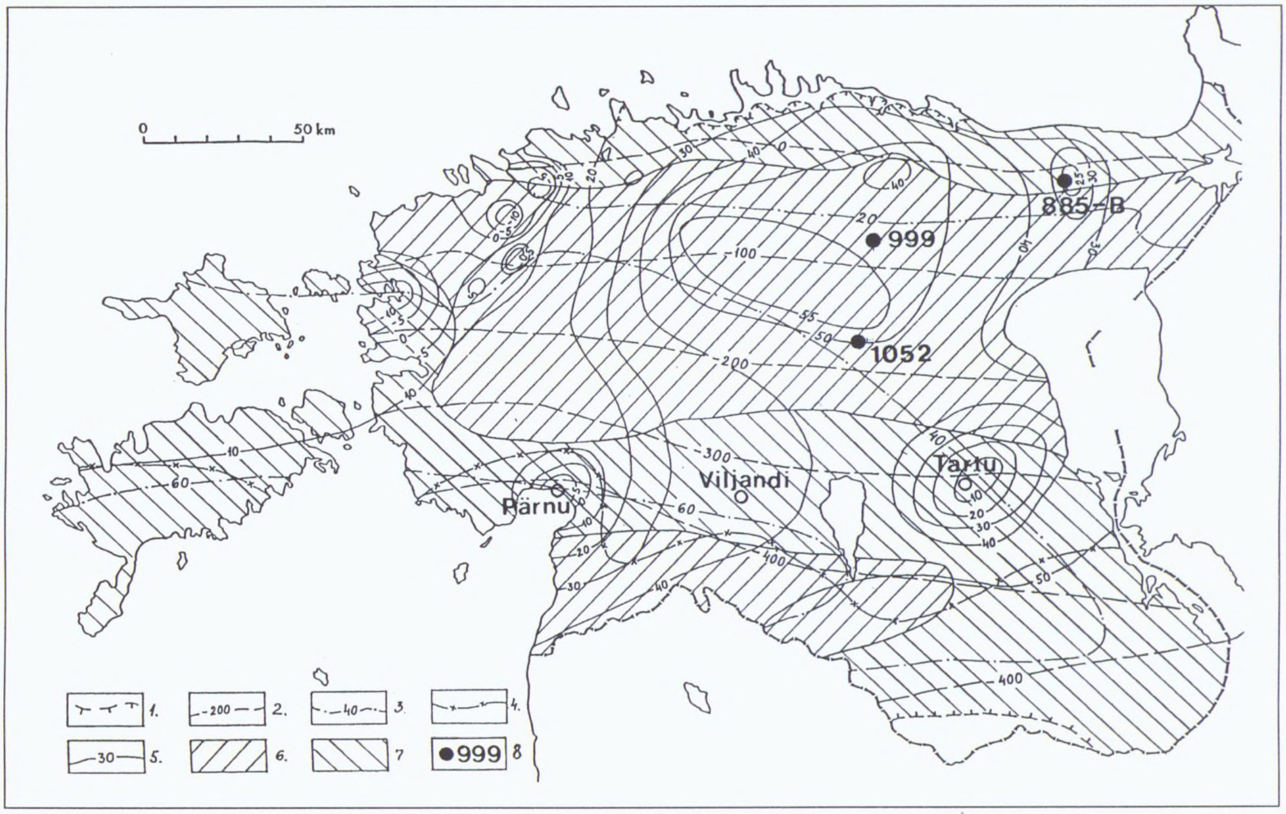


Fig. 98. Ordovician-Cambrian aquifer system: 1 - boundary of the aquifer system; 2 - isoline of the height of the roof of the aquifer system a.s.l., m; 3 - thickness of the aquifer system, m; 4 - boundary of brackish water; 5 - isoline of the height of the potentiometric surface a.s.l. in 1995, m; 6-7 - transmissivity, m^2/d: 6 - >50; 7 - <50; 8 - observation wells: 885 - Jõhvi, 999 - Väike-Maarja, 1052 - Tooma. *Compiled by R. Perens.*

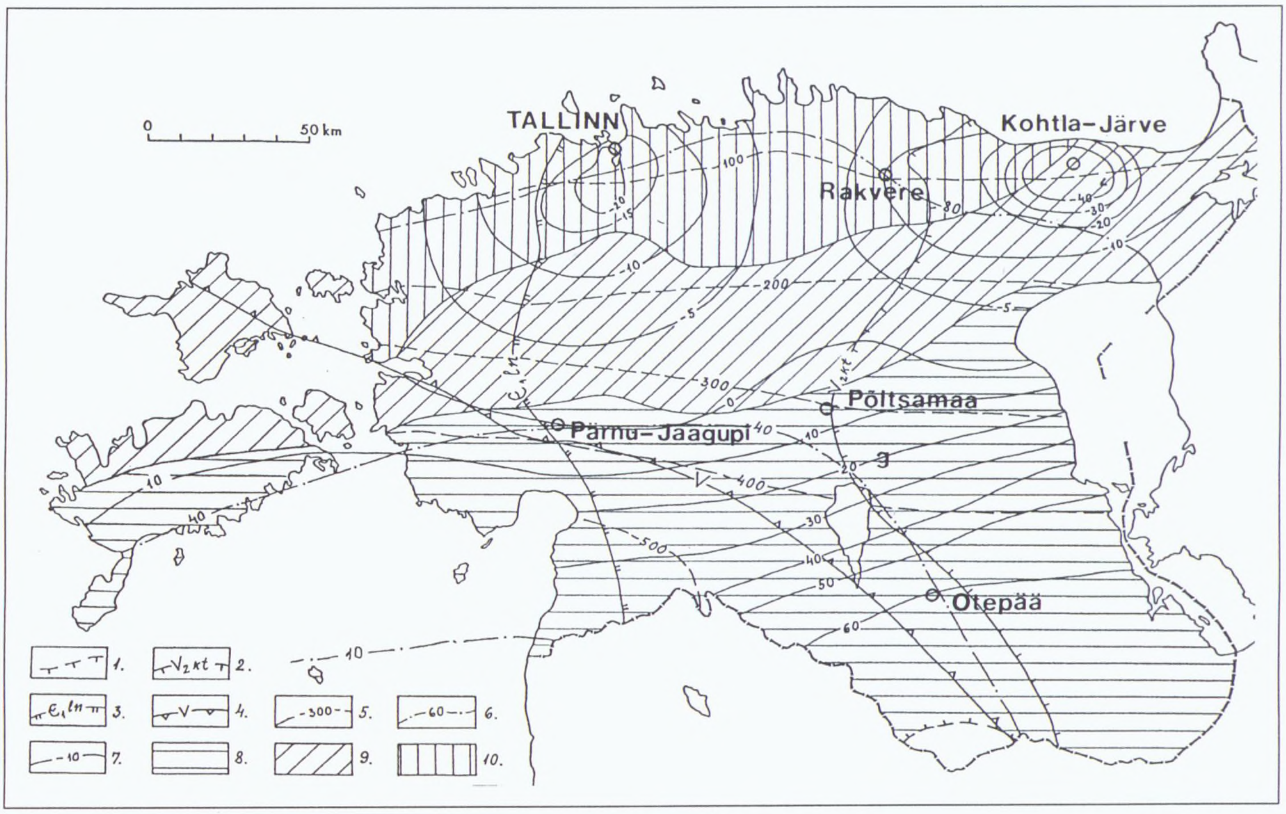


Fig. 99. Cambrian-Vendian aquifer system: 1 - boundary of the aquifer system; 2 - western boundary of the clays of the Kotlin Formation; 3 - western boundary of the clays of the Lontova Formation; 4 - boundary of Vendian rocks; 5 - isoline of the height of the roof of the aquifer system a.s.l., m; 6 - isoline of the thickness of the aquifer system, m; 7- isoline of the height of the potentiometric surface a.s.l. in 1995, m; 8-10 - transmissivity, m^2/d: 8 - up to 100, 9 - 100...300, 10 - more than 300. *Compiled by R. Perens.*

The difference between the cross-sections of western and eastern Estonia is obvious (Figs. 94, 95, 99). East of the Rakvere - Põltsamaa - Otepää line, the up-to-53m-thick clays of Kotlin Formation (V_2kt) divide the aquifer system into two aquifers. The upper, Voronka aquifer, consists of quartzose sand- and siltstone with a thickness of up to 45 m in northeastern Estonia. The lower, Gdov aquifer is formed of up-to-68-m-thick complex of mixed-grained sand- and siltstone.

In northern Estonia, the aquifer system is confined by 60...90-m-thick clays of the Lontova Formation. Westwards from the Tallinn - Pärnu-Jaagupi line, the Lontova Formation is gradually replaced by interbedding clay and sandstone of the Voosi Formation, which attain a thickness of 90 m in southwestern Estonia. On the West-Estonian islands, the Vendian deposits have also been pinched out and the water-bearing terrigeneous rocks consist only of Cambrian sand- and siltstones with interlayers of clay.

The Cambrian-Vendian aquifer system is the most important source of public water supply in northern Estonia. Intensive water extraction has led to the formation of two extensive depressions of potentiometric level (Fig. 99).

The Voronka aquifer (V_2vr) in eastern Estonia consists mainly of quartzose sand- and siltstones of the Voronka Formation, up to 40 m in thickness.The conductivity of rocks ranges from 0.6 to 12.5 m/d, being 2...6 m/d on average. Transmissivity decreases from 100...150 m^2/d in northern Estonia to 50 m^2/d and even less in the southern direction. In northern Estonia, the specific capacity of wells ranges from 0.2 to 5.8 (average 2) l/(s×m). In central and southern Estonia, it is 0.1...0.3 l/(s×m). Under natural conditions, the potentiometric level in the coast of the Gulf of Finland was 1.5...5.5 m above sea level.

The Gdov aquifer (V_2gd) consisting of mixed-grained sand- and siltstone with the thickness predominantly between 40...65 m lies directly on the Precambrian basement. The clay of the Kotlin Formation serves as an upper confining unit. In northern Estonia, the conductivity of water-bearing rocks is 0.5...9.2, average 5...6 m/d. Transmissivity in northeastern Estonia is 300...350 m^2/d; it decreases in a southerly and westerly direction to 100 m^2/d or less. Specific capacities of wells differ, the average value being 1.5...2.5 l/(s×m). Since most of wells tap both the Gdov and Voronka aquifers, their specific capacity is considerably higher. The potentiometric surface was 3...5 m above sea level under natural conditions in the coastal area of northern Estonia .

Westward from the line where the Kotlin clays are pinching out (Fig. 99), the Cambrian and Vendian water-bearing rocks form the steady Lontova-Gdov aquifer. On the West-Estonian islands, the Vendian sediments are absent and all the rocks deeper than the Lükati-Sõru aquitard ($Є_1lk\text{-}sr$) form the Voosi aquifer ($Є_1vs$). The productivity of wells on the islands of the West-Estonian Archipelago is 3...4 times lower than in northern Estonia.

The crystalline basement (PR_1) comprises groundwater in its upper weathered and fissured portion only. The specific capacity of wells does not exceed 0.1...0.2 l/(s×m). Natural potentiometric surface refracts the heads of the overlying Cambrian-Vendian aquifer system. Flowing wells occur in some places. Currently, the water stored in the crystalline basement is not used for water supply.

In extensive regional hydrogeological reviews the Estonian water-bearing formation has been treated as the northern slope of the Baltic Artesian Basin extending from the Gulf of Finland up to Minsk and Warsaw (Иодказис 1989). This viewpoint, although based on structural and geological aspects and a certain unity of formation of palaeohydrogeological conditions of deep groundwater, is still a theoretical construction. From the applicational point of view, the Estonian water-bearing formation should be considered as an independent artesian basin because the exchange of underground water with the neighbouring areas is less than 0.1% of the total annual groundwater recharge (Валлнер 1980).

GROUNDWATER FLOW

Main components of the budget and flow systems

In Estonia, groundwater is recharged from rain- and meltwater percolating through unsaturated soils. Mean annual precipitation is in the range of 500...750 mm or 1370...2060 $m^3/(d\times km^2)$, and is lower on the coast and some 10...20% higher on uplands.

The net infiltration (*I*) for Estonia (total groundwater recharge minus evaporation from the zone of saturation or capillary fringe) has been calculated preliminary from the budget equation comprising the main components of groundwater flow (Валлнер 1976, 1980):

$$I = R + Q + M - W \pm V \pm S$$

where *R* is the groundwater discharge (base flow) m^2/d to streams; *Q* is the pumpage from layers; *M* is the direct seepage of groundwater to the sea; *W* is the flux from streams into aquifers (induced recharge); *V* is the subsurface exchange of groundwater between Estonia and surrounding areas and *S* is the storage change.

The long-term groundwater discharge to streams (*R*) has been estimated on the basis of observations carried out during several decades at more than 100 hydrological gauging stations all over Estonia. Apart from the gauging stations, many irregular measurements of the low flow have been made approximately in 1000 stream cross-sections. The gained sporadic low flow data were modified to average base flow value by statistical methods using regular observations of gauging stations (Валлнер 1980). The pumpage data (*Q*) were obtained from state institutions checking the groundwater use. The subsurface fluxes to the sea (*M*) and groundwater exchange with adjacent areas (*V*) were calculated by Darcy's formula.

The total groundwater discharge to the channel network of Estonia (*R*) is approximately 7,700,000 m^3/d, but its intensity varies with regions (Fig.100). Average pumpage from wells and mines (*Q*) reaches 1,000,000 m^3/d causing the inverse fluxes (*W*) in an amount of *ca.* 500,000 m^3/d from surface bodies of water into aquifers. Direct groundwater seepage to the sea (*M*) averages 800,000 m^3/d. In the west and north, Estonia is bounded by the sea. In the east, its border runs along the Narva River and central portion of Lake Peipsi, in the south the border generally coincides with the divides of major streams. As a result, the exchange of groundwater with adjacent territories (*V*) is quite insignificant - not more than 10,000 m^3/d. For a long-term period the storage change *S* = 0. The total net infiltration (*I*) calculated by the budget equation averages 9,000,000 m^3/d.

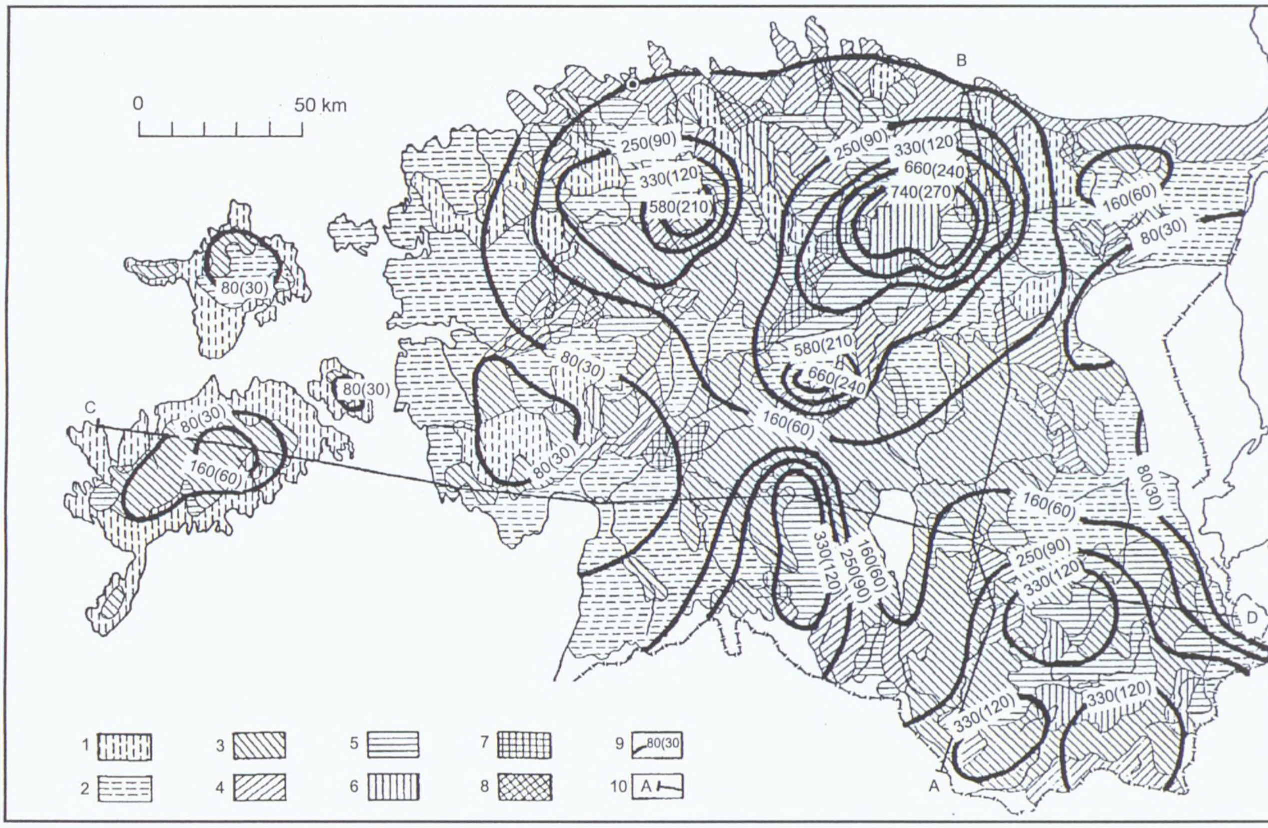


Fig. 100. Net infiltration and groundwater discharge to the channel network: 1- 8 - intensity of groundwater discharge, m³/(d×km²): 1 - 30...40, 2 - >40...90, 3 - >90...170, 4 - >170...260, 5 - >260...430, 6 - >430...600, 7 - >600...800, 8 - >800...1700; 9 - isoline of the net infiltration intensity, m³/(d×km²), in brackets mm per year; 10 - line of the cross-section of groundwater flow zones (on Fig. 101). *Compiled by L. Vallner.*

Groundwater discharge to the channel network (the base flow) and the instrumentally checked pumpage make up about 90% of the sum of the right-side members in the above water budget equation. Therefore, the value of net infiltration estimated by budget equation is probably more authentic than that based on indirect data, such as the air temperature and atmospheric humidity, evapotranspiration, *etc.* (Lerner *et al.* 1990). After completing the general budget of groundwater flow main components, the distribution of heads was simulated for the whole Estonian water-bearing formation (Валлнер и Тобиас 1984). Thereby the net infiltration was estimated by the trial and error method pursuing an optimum coincidence between the modelled and measured data.

The total amount of groundwater in the cracks and pores of Estonia's water-bearing strata is nearly 2000 km³. Both the distribution of the groundwater head and the direction of subsurface flows depend on spacial relations of the topography, surface bodies of water, and impermeable crystalline bedrock. Besides, human impact on the groundwater flow has been continuously increasing during the last five decades. Based on the total effect of the topography and geological structures, three main groundwater flow systems can be recognized in Estonia (Валлнер 1980, Tóth 1995)

The local flow system enfolds chiefly the unconfined or locally confined shallow groundwater moving from its recharge area toward the nearest ditches, creeks, rivers, and discharging directly to lakes or to the sea in coastal areas (Fig. 101). The length of the upper branches of the local flow system usually does not exceed a few kilometres, but lower branches which are not drained by small surface waterbodies and discharge to the middle courses of rivers can reach 10...15 km in length. The vertical thickness of the local flow system mostly ranges from 10 to 30 m.

The intermediate flow system takes its rise from Upper Estonia (Fig. 2) where the height of the terrain above sea level is more than 40 m. In the Harju Plateau, Otepää and Haanja heights, Pandivere, Jõhvi, Sakala and Karula uplands and in the Vooremaa area, the local maxima of the groundwater potential energy occur. In the Haanja and Otepää heights the groundwater table is 120...280 m above sea level, while in uplands and plateaus its height mostly ranges from 50 to 100 m. The groundwater head declines down from above in uplands; the difference of heads in some aquitard can be up to 50...60 m, and the corresponding head gradient up to 2.5...3. Consequently, a downward groundwater flow recharges the underlying aquifers from the overlying ones in highlands.

Groundwater moves from Upper Estonia towards Lower Estonia where the surface altitudes are 0...40 m above sea level. Branches of the intermediate flow system discharge to surface waterbodies between absolute heights of 30...40 m in the Võrtsjärv, Alutaguse, Valga and Varnja-Värska lowlands, and in the Väike Emajõgi Valley. The portions of the intermediate flow system formed in the Sakala Upland and in the western part of the Harju Plateau are drained by the lower course of the Kasari River between altitudes 0...30 m, while the lower branches of this flow system discharge to the offshore Baltic Sea. The intermediate flow system is completely confined. Its branches, approaching the discharge areas, bend up and recharge the local flow system from the underside. It is proved by decreasing of the groundwater head from below

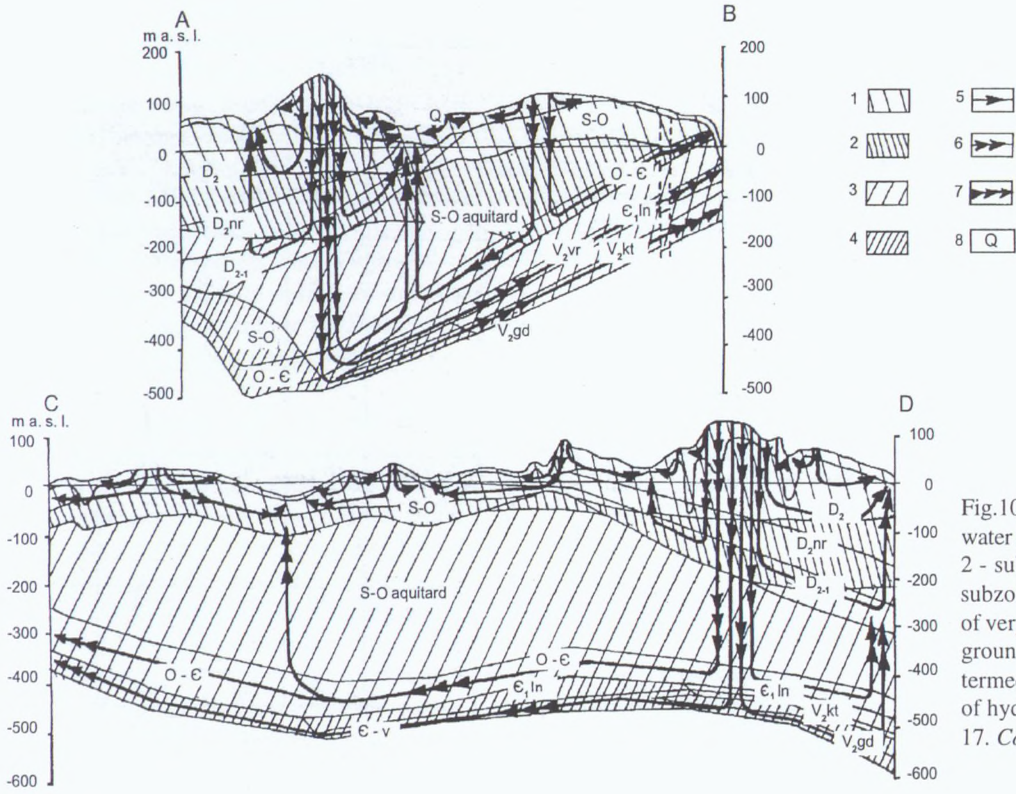


Fig.101. Vertical zoning of groundwater flow: 1 - subzone of fast flow; 2 - subzone of moderate flow; 3 - subzone of slow flow; 4 - subzone of very slow flow; 5-7 - systems of groundwater flow: 5 - local, 6 - intermediate, 7 - regional; 8 - index of hydrogeological unit, see Table 17. *Compiled by L. Vallner.*

upward observed in lake depressions and lower course valleys of most significant rivers where a lot of deep flowing wells occur. A portion of the intermediate flow system formed in the Harju and Viru plateaus, Pandivere and Jõhvi uplands discharges through springs or seeps out from layers on the North-Estonian Klint. The lateral length of the intermediate flow system is up to 100 km.

The regional flow system takes its rise in those parts of the Haanja and Otepää heights where the groundwater table is 180 to 280 m above sea level. There the head declines down from above attesting to the existence of a downward groundwater flow. This flow reaching the absolutely waterproof portion of the crystalline basement changes its direction and bends towards the discharge areas which are situated in the depressions of the Baltic Sea and the Gulf of Finland. The regional flow system underlies the intermediate flow system enfolding the portion of the lithosphere where the confined groundwater moves directly into the sea under natural (predevelopment) conditions. In western Estonia, the regional flow system includes the strata of the Ordovician-Cambrian and Cambrian-Vendian aquifer systems lying higher than 250 m below sea level. The Cambrian-Vendian aquifer system belongs completely to the regional flow system west- and northward from lakes Võrtsjärv and Peipsi. The length of lower branches of the regional flow system can reach 300 km between the Haanja Heights and the central part of the Baltic Sea.

The diversity of rock permeability complicates and modifies the head distribution in the subsurface hydrosphere. Therefore, the groundwater movement is mainly parallel to bedding in aquifers, but transversal in aquitards. The latter are often disconnected at tectonic disturbances where relatively intense vertical groundwater flow may occur. The configuration of the potentiometric surface and pumpage data show that all aquitards are more or less permeable in Estonia. Evidence is derived from declining of the head of the Ordovician-Cambrian aquifer system in the opposite directions - from the Otepää Heights and the Pandivere Upland towards the Emajõgi Valley. There the groundwater flows rise up and discharge to the river regardless of the Narva and Silurian-Ordovician confining aquitards with a total thickness of up to 300 m. The transversal permeability of regional aquitards is an eminent, but not completely clarified evidence until now (Tissot & Welte 1978, Brace 1980, Tóth 1995). Apart from porosity, it may well be caused by microfissurization of rocks.

While the potential direction of the groundwater movement is marked by the flow systems described above, the actual quantity and velocity of subsurface flows depend on permeability of layers. Various co-influences of head distribution and permeability are expressed by the vertical zoning of the groundwater flow. In general, the velocity of the groundwater movement decreases with the increase of the flow depth.

Vertical zoning

The zone of active water exchange enfolds the upper portion of the water-bearing formation (Table 18, Fig. 101). All aquifer systems overlying the Lükati-Lontova aquitard north of the Pärnu - Tartu line belong to this zone. Southward the zone is limited from underneath by the Ordovician aquitards or the altitude of 250 m below sea level. In the zone of active water exchange, an upper subzone of fast groundwater flow may be distinguished which includes the Quaternary cover and Devonian aquifers less than 50 m below sea level and the Silurian-Ordovician aquifer system, as a whole. Farther down, a subzone of moderate groundwater flow occurs comprising the portion of the Middle Devonian aquifer at a depth of 50-250 m below sea level and the Ordovician-Cambrian aquifer system down to the lower boundary of the zone of active water exchange.

The subzone of fast flow is immediately influenced by climate. In this subzone the infiltration is ac-

Table 18. Budget of groundwater flows, millions of m³/d

Groundwater flow unit		Inflow				Outflow						
		From above	Lateral	From below	From surface water-bodies	Directly to channel network and lakes	Through springs	Pumpage	Into sea	Up	Lateral	Down
Zone of active water exchange	Subzone of fast flow	9.0	<0.1	0.2	0.5	7.2	0.5	0.9	0.7	0	<0.1	0.4
	Subzone of moderate flow	0.4	<0.1	<0.1	<0.1	0	<0.1	<0.1	0.1	0.2	<0.1	0.1
Zone of passive water exchange	Subzone of slow flow	0.1	<0.1	<0.1	<0.1	0	0	0.1	<0.1	<0.1	<0.1	<0.1
	Subzone of very slow flow	<0.1	<0.1	0	0	0	0	0	<0.1	<0.1	<0.1	0
Total*		9.0	<0.1	0.2	0.5	7.2	0.5	1.0	0.8	0.2	<0.1	0

* Flow rates less than 0.1 million m³/d have not been accounted at summation

cumulating and evaporation from the aeration zone takes place. From the subzone of fast flow all underlying strata are recharged and also the whole groundwater flow discharging to stream channels permeates through this subzone. As mentioned above, in Estonia the total net infiltration reaches 9,000,000 m³/d, which makes an average of 200 m³/(d×km²) or 73 mm per year, but its actual value significantly varies with areas (Fig. 100). Infiltration is most intensive (350 to 770 m³/(d×km²) in the uplands of northern Estonia, where the carbonate bedrock abounds in karst phenomena. In an area of about 1000 km² in the central part of the Pandivere Upland, where the channel network is entirely lacking, most of rain- and meltwater, which has not been removed by evapotranspiration, percolates into karst interstices. Owing to the high permeability of the karsted carbonate bedrock, a significant portion of this newly formed groundwater quickly discharges to adjacent streams. That's why the groundwater table is relatively deep in the upland, being often 7...20 m below the ground surface. With such a depth, the discharge by evaporation from the aeration zone is negligible and the net infiltration reaches its maximum in Estonia - up to 900 m³/(d×km²).

In the uplands of southern Estonia, the net infiltration is twice as high as its average value. Highlands form only 16% of Estonia's total area but, nevertheless, about 40% of the total groundwater recharge takes place just on the uplands and their slopes. On the plateaus of northern and southern Estonia, the net infiltration is 170...260 m³/(d×km²) but in lowlands it is less than 170 m³/(d×km²).

Mires impede the recharge of deeper groundwater by infiltration. Fens are recharged from rising groundwater flows and bogs are usually located above an effective local aquitard which restricts the downward filtration. Similar conditions occur in the domains of glaciolacustrine varved clays covering an area of some thousand square kilometres.

The intensity of groundwater discharge to the channel network varies in a wide range, but its mean value is 170 m³/(d×km²). On the Harju and Viru plateaus, the groundwater discharge into streams averages 130 m³/(d×km²). The groundwater discharge to the channel network is at its highest, reaching 400...900 m³/(d×km²) in some topographic catchments around the Pandivere Upland where a lot of springs occur. In this area, the groundwater discharge averages 300 m³/(d×km²).

Springs are essential sources of recharge for many streams in northern Estonia. Commonly, they are associated with tectonic faults in the carbonate bedrock and occur in groups. The central part of the Pandivere Upland where the permanent channel network is lacking, is surrounded by a belt of gravity springs at an height of 80 to 100 m above sea level. Unconfined groundwater discharges through these, often intermittent springs (Heinsalu & Vallner 1995). A significant portion of confined groundwater formed in divides is discharged through ascension springs which are situated farther away from the upland in places where tectonic faults cross the river valleys. There groundwater moves upward along tectonic disjunctions of aquitards. About half of rising groundwater discharges directly to the streams dispersively or through subaquatic springs.

The total discharge of some spring groups draining the karsted bedrock is up to 50,000 m³/d during wet periods, but mostly it ranges from 500 to 10,000 m³/d. Besides the Pandivere Upland, there are a lot of springs in the upper course of rivers flowing out from the Harju Plateau and also in the upper course of the Navesti River. All in all, some 1500 springs occur in North-Estonian uplands and plateaus, but predominantly their discharge is less than 50 m³/d. The abundance of springs points to both the high permeability and vulnerability of the carbonate bedrock.

In the uplands and plateaus of southern Estonia, the subzone of fast flow includes mostly till, but also Devonian sand- and siltstones. The intensity of the groundwater discharge into the channel network ranges from 130 to 350 m³/(d×km²). Of an abundance of springs occurring in this area, the vast majority have a discharge of up to 10 m^3/d, and only a few springs 1000 to 1500 m^3/d. In Lower Estonia (Fig. 2), the intensity of groundwater runoff usually ranges from 50 to 100 $m^3/(d \times km^2)$, *i.e.* less than the average. This is due to the low and weakly dissected topography and low permeability of upper aquifers. In the Fore-Klint Lowland, the groundwater runoff reaches 200 $m^3/(d \times km^2)$ which is higher than the average value. The klint abounds in springs, but their discharge usually does not outnumber 10...20 m^3/d during the dry period. The total discharge of springs averages 500,000 m^3/d in Estonia forming about 7% of the total groundwater discharge to the channel network.

The total groundwater discharge from the subzone of fast flow directly to the sea is 700,000 m^3/d. In Estonia, the total lengths of the channel network and coastline are 31,000 and 3800 km, respectively (Hang & Loopman 1995). Thus, the intensity of groundwater discharge per a unit length of the channel network is about 250 $m^3/(d \times km)$ and per a coastline unit - 180 $m^3/(d \times km)$. These parameters are in good accordance and prove the reliability of the performed water budget calculations.

In northeastern Estonia, the subzone of fast flow enfolds the oil shale mines where the groundwater head has withdrawn up to 60 m. As a result of dewatering of mines, drawdown occurs in an area of about 700 km^2 and the induced recharge of Ordovician aquifers from the channel network reaches 400,000 m^3/d.

Of the total amount of groundwater formed in Upper Estonia, 430,000 m^3/d flows laterally into Lower Estonia; of that ca. 170,000 m^3/d into the Fore-Klint Lowland. The groundwater discharge directly to Lake Peipsi is 150,000 m^3/d.

Actual velocity of groundwater movement is variable in the subzone of fast flow. During wet periods, in the outcrop of the carbonate bedrock a lot of intermittent gravity springs, recharged from flows of shallow groundwater, come into being. Experiments carried out by means of dyestuffs have shown that the velocity of such groundwater movement is up to 5000 m/d. In deeper layers the actual velocity of groundwater v is much smaller and it is calculated by the formula $v = IK/n$, where I is hydraulic gradient, K is conductivity, and n is porosity.

In the above lateral water-conducting zones of the carbonate bedrock, the actual groundwater velocity v predominantly ranges from 1 to 10 m/d under natural conditions. In transversal fissures connecting these zones it is mostly 0.001...1 m/d. Consequently, in the karstified carbonate bedrock of high permeability it may take a month for the polluted groundwater to reach from the ground surface to a depth of 30 m, where usually the cased portion of a well ends. During another month, it may cover some 300 m in a lateral direction. After a year, the pollution may be found at a distance of about one kilometre from the pollution source. Approximately such velocities of groundwater movement have been observed in cases of the oil pollution on the outcrop of carbonate bedrock. In extremely permeable aquifers, groundwater may flow from the surface to a depth of 30...60 m even in a few days. However, in the Silurian-Ordovician carbonate bedrock the actual lateral velocity of groundwater movement ranges from 0.1 to 3 m/d. It is lower in Lower Estonia and in deeper layers.

In South-Estonian highlands, the hydraulic gradient of downward groundwater flows is between 0.01...0.4, but the gradient of lateral flows ranges from 0.0001 to 0.01. The lateral velocity of groundwater is 0.02...0.2 m/d in sandstones; the transversal velocity ranges from 0.001 to 0.005 m/d. In loamy till, the velocity of groundwater movement usually does not exceed 0.001 m/d, but in glaciolacustrial or glacial sandy loam it is up to 0.1 m/d. The velocity of groundwater movement is 0.0005...0.001 m/d in peat, 0.001...0.15 m/d in sand and 10...15 m/d in gravel.

A single water exchange may take place in some short and highly permeable branches of local flow system during a couple of months. Sometimes such branches with a length of up to 0.5 km occur between the local divide and the nearest stream. If the permeability of flow system is moderate or low, then under the same conditions a couple of years are required for a single water exchange. In the case of very low permeability of local flow system, when conductivity is 0.01...0.1 m/d (loam, peat, fine sand), it may take 50...150 years.

The time needed for a single water exchange has been calculated assuming that the existing groundwater will be replaced gradually along the branches of flow system which are isolated from one another. Actually, a more or less intensive water exchange takes place between all adjacent branches of groundwater flow. The most intensive water exchange occurs across the groundwater table directly affected by infiltration and evaporation. If an unconfined aquifer consists of Quaternary deposits or sandstone, the average thickness of which is 10 m and porosity 0.2, then at a common infiltration rate of 200 $m^3/(d \times km^2)$, the single water exchange takes about 25 years. Under the same conditions, in the carbonate bedrock with the porosity of 0.02 it takes only 3...4 years.

The subzone of moderate flow is recharged from the overlying subzone of fast flow. The amount of this downward flux is 400,000 m^3/d, *i.e.* only about 4% of total net infiltration. Of that, 160,000 m^3/d leaks through the Silurian-Ordovician regional aquitard with average intensity of 9.5 $m^3/(d \times km^2)$ in northern Upper Estonia. The intensity of downward fluxes penetrating the tectonic faults and recharging the underlying Ordovician-Cambrian aquifer system reaches 25...50 $m^3/(d \times km^2)$ or even more in the Pandivere Upland. The portions of the Middle-Lower Devonian aquifer system and the underlying Silurian aquifers which both belonging to the subzone of moderate flow are recharged through the Narva regional aquitard in South-Estonian highlands. In that case, the total amount of downward fluxes is 240,000 m^3/d and average intensity 20 $m^3/(d \times km^2)$. The lateral groundwater flux from Upper to Lower Estonia along the subzone of moderate flow is 220,000 m^3/d. Of that, 200,000 m^3/d returns into the overlying subzone of fast flow by uprising filtration. The flow to shelf deposits and from there into the sea averages 100,000 m^3/d, almost an equal amount goes to the underlying subzone of slow flow. Pumpage from the subzone of moderate flow was about 74,000 m^3/d in 1995.

The length of lateral groundwater flow branches belonging to the intermediate flow system ranged from 50 to 250 km under natural conditions, the hydraulic gradient was mostly 0.0003...0.0004. The actual velocity of groundwater movement was probably 0.005 m/d in Devonian and Cambrian sandstones and 0.05 m/d in Silurian carbonate rocks. At such velocities, a complete water exchange could have taken place only in the Silurian layers during the last 10,000 years, *i.e.* since the time the ice sheet retreated from Estonia's territory (Raukas & Rõuk 1995). During the same period, in sand- and siltstones the groundwater could have moved forward only some thirty or forty kilometres.

Owing to the intensive pumpage, a piezometric depression has formed in the subzone of moderate flow. Local cones occur in Haapsalu, Paldiski, Vasalemma, Tallinn, Kohtla-Järve, Pärnu, and Tartu where the drawdown ranges from 20 to 94 m. At the time being, the groundwater moves in many different directions towards the intakes in the subzone of moderate flow. The hydraulic gradient is 0.005...0.01 or even higher (Fig. 98). The actual velocity of the groundwater movement predominantly ranges from 0.1 to 2 m/d.

The zone of passive water exchange under the subzone of moderate flow enfolds the Silurian-Ordovician regional aquitard and all underlying strata south of the Tartu latitude (Fig. 101). Farther in the north, it comprises the Lükati-Lontova regional aquitard, the Cambrian-Vendian aquifer system, the water-bearing portion of the crystalline basement, and in the West-Estonian Archipelago partially also the Ordovician-Cambrian aquifer system. In the zone of passive water exchange, the water moves at a considerably lower actual velocity than in the zone of active water exchange. Therefore, the above-mentioned relatively thick regional aquitards belong to the zone of passive water exchange. Under natural conditions, the zone of passive water exchange recharges from overlying strata in South-Estonian highlands only, in the Pandivere Upland such recharge is negligible.

The subzone of slow flow embraces the upper portion of the zone of passive water exchange to a depth of 350 m below sea level. Under natural conditions, the downward filtration from the overlying subzone of moderate flow into the subzone of slow flow averaged 30,000 m^3/d which was only 0.3% of the total net infiltration. The uppermost branches of the intermediate flow system with a length of up to 50 km, which belonged to the subzone of slow flow rised up in the region of Lake Peipsi recharging the subzone of moderate flow from below. The lower branches headed along the regional flow system towards discharge areas in the depressions of the Gulf of Finland or central Baltic Sea. The lateral hydraulic gradient of deep groundwater flows ranged from 0.0001 to 0.0003.

The calculated velocities of deep groundwater movement are between 0.0005...0.005 m/d under the above-described conditions, which means that during the last 10,000 years the deep groundwater could have move forward only by some twenty or thirty kilometres and a complete water exchange along flow branches was impossible. This viewpoint is proved by the isotopic composition of Cambrian-Vendian fresh water which shows values of $d^{18}O$ from -1.8 to 2.2% (Vaikmäe & Vallner 1991). Such water must have originated from the thawing of the ice sheet at the end of the Pleistocene and its preservation in the subzone of slow flow is only due to an extremely low velocity of groundwater movement.

At present, pumpage from the subzone of slow flow is about 110,000 m^3/d. Pumping wells are mostly situated in the coastal area of northern Estonia within 20 km from the sea. Pumpage is most intensive in Tallinn and Kohtla-Järve where the local centres of piezometric depression have formed and the maximum drawdowns reach 25 and 50 m, respectively (Figs. 99, 102). At the present time, the water moves to the centres of piezometric depressions in the subzone of slow flow. North of groundwater intakes the direction of flows is from the sea to the mainland, *i.e.* contrary to that in predevelopment conditions. Therefore, an encroachment of brackish sea water into coastal aquifers is taking place in the near-shore area of northern Estonia.

The subzone of very slow flow enfolds the lower portion of the zone of passive water exchange at a depth greater than 300 m below sea level. This subzone includes the lower strata of the Ordovician-Cambrian and Cambrian-Vendian aquifer systems (Fig. 101) south of Elva latitude which comprise water with TDS ranging from 1 to 22 g/l. Due to the lack of experimental data, the velocity of groundwater movement in this subzone is not yet clear. In the lowermost portion of the Estonian water-bearing formation which lies at a depth of 500...700 m below sea level in the vicinity of Ruhnu Island (Fig. 94), the water may be stagnant (Mazor 1995, Mazor *et al.* 1995). In any case, the water has not become fresh in the subzone of very slow flow during the postglacial period though the TDS of water might have decreased.

Water budget of aquifer systems

In Estonia, the total net infiltration enters first the Quaternary cover (Fig. 103). The downward flow from the Quaternary deposits into the underlying bedrock averages 5,300,000 m^3/d, while the direct discharge into the channel network is 3,000,000 m^3/d. Discharge through springs is 43,000 m^3/d and 320,000 m^3/d flows directly to the sea. The flux of bedrock water rising upward and recharging the Quaternary cover from below is 4,100,000 m^3/d.

The amount of water flowing from the Quaternary cover into the Upper and Middle Devonian aquifer systems reaches 960,000 m^3/d and the recharge from the underlying Middle Devonian aquifer system is 510,000 m^3/d. About ¾ of inflow discharges in the form of lateral flows to the channel network across the streambeds. The remaining amount discharges through springs, is extracted by pumping or leaks into deeper layers.

The Middle-Lower Devonian aquifer system is recharged from the Quaternary deposits on its outcrop and farther in the south from the overlying Middle Devonian aquifer system. The total inflow is 350,000 m^3/d of which some 80% rises upwards into the overlying Middle Devonian aquifer system or discharges into streams through the Quaternary deposits. Approximately 15% is pumped out or drained directly by the sea, while 5% goes into the underlying Silurian-Ordovician aquifer system.

Of the total downward flow formed in the Quaternary cover, the Silurian-Ordovician aquifer system receives 75% or 4,000,000 m^3/d. The induced recharge of the carbonate bedrock due to the dewatering of mines is 500,000 m^3/d. The upward recharge from the underlying Ordovician-Cambrian aquifer system reaches 70,000 m^3/d in Lower Estonia. Discharges to the channel network through Quaternary deposits and springs are 3,000,000 and 450,000 m^3/d, respectively. Pumpage from the Silurian-Ordovician aquifer system, including mine water, averages 760,000 m^3/d and discharge into the sea through the shelf is 300,000 m^3/d.

The Ordovician-Cambrian aquifer system is recharged

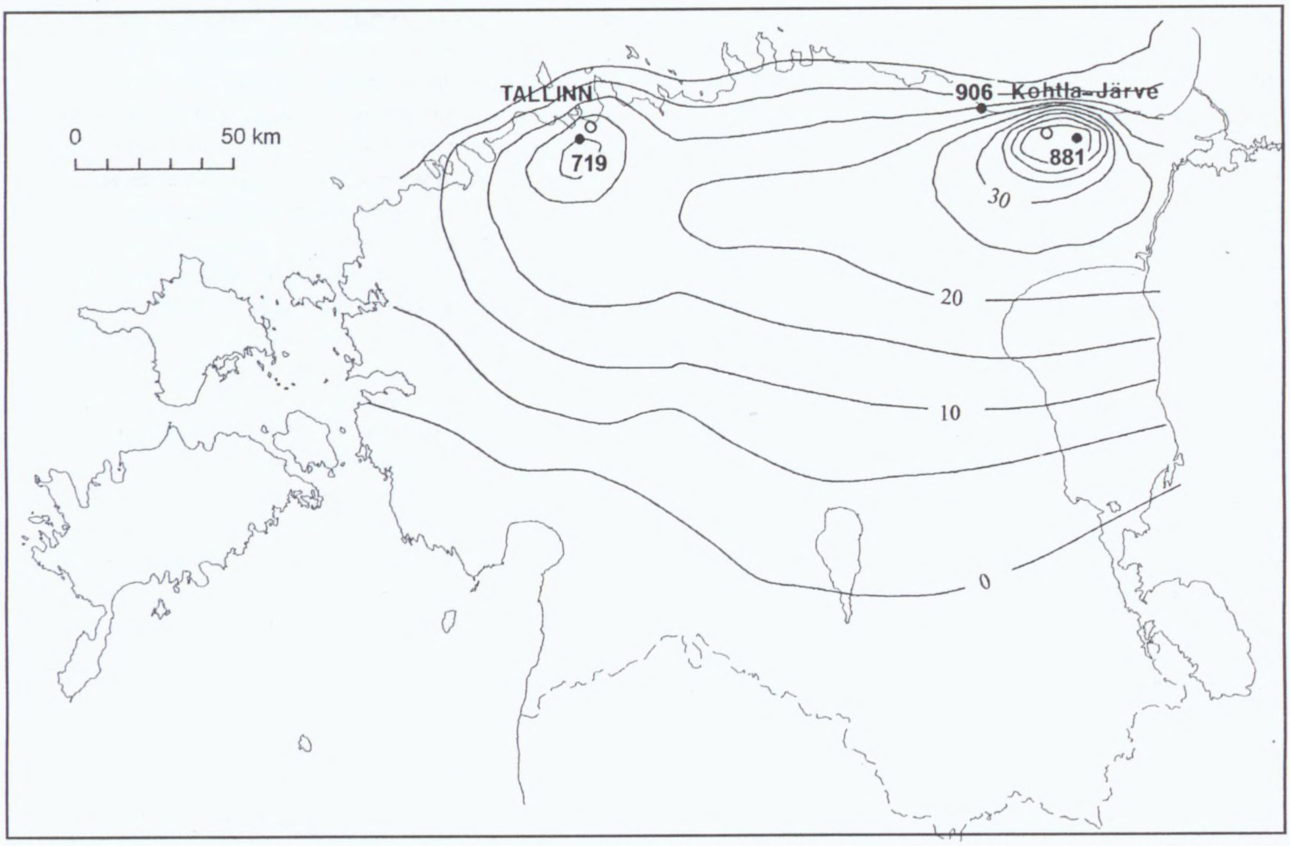


Fig. 102. Drawdown of the Cambrian-Ordovician aquifer system in 1994, m. Black circles mark observation wells: 719 - Tallinn, 906 - Purtse, 881 - Sompa. *Compiled by L. Savitski.*

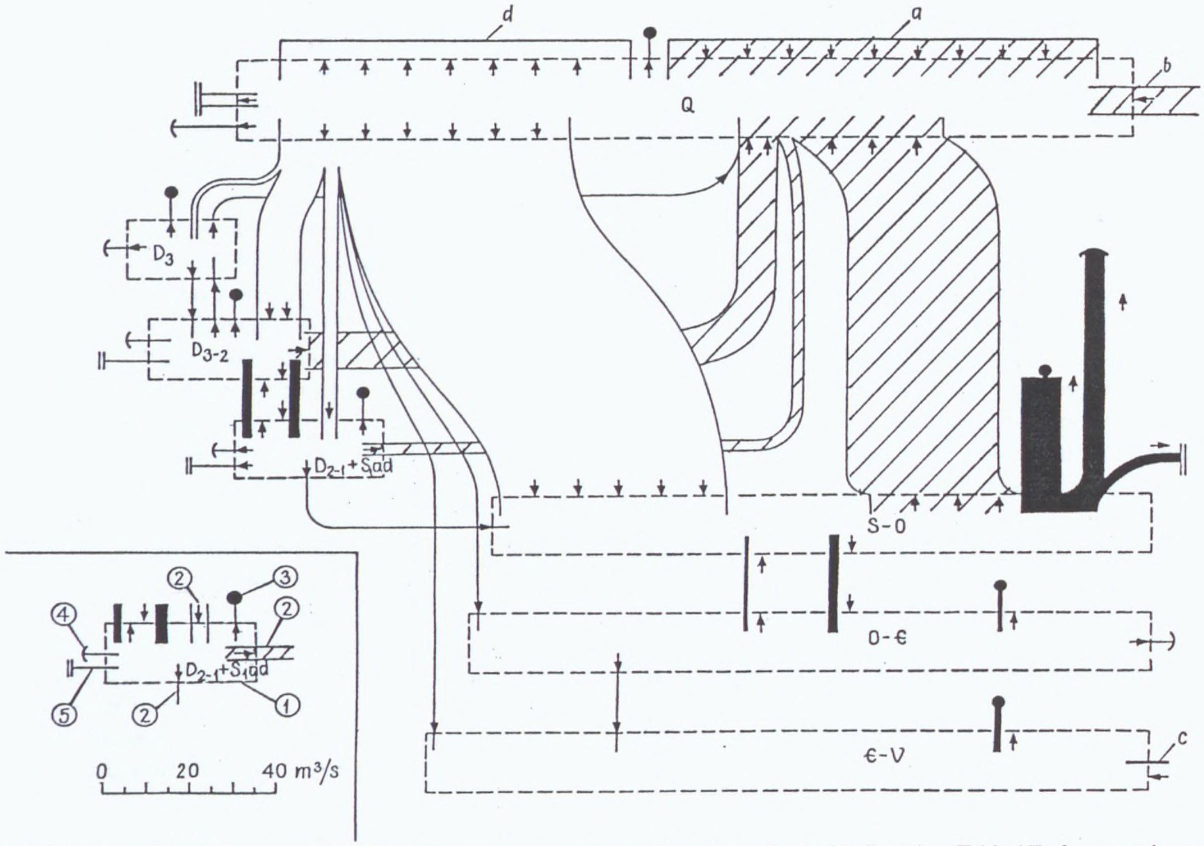


Fig. 103. Sketch of groundwater budget: 1- aquifer or aquifer system with hydrogeological indices (see Table 17); 2 - groundwater flow to the Quaternary cover (striped) and out of it (unstriped), the arrow shows the flow direction, discharge corresponds to the width of the flow shown in the figure (see the scale, m^3/s); 3 - pumpage of groundwater; 4 - discharge through springs; 5 - direct discharge to the sea; a - recharge from precipitation; b - induced recharge from streams; c - induced recharge from the Gulf of Finland; d - discharge to the channel network. *Compiled by L. Vallner.*

mainly from downward flow reaching 150,000 m^3/d which comes from the overlying Silurian-Ordovician aquifer system. Of the above amount of inflowing water, one third recharges the underlying Cambrian-Vendian aquifer system, one third rises up into the overlying Silurian-Ordovician aquifer system in Lower Estonia and one third is pumped out or flows into the sea.

Pumpage from the Cambrian-Vendian aquifer system was approximately 110,000 m^3/d in 1995. About half of it leaks through the Lükati-Lontova aquitard from the Ordovician-Cambrian aquifer system, and another half has formed on account of lateral flows coming from the side of the Gulf of Finland and central Estonia.

Fluctuation of the groundwater table under natural conditions

Natural fluctuation of the groundwater table has been investigated in many places all over Estonia. The statistical analyses of observation data have shown (Vallner 1982) that the groundwater table in the Quaternary deposits covering the outcrop of the Devonian rocks is 0.1 m higher in the beginning of the year than the annual average level of the water table. Later on, the groundwater table is lowering more or less evenly until the beginning of March when it is below the mean level by 0.05 m. When the air temperature rises above 0°C, the melt-water will percolate into the soil and the spring phase of intensive infiltration starts, lasting until snow has melted in the last decade of April. The amplitude of the spring rise of the groundwater table is 0.4 m whereby the maximum point exceeds the annual mean level by 0.35 m. In late spring and in summer, the amount of groundwater mostly decreases due to the intensive evapotranspiration of the soil moisture. As a result, the groundwater table will lower during about 140 days until the first or second decade of September when the decrease of evapotranspiration caused by lowering of the air temperature will be balanced with infiltration. The amplitude of groundwater level lowering is 0.6...0.7 m in the warm period. The minimum point of the groundwater table is below the annual mean level by 0.3 m in September. The summer lowering of the groundwater table can be retarded or even changed to rising because of occasional rain periods. Intensive infiltration recurs in autumn when a systematic rainfall starts and evapotranspiration is low. Then the groundwater table rises until the soil will freeze in December. During a cold winter the amount of groundwater predominantly decreases due to restricted infiltration.

The fluctuation of the groundwater table in western Estonia is very similar to that in southern Estonia as described above. Only the fluctuation amplitudes in southern Estonia are by 0.1 m less in winter and spring. It may be explained with the higher air temperature in winter, owing to which a portion of melt-water percolates into the soil or discharges directly to stream channels before the main thawing period starts in spring.

The annual amplitude of the groundwater table fluctuation is significantly greater in the karstified and fissured carbonate bedrock of northern Estonia. The spring rising reaches 0.7 m and the summer lowering is about 0.9 m. In the areas of abundant karst phenomena, the annual amplitude of fluctuation can range from 4 to 6 m, exceeding occasionally even 10 m. The spring maximum point and the summer minimum point arrive by one decade earlier than in southern Estonia. Such relatively great fluctuation amplitudes are caused by the karst cavities in the carbonate bedrock which accumulate a significant amount of water in spring, but this water quickly discharges to streams in summer.

Seasonal fluctuations of the water table are remarkably small in peat. The spring rising amplitude does not exceed 0.1 m and the summer lowering averages 0.2 m. The winter lowering lasts until the third decade of March and the spring maximum point arrives in the middle of May.

In confined bedrock aquifers the seasonal fluctuation of the head is commonly similar to the fluctuation of the groundwater table, but the amplitude decreases with depth. Owing to the intensive pumpage, the character of natural seasonal fluctuation can be more or less perverted.

COMPOSITION AND PROPERTIES OF GROUNDWATER UNDER NATURAL CONDITIONS

Zone of active water exchange

Infiltration water, comprised in the active water exchange zone of the Estonian groundwater system, obtains the chemical composition typical of groundwater mostly in the aeration zone (Table 19, Fig. 95). The upper 30...50 metres of the active water exchange zone are characterised by oxidized state, while in the lower part a transition from oxidizing to reducing conditions takes place (Põllumajanduslik ... 1994). The passive water exchange zone is entirely under reduced state.

In the active water exchange zone, calcium and magnesium carbonates are practically the only dissolved compounds. Therefore, regardless of the lithological compositon and the redox state of the groundwater the HCO_3-Ca-Mg (frequently also HCO_3-Mg-Ca) type of groundwater is formed, with the content of dissolved mineral salts under natural conditions being 0.1...0.6 g/l, most commonly 0.3-0.4 g/l. The content of free CO_2 in the upper part of the active water exchange zone is prevailingly 20...30 mg/l (the boundary content limits are 0.5...50 mg/l, occasionally even 100 mg/l). With the pH values 7.2...7.6, the concentration of balanced HCO_3^- in the water is 200...400 mg/l. When carbonates dissolve, then together with HCO_3^- also Ca^{2+} and Mg^{2+} reach water in proportional amounts (most frequently 40...95 and 11...30 mg/l, respectively). Besides, the water is enriched with Na^+, Cl^- and SO_4^{2-} (2...20, average 10 mg/l) originating from precipitation or soil.

In the active water exchange zone, neither the groundwater composition nor the amount of dissolved mineral salts is controlled by the lithological composition of rocks. Evidence is derived from the uniform chemical composition of water in springs and the amount of mineral salts dissolved in water. As an exception serves the water stored in the Quaternary sands, in which the proportion of quartz reaches 90%, and also the water stored in Devonian sandstones in southern Estonia in the areas where the Devonian is covered by sand, not by till. In that case the total content of dissolved salts (TDS) may reach 0.1...0.2 mg/l. This kind of water is unsaturated with carbonates, and has maintained the potential ability of dissolving carbonates. In practice, the cases are known, when groundwater in sandy areas has corroded concrete well curbs. This kind of water is used for removing scale from steam boilers.

Table 19. Chemical composition of precipitation in Estonia (after Karise 1965 and Simm 1975)

Type of precipitation	pH	Dissolved mineral salts, mg/l	Content of ions *, mg/l								Chemical type of water**
			Ca^{2+}	Mg^{2+}	Na^{+}	K^{+}	HCO_3^-	SO_4^{2-}	Cl^-	NO_3	
Snow water (12 samples)	4.4-7.0	6.4-34.6 18.2	0.2-8.0 3.5	0.0-1.0 0.4	0.0-2.5 0.9	0.3-1.2 0.6	0.0-23.2 8.2	0.0-15.8 4.9	0.7-6.0 2.0	0.0-4.0 1.4	HCO_3-Cl-Na-Ca HCO_3-SO_4-Ca-Mg SO_4-HCO_3-Ca SO_4-HCO_3-Cl-Ca *etc.*
Rain water (21 samples)	4.4-6.2	4.8-45.5 31.5	0.2-7.4 4.1	0.1-3.7 1.5	0.0-5.5 3.0	0.0-3.7 1.4	0.0-35.0 17.1	2.3-10.5 4.5	0.5-5.5 4.8	0.0-8.0 3.0	HCO_3-Cl-Na HCO_3-Cl-Ca-Na HCO_3-Cl-Na-Ca-Mg HCO_3-SO_4-Ca-Mg-Na *etc.*
Average in precipitation		23.0	3.7	0.8	1.7	0.9	11.2	4.6	3.0	2.0	

* In numerator - boundary content values, in denominator - mean content.
** The chemical type of water considers the content of major macrocomponents, the ions are given in the descending order of content values

Groundwater in the active water exchange zone is generally weakly alkaline (pH = 6.8...7.6). The unsaturated water in sandy regions is slightly acid (pH = 5.5...6.5). The water in bogs is acid (pH = 3.0... 5.0).

In the upper part of the active water exchange zone, the groundwater always contains free oxygen (O_2). Its content is much the same as in the surfce water, which in Estonia is 8...12 mg/l, as an average (Simm 1975). Due to the presence of free oxygen, the redox potential (Eh) of the water is always positive, because even with a small amount of free oxygen available, Eh cannot be less than +0.17...+0.18V (Посохов 1975). The higher the concentration of free oxygen in water, the higher the redox potential (maximum +0.7V). Generally, the water under oxidized state does not contain iron, because Fe-oxides and -hydroxides are insoluble in water, which means that Fe comprised in water-bearing rocks does not reach groundwater. However, in several cases Fe^{2+} (0.7...5.0 mg/l) has been determined in the water of bored wells tapping the Devonian sandstones or Silurian and Ordovician carbonate rocks, and in several cases also in the water of springs flowing out from Devonian sandstones which is indicative of reduced state at that depth. Aqueous environment, where pH = 7.0 and Eh < +0.3 V, is reducing in respect of iron. The latter stays dissolved under such conditions, although there is a small amount of free oxygen available in the water (Carrels & Christ 1965, Шварцев 1982).

In natural, uncontaminated groundwater which is in oxidized state the content of NO_3^- is commonly 5...6 mg/l. In the spring water flowing out from Devonian sandstones the concentration of NO_3^- is only 1...3 mg/l, in the water of excessively damp areas and peatlands it is less than 1 mg/l. This is due to reducing conditions under which, as a result of denitrification, part of the initially dissolved NO_3^- has been reduced to free oxygen (N_2) which volatalizes (Põllumajanduslik... 1994).

In northern Estonia, in the areas with a thick (up to 100 m) Quaternary cover where sediments contain buried organic matter, emissions of burning gas from bored wells have sometimes been recorded (Keri, Prangli and Mohni islands, Viinistu, Püssi). The gas comprises methane, hydrogen, nitrogen, hydrogen sulphide and, to a lesser extent, also helium, argon, oxygen and carbon dioxide and other compounds (Войтов и др. 1982). Frequently, there are gas emissions from the bored wells tapping the bedrock. In this gas the main component is free nitrogen (N_2) which accounts for 80...90% of gas volume. Sometimes, hydrogen sulphide is emitted together with the gas. In places, particularly in northern and southwestern Estonia, the water of borings contains helium in a rather high concentration (up to 0.6 ml/l) which originates in the crystalline basement and reaches groundwater through tectonic disturbances (Тибар 1987).

Zone of passive water exchange

A great part of the Estonian water-bearing formation is situated in the passive water exchange zone with reduced state. Due to combined effect of several factors, such as the very slow groundwater flow, connate water occasionally present in rocks, soluble mineral salts, *etc.*, the chemical composition and TDS in groundwater in the passive water exchange zone differ with regions. The HCO_3-Mg-Ca water with the TDS ranging from 0.5 to 0.6 g/l is of limited distribution and occurs in the Middle-Lower-Devonian aquifer system along Pärnu - Tartu - Viljandi line. The HCO_3-Cl-Na-Mg-Ca water with the TDS in between 0.3...1.5 g/l is more widespread. It occurs in central and southern Estonia in the Middle-Lower Devonian, Silurian-Ordovician and Ordovician-Cambrian aquifer systems. The Cl-HCO_3-Na-Ca and Cl-HCO_3-Ca-Na water with the content of dissolved mineral salts 0.4...1.0 g/l is stored in the Cambrian-Vendian aquifer system in northern Estonia.

In several places around Tallinn, the content of $\delta^{18}O$ in this water is -18...-22‰, but in the area of Loksa, Võsu, Kunda, Toila and Salutaguse it is -11.9...-16‰ (Vaikmäe & Vallner 1989, Savitskaja & Viigand 1994). In all likelihood, in the former case we have a typical glaciogenic palaeowater, in the latter case the glaciogenic water seems to have mixed with infiltration water formed under the conditions of moderate climate.

The Cl-Na, Cl-Na-Ca and Cl-Ca-Na water with the TDS from 2 to 22 g/l is widespread in the passive water exchange zone. This type of water has been established in the Ordovician strata on Hiiumaa Island (Kärdla), Ordovician-Cambrian aquifer system in southern, Cambrian-Vendian aquifer system in northeastern, southwestern and southeastern Estonia, on the islands of Saaremaa and Ruhnu and in the Lower Proterozoic strata in northern and southwestern Estonia.

The SO_4-Cl-Ca-Na water with the TDS up to 4.6 g/l, which very rarely occurs under such conditions as prevailing in Estonia, is found at a depth of 260 m at Värska in the Middle-Lower Devonian aquifer system (Fig. 94). Formation of sulphate-rich groundwater is due to the occurrence of gypsum in those layers south of Värska. Since the depth interval 250...800 m is charactericed by reduced state, the water there does not contain O_2 or NO_3^-; SO_4^{2-} is absent or present in a very low amount (1...5 mg/l). The content of Fe^{2+} is occasionally very high (up to 10 mg/l) and pH of deep water is usually 7.8...8.5. One reason is the low concentration of free CO_2 (often 0...2 mg/l, occasionally up to 6 mg/l). The content of HCO_3^- is also low, commonly 90...120 mg/l, seldom more.

Total hardness of Estonian groundwater differs in a wide range: it is soft in non-carbonaceous sediments, hard or very hard in carbonate-rich strata. Total hardness of bog water is 1...25 mg/l as $CaCO_3$, the hardness of water stored in sands under the influence of bog water is 20...85 mg/l, in extensive sand areas 65...195 mg/l and in sedimentary rocks 145...360 mg/l. Total hardness of mineral water with the TDS reaching 22 g/l, may be up to 4250 mg/l.

Microelements

The concentration of microelements in the water stored in oxidized state in the active water exchange zone is generally very low. The low content of some physiologically important microelements, such as fluor and iodine in drinking water, may cause health troubles. The incidents of endemic struma and caries, in some regions more numerous than in others, are associated with the low content of iodine and fluor in drinking water, respectively (Куйк 1961). The groundwater, which occurs under reducing conditions in the passive water exchange zone, is richer in microelements. In some places the content of microelements is even in excess of the optimum value established for drinking water. For instance, in the Silurian-Ordovician aquifer system the content of fluorides (F^-) is 5.5...7.2 mg/l in western and southwestern Estonia, 3.2 mg/l in Tartu and 2.4 mg/l at Abja, being well in excess of the standard established for the drinking water in Estonia and exceeding the level permitted by the World Health Organisation, which is 1.5 mg/l (Куйк 1963, Viigand & Vatalin 1992, Guidelines... 1993). This explains the incidences of fluorosis in these areas (Куйк 1961). In Pärnu, the content of fluorides in the Silurian-Ordovician aquifer system is almost optimal - 0.8...1.0 mg/l (Boldõreva *et al.* 1993), being elsewhere well below it.

In the Ordovician-Cambrian aquifer system, the content of microelements is low in the areas where the TDS is less than 1.0 g/l. South of the Pärnu - Viljandi - Tartu line where the Ordovician-Cambrian aquifer system stores Cl-Na-(Ca) water with the content of dissolved mineral salts up to 14 g/l (Ikla), the content of microelements is higher. Thus the contents of cadmium (Cd^{2+}), lead (Pb^{2+}) and lithium (Li^+), in excess of the norms established by the drinking water standard in Estonia, have been registered at Värska (Fig. 94). Since the water derived at Värska is used as a mineral water, and half of the amount produced is dilluted, then it does not pose any threat to human health.

In the groundwater stored in the Cambrian-Vendian aquifer system east of Tallinn, the concentration of iodides (J^-) is heightened, reaching 120...280µg/l (Куйк 1961). This water, if used for drinking, will cover the need for iodine in the population. Generally, the content of all microelements increases in northern Estonia towards the east in this aquifer system. In northeastern Estonia, in some bored wells the content of cadmium (Cd^{2+}), lead (Pb^{2+}) and lithium (Li^+) is slightly in excess of the norms. In this aquifer system the concentration of microelements is at its highest at Värska in southeastern Estonia where in Cl-Na-Ca water the TDS ranges from 6.0 to 19.0 mg/l. In this water the content of cadmium (Cd^{2+}), lithium (Li^+), manganese (Mn^{6+}) and lead (Pb^{2+}) is in excess of permitted boundary limits. This type of Värska water is not used directly for drinking, but due to its medicinal effect it is used in baths and as curative drinking water. The origin of the microelements in the water of Ordovician-Vendian and Cambrian-Vendian aquifer systems is not yet unambiguously clear. They may partly reach the water from the steel casing of wells, continuously corroded by salt water stored in reduced state, however, part of microelements is evidently of natural origin (Savitskaja & Viigand 1994).

In several regions of Estonia, the heightened concentrations of bromides (Br^-) have been detected (Fig. 94, Table 20):

- in the Ordovician-Cambrian aquifer system: 31 mg/l at a depth of 645...658 m at Ikla and 50...54 mg/l at a depth of 707...784 m on Ruhnu Island;
- in the Cambrian-Vendian aquifer system: 13 mg/l at a depth of 540...555 m at Kuressaare, 16...17 mg/l at a depth of 520...535 m and 51...56 mg/l at a depth of 540...600 m at Värska;
- in the water of the crystalline basement 51...61 mg/l at Hirvli and Pudisoo.

Mineral water

The groundwater in which the TDS is 2 g/l or more is rated as mineral water in Estonia. A. Verte was the first to predict the existence of different types of mineral water in Estonia (Photo 33). The first mineral water deposit was discovered at Pärnu in 1959 by the researchers of the Geological Survey of Estonia. For the purposes of structural geological investigations a well, deeper than 500 m, was sunk in Pärnu. It tapped the Lower Proterozoic crystalline bedrock and yielded Cl-Na water with the TDS about 22 g/l (Vingissaar 1978). Mineral water has been found in 16 different sites all over Estonia. At Värska, it occurs in four aquifers, at Kuressaare and Arumetsa in two aquifers (Table 20, Fig. 94).

In Estonia the bottling and marketing of mineral water was started in 1968. Currently, the Cl-Na-Ca water of the Ordovician-Cambrian aquifer system obtained at Värska with the content of TDS ranging from 2.0 to 2.2 g/l is bottled. The mineral water derived from Värska and Kuressaare is used for curative purposes both for drinking and in baths. In earlier years, the mineral water derived at Arumetsa, Häädemeeste, Ikla, Kuressaare and Kärdla was also bottled and sold. During the bottling, the water is often enriched with carbon dioxide (5...7 g/l). The reserves of Estonian mineral waters are estimated at about 6000 m^3/d. Total reserves of salty and salt water are very large in Estonia and may amount to hundreds of cubic kilometres.

Temperature

The temperature fluctuations caused by meteorological factors occur in the upper part of Estonia's water-bearing formation with an average thickness of some 18 m (Юрима 1984). The maximum thickness of this zone, marking the depth of the so-called neutral layer, reaches 30 metres in the Pandivere Upland. In western Estonia and on islands, the thickness of the zone subjected to annual temperature fluctuations is 10...15 m and in the coastal plain of northern Estonia it is 5...11 m. The temperature of water in springs and up-to-30-m-deep wells in the Pandivere Upland ranges from +4°C to +6°C, elsewhere it is +6...+7°C. The water is coldest in March-April when a great quantity of melt-water percolates into the ground, and warmest in September - October. The temperature in the neutral layer is stable, being prevailingly +7°C, in some places also +6°C. Downwards the temperature rises steadily (Fig. 104). At a depth of 50 m in the area of Silurian and Ordovician carbonate rocks, the temperature is +6.2...+6.3°C, in southwestern Estonia under Devonian sandstones +9.4...+9.5°C and in Lontova clays on Estonia's north coast +7...+8°C. At a depth of 100 m the average temperature is +7.6°C. In the uplands with intensive groundwater recharge the temperature is lower than the average (+6.5...+7.2°C), whereas in northeastern Estonia and on the Pärnu Lowland it is higher, being +9.2...+9.5°C and +9.8...+10.2°C, respectively. At a depth of 200 m in the carbonate bedrock of central Esto-

Photo 33. Artur Verte (1901-78), the most prominent Estonian hydrogeologist after World War II. He predicted the existence of different types of mineral water in Estonia and compiled several hydrostratigraphical schemes.

Table 20. Estonian mineral waters

Number on Fig. 17	Location of bored well	Depth interval of water, m	Aquifer or aquifer system (table 17)	Chemical type of water; content of dissolved mineral salts, g/l
1.	Põlva County, Värska	259-314 451-500 520-535 575-595	D_{2-1} (Pärnu Stage) O-€ €-V (Voronka) €-V (Gdov)	SO_4-Cl-Ca-Na; 4,6 Cl-Na-Ca; 2,0...2,2 Cl-Na-Ca; 5,6...6,0 Cl-Na-Ca; 15,0...19,0
2.	Pärnu County, Ikla	645-658	O-€	Cl-Na-Ca; 14,0
3.	Pärnu County, Arumetsa	538-597 602-632	O-€ $€_1$	Cl-Na; 5,0 Cl-Na-Ca; 3,2
4.	Pärnu County, Häädemeeste	540-600	O-€	Cl-Na-Ca; 5,4
5.	Pärnu	ca 500	PR_1	Cl-Na; 21,7
6.	Hiiu County, Kärdla	290-336	O_2id	Cl-Na-Ca; 3,0...3,5
7.	Saare County, Kuressaare	458-502 540-555	$€_1$ (Soela) $€_1$ (Sõru)	Cl-HCO_3-Na; 2,1 Cl-Na-Ca; 3,8...4,0
8.	Saare County, Ruhnu Island	707-784	O-€	Cl-Ca-Na; 17,0
9.	Harju County, Hirvli	271	PR_1	Cl-Na; 11,6
10.	Harju County, Pärispea	240	PR_1	Cl-Ca-Na; 20,4
11.	Harju County, Pudisoo	204	PR_1	Cl-Na; 10,7
12.	Harju County, Rammu Island	ca 153	€-V+ PR_1	Cl-Na; 5,0...5,7
13.	Lääne-Viru County, Käsmu	187	PR_1	Cl-Na; 4,5
14.	Lääne-Viru County, Põhja-Uhtju Island	ca 126	€-V+ PR_1	Cl-Na; 5,0...5,7
15.	Ida-Viru County, Narva-Jõesuu (Meriküla)	ca 215	€-V (Gdov)	Cl-Na; 2,6...2,9
16.	Võru	504-535	O-€	Cl-Na-Ca; 3,2

nia, the temperature is only +8.0°C which shows that the subsurface groundwater percolates quickly to a greater depth. At the same depth in northeastern Estonia, the temperature is +14.0...+14.5°C and in southwestern Estonia +10.3...+10.6°C. At this depth, the highest temperature (+15...+16°C) has been measured in the crystalline basement of northeastern Estonia (Юрима 1984).

The mean value of geothermal gradient in Estonia is 1.2°C/100 m, and it increases with depth being 1.0°C/100 m in Silurian-Ordovician carbonaceous rocks and Devonian sandstones, 2.0...3.5°C in deeper bedded Cambrian sandstones, and 5.0...6.0°C (average 4.0°C) per 100 m in the underlying Cambrian clays and silts (Lükati-Lontova regional aquitard) (Юрима 1984).

MAN-MADE CHANGES OF GROUNDWATER QUALITY

Pollution load on groundwater

As a result of extensive economical activity and high vulnerability of the uppermost aquifer, the shallow groundwater is in places heavily polluted and therefore unfit for drinking. The point-pollution sources are different constructions and pipelines in poor condition, such as boilerhouses, fuel storages, storages of chemicals and manure, settling basins, sewerage, leaching beds, gas stations, landfills, burial places of domestic animals, *etc*. Extensive fuel leakages have occurred on military airfields (Fig. 105) and in railway junctions. The majority of gas stations that were state-owned during the Soviet period contaminate environment with oil products. The asphalt concrete plants using primitive equipment and technology (Tiitso, Riisipere, *etc*.) are also sources of extensive pollution of groundwater.

In the Ida-Viru County, essential point-pollution sources are the spoil heaps of oil shale mines and ash plateaus of thermal power plants (Fig. 105). The groundwater leaching from oil shale ash is polluted with phenols and compounds of heavy metals, its pH-value is 12 and even more. Extensive surface water and groundwater pollution has been caused by cracking processes accompanying fires in oil shale mines (Vallner & Sepp 1993, Vallner 1994).

Non-point pollution is caused by the misuse of mineral fertilizers on arable lands and the underutilization of slurry from pig farms, but also by the treatment of fields with toxic chemicals. In the Ida - Viru County powerful thermal power plants annually eject into the atmosphere 120,000 tonnes of fly-ash and 80,000 tonnes of aerosol fractions containing harmful elements (S, F, Cl, V, Cr, Ni, Br, Sb, Cd, Pb, As) and radioactive isotopes (Õispuu & Rootamm 1994). The ash and gaseous pollutants transported by wind over large areas percolate together with rainwater into the groundwater. Higher concentrations of these pollutants in the atmosphere promote formation of sulphuric and nitric acids. The impact of acid rains is mitigated by fly-ash from thermal power plants which may spread to a distance of 100...150 km (Frey *et al.* 1987).

In 1994, about 14,000,000 tonnes of solid waste was generated in Estonia; of that amount 46 % by the oil-shale-based energy production, 40 % by oil shale mining, 8 % by the chemical industry and only 6 % in other spheres (Keskkond... 1995). Of 1,962,000,000 m^3 of waste water produced, 70 % was the cooling water heated through use in an industrial process which raised the temperature of both surface and groundwater. The amount of waste water needing purification was 380,000,000 m^3. Approximately 1% of waste water (1,800,000 m^3) was

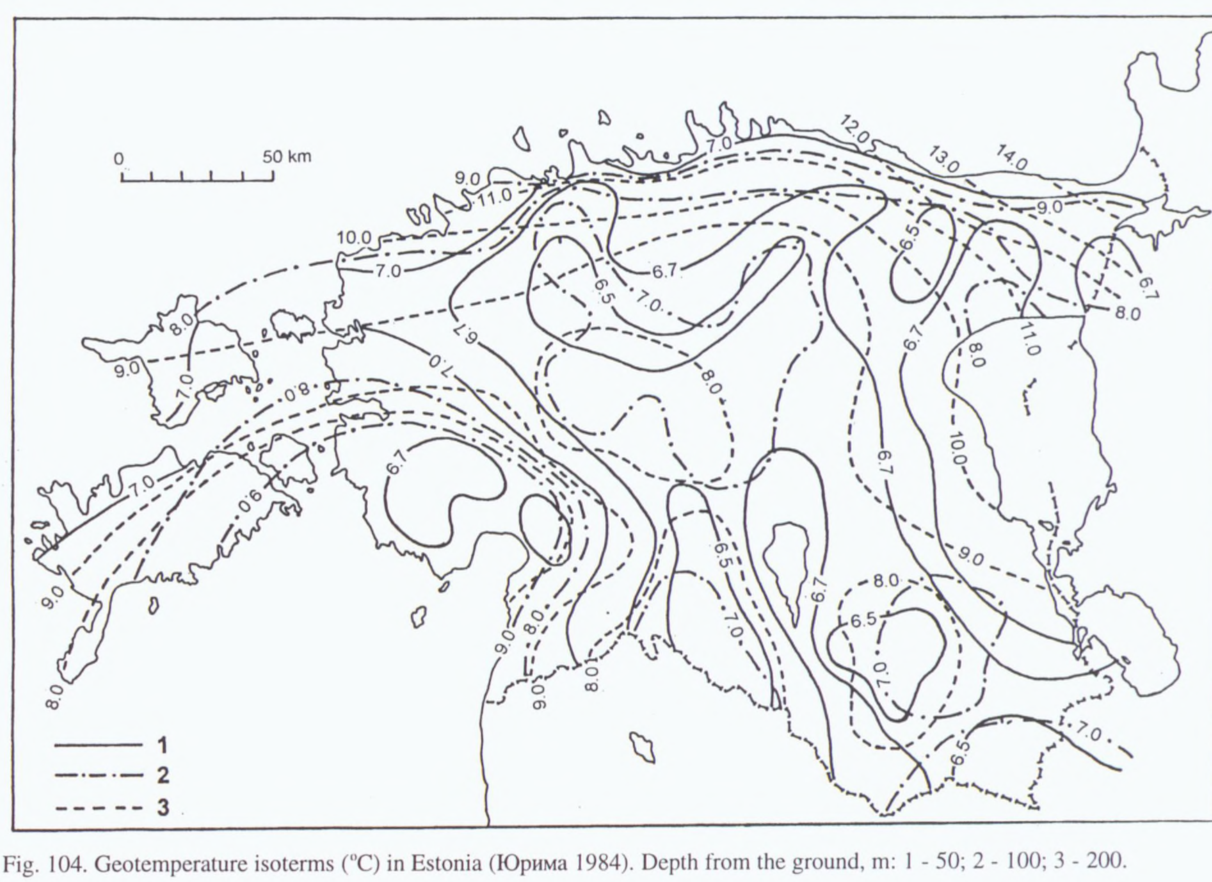


Fig. 104. Geotemperature isoterms (°C) in Estonia (Юрима 1984). Depth from the ground, m: 1 - 50; 2 - 100; 3 - 200.

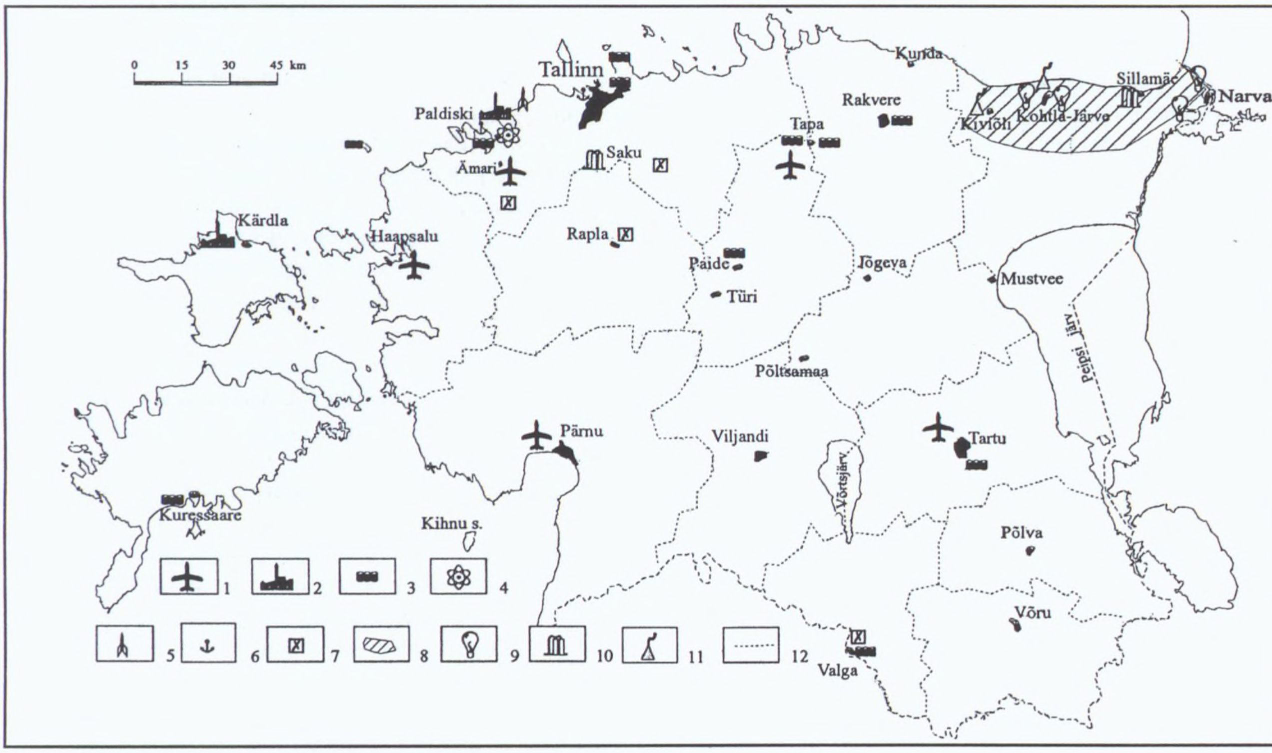


Fig. 105. Significant point-pollution sources: 1- airfield; 2 - boiler house; 3 - fuel depository; 4 - nuclear reactor, liquidated; 5 - missile base, liquidated; 6 - harbour; 7 - asphalt concrete works; 8 - industrial region of North-East Estonia; 9 - power station; 10 - radioactive waste depository; 11 - oil shale processing enterprise; 12 - boundary of the county. *Compiled by the joint-stock company MAVES.*

discharged directly into soil and groundwater. Compared to 1989-90, the total industrial capacity has decreased and the technology has improved. As a result, the amount of solid waste and waste water has decreased by about 20 and 40 %, respectively.

In 1994, a total of 41,000 tonnes of mineral fertilizers (as N) and 1,100,000 tonnes of manure were used in Estonia, which is about 6.5 times less than during 1988-89. The average pollution load of solid waste was of 320 t/km^2, and that of untreated waste water 8,700 m^3/km^2. The amount of mineral fertilizers (as N) was 0,04 t/ha and that of manure 1.1 t/ha. The pollution load was highest in the Ida - Viru County due to oil shale enrichment waste and ash of thermal power plants. In 1994, the average pollution load of solid waste in this region was 3,840 t/km^2 (1.6 times less than in 1990), while the load of waste water in need of purification was 69,000 m^3/km^2.

In northeastern Estonia, the Ordovician carbonate rocks covered with a thin Quaternary mantle are intensively polluted with shale oil and phenols. In the undermined area (about 200 km^2) the lowering of groundwater level has resulted in oxidation of pyrite in the aeration zone, due to which the content of sulphates in groundwater has increased up to 650 mg/l (under natural conditions it is less than 20 mg/l).

Groundwater is polluted with oil products at Tapa (Fig. 105) and in its surroundings. Since 1966, several big accidents of fuel tanks and constant leakage of fuel pipelines took place on the former military objects of this region. Particularly intensive pollution with oil products has been recorded on the former military airfield, where the stopcocks, valves and pipelines of fuel tanks were leaking continuously. Near the railway oil receiving centre, an oil lake had formed on the surface. Due to carelessness or leakage from boiler-houses and fuel tanks, groundwater is polluted with oil products also in Tallinn, Tartu, Rakvere, Kohila, Rapla, Tamsalu, Aruküla and several other places. In these regions, the water supply is more or less disturbed.

In the area of the former galvanic departments and in the surroundings of landfills, groundwater contains heavy metals. Due to the diffusion of filtrates from Tuula landfill near Keila, the concentrations of hazardous compounds in the neighbouring wells are the following: Pb^{2+} – up to 0.13 mg/l, Mn^{2+} – up to 1.4 mg/l, Cd^{2+} – up to 0.0035 mg/l (Tennokesse *et al.* 1992); the maximum permissible concentrations of these elements in drinking water are 0.01, 0.1 and 0.003 mg/l, respectively (Eesti standard...1995).

In the regions of agricultural activity, groundwater is contaminated mainly by nitrogen compounds (Fig. 106), but the concentration of chlorides and sulphates has also increased to 40...60 mg/l (under natural conditions 20 mg/l), as an average. In the surroundings of Viiratsi piggery near Viljandi, the concentrations of ions are the following: NH_4^+ – up to 80 mg/l, Fe^{2+3} – up to 128 mg/l, Cl^- – up to 262 mg/l and SO_4^{2-} – up to 330 mg/l (Boldõreva *et al.* 1992). The high content of ammonium and iron, as well as the lack of nitrates indicate reducing conditions. In Estonia, the maximum permissible content of NO_3^{2-} in drinking water is 45 mg/l (Eesti standard...1995). In 1990, the groundwater did not meet the requirements established for drinking water in 40...70 % of the total number of shallow wells (depth to 15 m) in southern, 20...40 % in northern, 30...60 % in central Estonia and in 10 % of wells on the islands of the West-Estonian Archipelago (Põllumajanduslik... 1994). In the wells with a depth of 30...100 m, the content of nitrogen compounds was remarkably lower, especially in southern Estonia (Tennokesse *et al.* 1992).

Groundwater protection

Due to a thin aeration zone (mainly 1.5...3 m) the unconfined groundwater is generally weakly protected against surface pollution. The problems are acute in northern and central Estonia where the Quaternary cover is less than 2 m thick or practically lacking (alvars) and the fissured and highly cavernous limestones crop out on the surface.

On the basis of numerous experimental data and calculated infiltration velocity of waste water, certain criteria for the assessment of the degree of natural protection of shallow groundwater from agricultural and municipal pollution have been worked out (Savitskaja 1987).

Groundwater is considered unprotected on alvars (the thickness of the Quaternary cover is less than 0.5 m) and in the areas where aquifers are covered with a up-to-2-m-thick layer of loamy sand (conductivity value K=0.1...0.5 m/d), or up-to-20-m-thick layer of sand or gravel (K=1..5 m/d). Groundwater is weakly protected when the thickness of the confi-ning layer of loamy sand ranges from 2...10 m, or that of the clay layer (K=0.0001...0.005 m/d) is less than 2 m. Groundwater is regarded as moderately protected when aquifers are overlain by a 10...20-m-thick layer of loamy sand, or with a 2...5-m-thick layer of clay, and protected when the thickness of the confining layer of loamy sand exceeds 20 m, or that of clay layer is over 5 m. In case of intensive water consumption, these criteria must be corrected, because under such conditions the transport velocity of pollutants in soil may increase noticeably.

The map showing the degree of shallow groundwater protection demonstrates that groundwater is weakly protected or unprotected in ca. 40 % of Estonia's area. A comparison of the maps of groundwater protection degree and nitrate concentrations (Fig. 106) has shown that the intensity of distribution of nitrates depends closely on groundwater protection degree.

Since 1991, the impact of agriculture on groundwater state has noticeably decreased due to manifold reduction of the use of fertilizers and decrease in the number of domestic animals. Long-time observations of 19 springs on the slope of the Pandivere Upland show that the content of nitrate in spring water has decreased which also indicates to general lowering of pollution load in this region (Savitski *et al.* 1996).

In 1994, a decree of the Minister of the Environment addressing reduction of pollution load, enacted restrictions to the use of fertilizers (Table 21) and the number of domestic animals by regions.

In Estonia's environmental strategy, priority has been given to the problems of groundwater protection, the most important being elimination of sources of groundwater pollution and regulation of groundwater use. The Water Code of Estonia and the Law of Sustainable Development enact general regulations for economically effective and environmentally sound use of water resources.

Table 21. Maximum permissible amounts of nitrogen given with mineral fertilizers

Region	Annual permissible amount, kg(N)/ha
Estonia, as a whole	100
Areas with karst and thin Quaternary cover	80
Area of morainic hills in southern Estonia	70
Islands	60

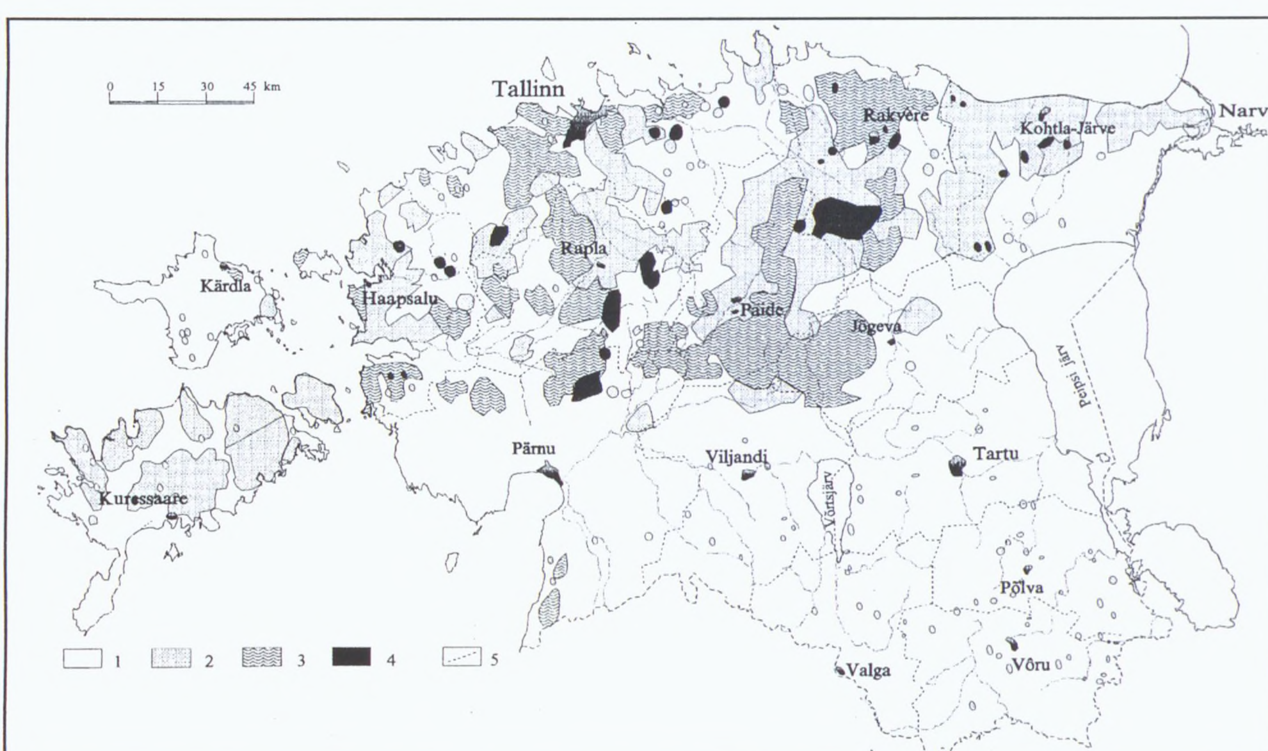


Figure 106. Distribution of nitrates in groundwater to a depth of 30...100 m from the ground in 1990: 1-4 - concentration of nitrates in groundwater, mg/l: 1 - <4; 2 - 4...20; 3 - 20...45; 4 - >45; 5 - boundary of the county. *Compiled by L. Savitskaja.*

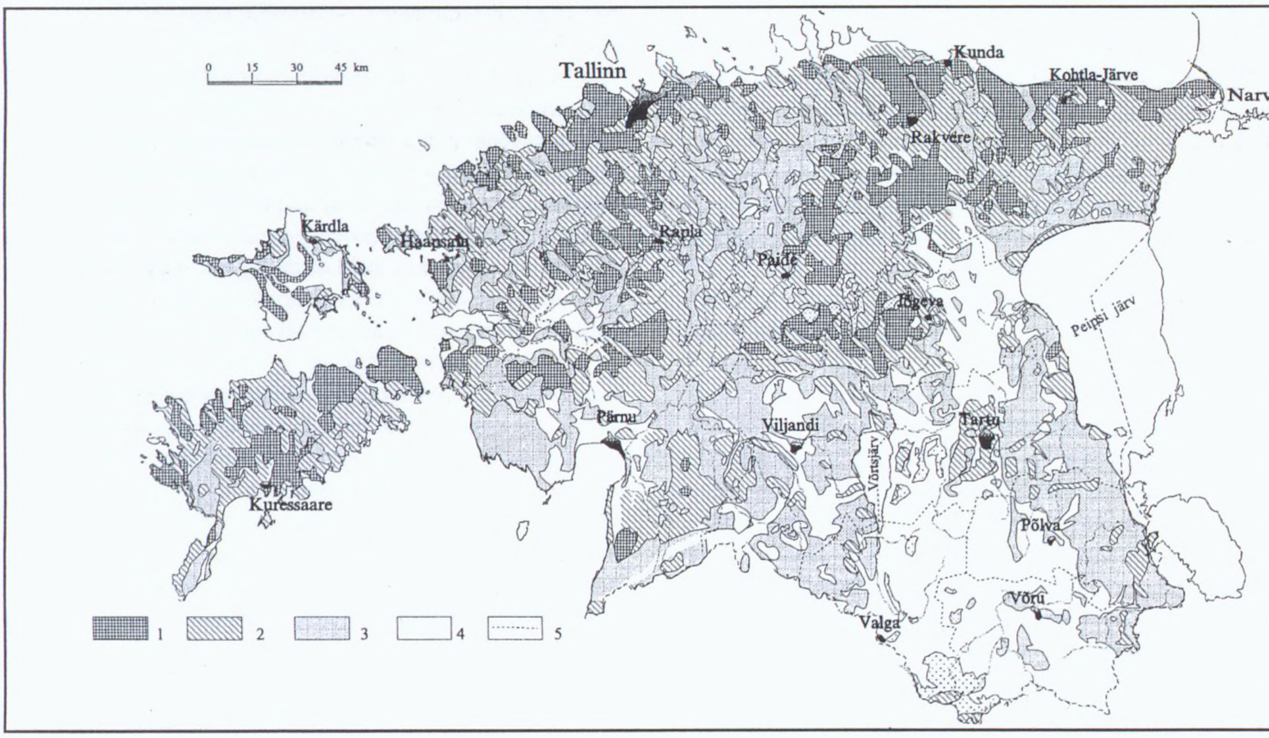


Fig. 107. Degree of groundwater protection: 1 - unprotected; 2 - weakly protected; 3 - moderately protected; 4 - protected. *Compiled by L. Savitskaja.*

GROUNDWATER EXTRACTION AND SAFE YIELD

Groundwater use

Until 1944, the groundwater was mostly used for domestic needs only. According to studies of J. Kark, there were only a few hundred bored wells, deeper than 50 m in Estonia which were used by public waterworks in central parts of towns and predominantly for the purposes of food industry. Pumping of water from a few production wells did not induce any cone of depression worth of mentioning. There was only one public water intake tapping the glaciofluvial water-bearing gravel and sand of the Meltsiveski buried valley which was put into operation in Tartu with the pumping rate of 12,000 m^3/d (Orviku 1946). The total number of wells was approximately 250,000. Predominantly these were domestic wells with a depth of up to 8 metres.

Since 1945, unbalanced industrialization, urbanization and militarization of Estonia was carried out by Soviet authorities. The need for a centralized water supply increased rapidly and the number of deep bored wells augmented from a few hundred in 1950 to 3300 in 1964 (Архангельский 1966). Since 1964, 200...300 new wells were bored every year, but at the same time there was a need to liquidate a lot of amortized wells. In 1991, the number of wells deeper than 60 m reached 8400 in Estonia.

The total groundwater extraction grew from about 30,000 m^3/d in 1950 to 470,000 m^3/d in 1991 (Boldõreva *et al.* 1993). During the first two decades, the rate of water extraction increased steadily, thereafter began to decelerate (Fig. 108). During 1964-94, the contribution of different aquifer systems in the total groundwater supply was following: Quaternary, Middle Devonian and Middle Devonian-Silurian aquifer system - 6...16% each, Silurian-Ordovician aquifer system - 22...25%, Ordovician-Cambrian aquifer system - 9...12% and Cambrian-Vendian aquifer system - 34...39% (Savitski *et al.* 1995). Consequently, the proportion of the aquifer systems exploitation did not change significantly during 30 years.

Until 1991, approximately 60% of total groundwater supply was consumed in towns, of that nearly half in Tallinn and in the Kohtla-Järve region; 15% in Tartu, 5% in Pärnu and the remaining 30% in other towns. In rural areas, the deep bored wells were predominantly used to supply the settlements and big livestock farms.

In 1995, about 290,000 m^3 of drinking water was extracted from bored wells, which is 60% less than in 1990 (Kivit *et al.* 1991 & Savitski *et al.* 1996). Approximately 80% of this water was spent for supplying of towns and industrial settlements, the remaining 20% was used in agricultural regions (Fig. 109). The decrease of groundwater extraction has been caused by the decline of industrial and agricultural production and by the more sustainable use of groundwater. As an average, 110,000 m^3/d was extracted from the Cambrian-Vendian aquifer system, 59,000 m^3/d from the Silurian-Or-

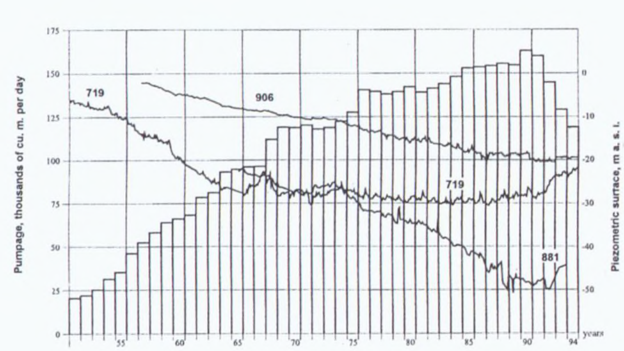


Fig. 108. Total pumpage from the Cambrian-Vendian aquifer system and change of the head in observation wells 719, 881 and 906 (see Fig. 102). *Compiled by L. Savitski.*

dovician aquifer system, and 17,000...47,000 m³/d from each of the remaining aquifer systems (Table 22).

Currently, groundwater forms 70% of consumed drinking water in Estonia. Only in Tallinn and Narva the consumption of purified surface water exceeds that of groundwater, while elsewhere in Estonia groundwater is the only source of public water supply. The major amount of groundwater is extracted from about 20,000 bored wells from which only a half are deeper than 20 m. The depth of bored wells, used in public water supply, remains predominantly in the interval of 80...150 m. However, in central Estonia some wells have a depth of up to 450 m.

Groundwater safe yield

In Estonia, the groundwater safe yield is considered to be an amount of groundwater which could be withdrawn during a calculation period without producing an unpermitted deterioration in the quality of water pumped. Since 1963, the groundwater safe yield has been determined on the basis of detailed hydrogeological investigations to get a clear conception about the perspectives of public water supply. To calculate the safe yield Q_s the modification of Theis' equation was mostly used (Бочевер 1968):

$$S=\sum_{i=1}^{n} \frac{Q_i}{4\pi T} W_i(u)$$

where s is a given suitable and really feasible drawdown in a given point A of wellfield; n is the total number of wells under consideration; Q_i is the given constant discharge of a well i; T is the transmissibility of the aquifer, and $W_i(u)$ is the so-called well function with the argument $u = r_i^2S/4Tt$ where r_i is the radial distance of the well i from the point A; S is the storage coefficient, and t is the time since beginning of pumping. The safe yield was calculated as a sum $Q_s = Q_1 + Q_2 + ... + Q_n$ by the trial and error method.

In calculating the safe yield, a series of simplifying assumptions, such as infinite areal extent of an aquifer, its uniform thickness, homogeneity, isotropy, etc., were posed. In spite of this, in most cases the calculated prognoses fitted quite well to observations, made later in operation of water intake. The necessary hydrogeological parametres for calculating the safe yield were taken from the materials of hydrogeological mapping of Estonia in scale 1:200 000 and partially 1:50 000 but also from the results of previous hydrogeological exploration, including pumping tests and groundwater quality analyses.

Valuable information was gained from the observation network of regular hydrogeological monitoring, which was established in 1946 on some water intakes in northern Estonia and in the region of oil shale mines. Later on the observation network of the state geological survey was steadily widened and developed: in 1990, it consisted of 846 observation wells (Kivit *et al.* 1991). The aquifer systems were monitored all over Estonia. Besides, 60...120 observations wells were bored on more essential water intakes and in mining districts (Tallinn, Kohtla-Järve, Vasavere) Groundwater level was measured at observation points once every three days and the chemical analysis of water was made once every three months to determine the content of 8...12 main components. The sanitary state of groundwater was steadily controlled by sanitary survey and the amount of consumed groundwater was registered by the owner of the well.

Until 1995, the safe yield of 131 well fields was estimated (Savitski *et al.* 1996) at about 550,000 m³/d which is nearly three times as high as the real extracted amount (Table 22, Fig. 109). About 40% of the safe yield associates with the Cambrian-Vendian aquifer system, while the Quaternary and Ordovician-Cambrian aquifer systems provide only 8 and 5%, respectively. The safe yield of the Quaternary and Silurian-Ordovician aquifer systems is rapidly recurrenting, but these systems are vulnerable to pollution and require extensive zones of sanitary protection. The safe yield of the Middle Devonian-Silurian, Ordovician-Cambrian and Cambrian-Vendian aquifer systems will be restored slowly, or it will never be restored (glaciogeneous water in the Vendian strata). To ensure sustainable use of the safe yield, the slowly recurrenting groundwater must be used only for drinking water supply.

The regional safe yield for main aquifer systems and bigger water intakes was simulated by means of an analogue computer in 1976. The total, so-called presumptive safe yield, determined by this method with relatively smaller accuracy, is 1,530,000 m³/d of which 42% falls to the Silurian-Ordovician aquifer system and 15% to the Cambrian-Vendian aquifer system (Table 22).

In 1994, about 70% of the total pumpage was got by intakes having an estimated safe yield and enfolding 925 production wells. The remaining portion of water was derived by 7,000 wells scattered over Estonia. Despite some deviations, the used safe yield generally meets the drinking water requirements (Eesti standard... 1995). The increased contents of iron and manganese have been found in the water of Devonian strata in southern Estonia, and the increased content of fluorine in the water stored in the Silurian strata in western Estonia. The water of some wells tapping the Cambrian-Vendian aquifer system contains iron, manganese and chlorine in excess of the sanitary norms (Kohtla-Järve, Jõhvi, Ahtme, *etc.*). To reduce the concentration of hazardous components, the groundwater must be treated before use.

Consequences of the intensive groundwater use

The most serious consequences of intensive groundwater use include the formation of regional depressions of potentiometric level which has caused cardinal changes in the direction and velocity of filtration flows in the lower portion of the water-bearing formation (Figs. 102, 103). As a result of heavy pumpage, the groundwater inflow into the strata increases. If in natural conditions the inflow into the Cambrian-Vendian aquifer system was less than 30,000 m³/d, then in 1990 it was 163,000 m³/d or even more. Drawdown contours show that half of the increased inflow is coming from the seaside. Consequently, the other half is supplied by downward, upward and southward lateral flows. The downward flow provides the Cambrian-Vendian aquifer system mostly with fresh water, but the rising flow may be connected with upconing of brackish water from the crystalline basement. Lateral flows conduce the transport of connate brackish water from the deeper portion of an aquifer or sea-water intrusion to groundwater intakes.

The sea-water intrusion into water intakes can take place first in Tallinn where a lot of production wells tapping the

Cambrian-Vendian aquifer system are situated close to the sea. However, in spite of several threatening prognoses the intrusion of sea-water into water intakes in Tallinn has not been observed yet. The increase of TDS in the water of some wells from 0.5 up to 1.3 g/l has been observed in the northern part of Tallinn, but the data of isotope analyses confirm that it is caused by the inflow of brackish water from the lower part of aquifer system or from the crystalline basement (Savitski *et al.* 1995). The potential intrusion of sea-water into coastal bored wells, tapping the Silurian-Ordovician aquifer system, is a serious problem also in Pärnu, Haapsalu, Kuressaare and some other places.

The potentiometric surface of the Ordovician-Cambrian aquifer system has dropped at least 7 metres in the central part of the Pandivere Upland due to regional impact of water consumption. Deep local depressions have also formed in Tartu, Rakvere and several other places (Figs. 102, 110). Because of the significant head withdrawal, pollutants can easily intrude into the Ordovician-Cambrian aquifer system from surface in Tallinn, Tapa and Kohtla-Järve.

To intensify the water supply of Kohtla-Järve urban area, a groundwater intake with the planned capacity of 25,000 m^3/d using the glaciofluvial aquifer of the Vasavere buried valley was put into operation in 1971. The intake was within the Kurtna Landscape Reserve featured by wooded hillocks and ridges with a lot of picturesque lakes, many of them with unique hydrobiological biocenoses. Unfortunately, already at the pumping rate of 10,000 m^3/d the water table of several lakes lowered below the acceptable minimum level and the unique biocenoses were irreversibly damaged (Mäemets 1987).

In Tartu, many historical buildings have been founded on timber piles which were initially submerged by groundwater. After 1960, the pumping from the Toome-Meltsiveski buried valley and bedrock aquifers was significantly increased which caused a drop of the groundwater level by 2.5...3.5 m in the old part of Tartu. The upper portion of timber piles remaining in the aeration zone were intensively attacked by the wood borer *Cossonus parallelepipedus* and began to decay quickly (Oll 1967). As a result, the buildings started to crack. Decaying of architectural memorials could have been avoided by correct groundwater pumping; now tens of millions of US dollars are needed for their reparation.

An uneven land subsidence occurs in the area of buried valleys permeating the territory of Tallinn (Арбейтер и др. 1982, Vallner & Lutsar 1966). The total subsidence amounted to 0.6...0.8 m and its maximum annual rate reached 36 mm in the vicinity of the trading port in 1964. Deformation of the ground caused the bench-marks to shift, spoiled the designed inclination of sewage collectors, created cracks in the walls of buildings. The subsidence was caused by the compaction of marine and glaciolacustrine clays confining the Cambrian-Vendian aquifer system in buried valleys. Due to intensive pumping from this aquifer system, the potentiometric surface was lowered by 20...25 m and the compacting intergranular pressure in the overlying highly compressible Quaternary aquitard increased. However, in 1972, the potentiometric surface of the Cambrian-Vendian aquifer system dropped below the bottom of the confining Quaternary aquitard and since then land subsidence has gradually expired. Calculations have shown that the maximum subsidence does not exceed one metre in the area of the old trading port of Tallinn (Валлнер

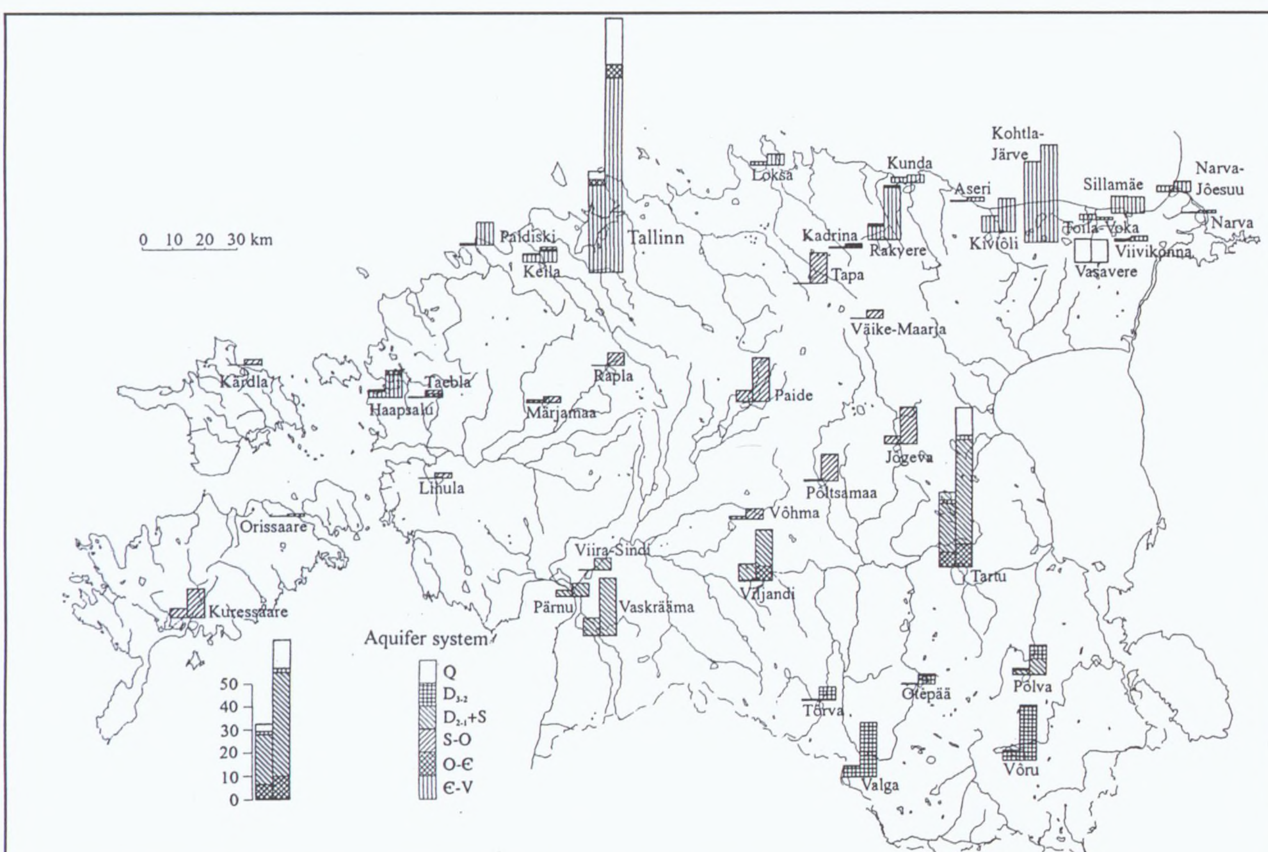


Fig. 109. Groundwater safe yield (right diagram) and actual pumping rate, thousands of m^3/d (left diagram). *Compiled by L. Savitski.*

1989). An analogous subsidence, only of smaller extent, occurred in Pärnu, where the compaction of Quaternary clayey sediments has been caused by the essential lowering of potentiometric surface in the underlying Silurian-Ordovician aquifer system due to intensive pumping.

Dewatering of mines and drainage of arable land

As a result of dewatering of oil-shale mines, the groundwater level of Quaternary and Ordovician aquifers has lowered by 15...65 metres and several local cones of depression, influencing each other, have formed over an area of 600 km^2 between Purtse and Narva rivers. In 1984-94, depending on precipitation, 600,000 to 900,000 m^3 of water was pumped out from the mines every day. The yearly amount of the water pumped out from the mines was 1.4...1.9 times as high as the total annual groundwater supply during the last decade. However, the proportion of groundwater in total mine water does not exceed 20...50%, the remainder is formed of surface water intruding directly into goafs through cracks, ventilation holes, unsealed wells, *etc.*, or precipitation which accumulates in open-pit quarries. During a wet period, 3 to 10 times as much water is pumped out from mines as during droughts (Газизов 1971, Норватов 1987, Савитский 1980). More than 60% of the whole amount of annual intrusion intrudes into the mines, located in the northern and central part of the oil shale region. It takes place during 4...5 months of water-abundant periods in spring and autumn. Drawdowns of the groundwater head increase southward in accordance with the dipping depth of the commercial oil shale layer. Depressions of the head are extending to a distance of 4...8 km outward from the mines.

The shallow wells, situated in the area of mine influence, often dry up. Dug into Quaternary cover or hewed into carbonate bedrock, they usually do not reach deeper than half a metre below the lowest groundwater level under natural conditions (Vallner 1996b). A groundwater drawdown of only 0.5 m caused by mine dewatering unfit these wells for water supply in dry period. In the areas, suffering from the greater drawdown, the dug wells are out of use during a longer period or they dry up completely. Besides of intensive pumping from the Vasavere intake, the mine dewatering is another reason why the groundwater table has lowered in the Kurtna Landscape Reserve (Kurtna... 1996). Owing to the depressurization of Ordovician carbonate aquifers in the area of the Jõhvi Upland, the recharge of the underlying Ordovician-Cambrian aquifer system has decreased causing the lowering of potentiometric surface by up to 8 metres (Vallner 1996b).

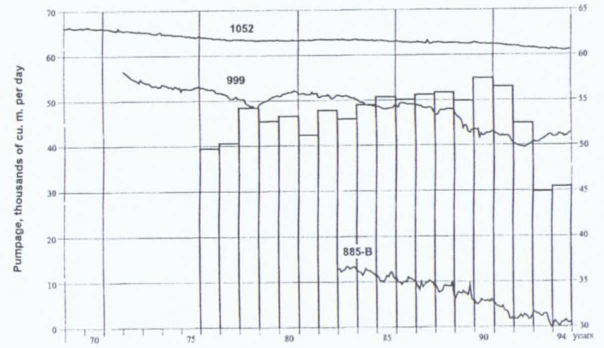


Fig. 110. Total pumping from the Ordovician-Cambrian aquifer system and change of the head in observation wells 885-B (Jõhvi), 999 (Väike-Maarja) and 1052 (Tooma). See also Fig. 98. *Compiled by L. Savitski.*

Since 1950, about 11, 000 km^2 of arable land and woodland have been drained by ditching and covered trenches. Altogether about 75% of arable agricultural land and about 30% of woodland has been drained. In the drained area, the groundwater level has dropped a metre, but the confined aquifers of the carbonate bedrock were also tapped in places which increased the local groundwater runoff. Therefore, an apprehension of the groundwater overdraft arose. However, the more detailed calculations showed that due to the lowering of the groundwater level, evaporation from its surface decreased inducing an increase of the total infiltration in the drained areas by 15 mm (Валлнер и Метсур 1988). Thus, as a result of the agricultural land drainage and mine dewatering, the groundwater recharge has not diminished, but, on the contrary, augmented.

Table 22. Groundwater safe yield and use in 1995

Characteristics of aquifer system	Aquifer system						
	Q	D_2	D_{2-1}+S	S-O	O-€	€-V	Total
Lying depth, m	20...60	60...18	120...140	20...180	50...420	120...250	
Number of wells	29	73	180	138	124	393	937
Pumpage from groundwater intakes with estimated safe yield, m^3/d	16, 041	11, 644	39, 858	16, 828	12, 168	105, 408	201, 947
Maximum draw-down (m) and its location	45 Tartu	75 Valga	75 Tartu	82 Jõgeva	94 Tartu	81 Kohtla-Järve	
Safe yield, m^3/d	42,025	64,608	103,780	97,165	28,900	215,890	552,368
Presumptive safe yield, m^3/d	40,000	281,500	218,000	646,900	77,200	270,000	1,533,800
Total water extraction, m^3/d	17,000	24,200	46,900	59,000	27,400	113,300	287,800
Share of aquifer system in total water extraction,%	6	9	16	20	10	39	100

VII TECTONICS

Deep structure

In the light of geophysical studies of the deep crustal structure conducted in the Fennoscandian (Baltic) Shield during the last decades and the much sparser data available on the adjacent Russian Platform area, the position of Estonia in the centre of the thickest crustal domain within the northwestern part of the East-European Craton becomes evident (Fig. 111). Main features of the regional crustal structure were formed during the Svecofennian orogeny (1.8 to 1.9 Ga). In the present morphostructure, the Baltic Sea Drainage Basin conforms well with the Svecofennian Crustal Domain (Puura & Flodén 1997). The Baltic Sea Depression (*sensu stricto*) occupies the central position in both the drainage basin and the crustal domain. In terms of the crust's age, Estonia locates within a unitary super- and morphostructure. The apparent differentiation in the crustal thickness (Moho depth) in Estonia and adjacent areas (Fig. 111) can be explained by both Svecofennian orogenic processes and subsequent crustal changes induced by tectonic and magmatic events.

Compared with Fennoscandia, less information is available on the deep structure of Estonia where the first data were obtained only a few decades ago. The data of the magnetotelluric survey (Fig. 112), carried out in 1970-72 (Андра и др.1974), gave 75 km for the mean depth of the **highly conducting region in the mantle**. At three stations, it was also possible to estimate the depth (140 to 180 km) of another highly conducting region.

Based on the studies, carried out on the FENNOLORA seismic refraction profile in 1979 and on the Sovetsk - Kohtla-Järve profile (Fig. 111) in 1983-86, Luosto (1991) demonstrated an almost east-west oriented **Moho depression** extending from the eastern coast of Sweden over the Baltic Proper to the Baltic States. On these profiles above the assumed trough of the depression, the crust is 55 to 64 km thick. Later studies showed that on the Baltic Sea profile between the islands of Gotland and Saaremaa the crust is thinner with its thickness ranging from 41 to 44 km (Ostrovsky *et al.* 1994). Hence, the Moho depression is located mostly on mainland (Fig. 111).

According to the calculations by Sadov and Penzina (Анкудинов и др. 1994), based on the Sovetsk - Kohtla-Järve profile, the crust is up to 64 km thick in the middle of the depression (Fig. 111) southeast of Riga, Latvia, and 46 to 51 m thick in the Estonian part of the profile.

In 1976, a seismic survey was carried out close to the epicentre of the Osmussaar earthquake of October 25, 1976. As a consequence, Bulin (Булин 1978) suggested the preliminary depth of about 42 to 47 km for the Moho depression in northwestern Estonia (Fig. 113, left). The reinterpretation (Булин и др. 1980) gave smaller depth values of 36 to 44 km for this area (Fig. 113, right) suggesting a local Moho uplift (Fig. 111). Considering the depth of the Moho surface in southern Finland (Luosto 1991), it seems that in Estonia the thickness of the crust increases slowly from 44 to 51 km towards the south.

In the Moho depth map (Fig. 111), the thickest crustal area in southeastern Estonia and its gradient zone are referred to the Svecofennian orogenic features, while the rised Moho surface in the West-Estonian Archipelago and in northern Estonia are contiguous to crustal thinning areas developed during the rapakivi-age continental rifting and basaltic underplating (Puura & Flodén 1996).

Like in the whole Russian Platform, the crust in Estonia is divided into two structural stages: (1) the strongly disturbed and metamorphosed Precambrian **basement**, and (2) an unconformably overlying thin (less than 800 m) **cover** of little deformed and gently tilted sedimentary strata. The crust is broken into blocks by deep-seated faults (Fig. 111).

Basement features

As shown above, Estonia belongs to the Svecofennian Crustal Domain, the structural patterns of which have been studied in particular detail in the Fennoscandian Shield. In Estonia, the data for crystalline basement studies has been basically obtained by geophysical (gravimetry, magnetometry, electrometry *etc.*) survey and deep drilling (Пуура и др. 1983, Koistinen 1994, 1996). As a whole, the main structural elements of the buried basement of Estonia are quite well established: 1) areas of folded metamorphic rocks, 2) plutonic rocks, 3) regional fault zones. The folded structure of the orogenic rocks can be characterised only in general lines, basing on observations of geophysical anomaly field patterns (Fig. 6) and drill core samples. Localities of anorogenic plutonic rocks with areas large enough for geophysical determination have been well mapped by deep drilling (Fig. 114). Fault zones of different age and different depth of formation (brittle or ductile deformation) have been studied by detailed geophysical profiling and veryfied by drilling.

Traditionally, the folded basement of Estonia has been divided into several structural zones (Fig. 114) differing in the rock composition and level of regional metamorphism (Пуура и др.1983). In the northern - northeastern part, the basement consists of metamorphic belts of amphibolite facies (Tallinn and Alutaguse structural zones) with local high-grade blocks (Jõhvi, Tapa, *etc.*). In the south and south-west, a large area of prevailing mafic to intermediate granulites with enderbites-charnockites has been distinguished, while mafic and intermediate rocks of metabolite facies occur in the west and northwest. The boundaries of the zones and blocks are usually marked by faults occurring as specific gradient zones in geophysical anomaly fields. Late-, post- and anorogenic plutonics accompanied by nonlinear anomalies show their cutting position in the linear anomaly fields of metamorphic rocks (Fig. 114).

According to the recent reinterpretation of the tectonic structure of the Fennoscandian Shield (BABEL... 1990, 1993, Korja 1993) and adjacent areas, considering the new data on the deep crustal processes in the region (Koistinen 1996, Puura & Flodén 1996), the Svecofennian structure of the basement is composed of a collage of metasedimentary basins squeezed between crustal blocks of the island arc origin (Korja 1993). The roots of the Svecofennian mountains have preserved along crustal bulges exceeding 50 km in thickness. In this interpretation, the Tallinn Structural Zone of the North-Estonian basement (Пуура и др. 1983) belongs to a prevailingly volcanic block of presumably volcanic arc origin. Its continuation might

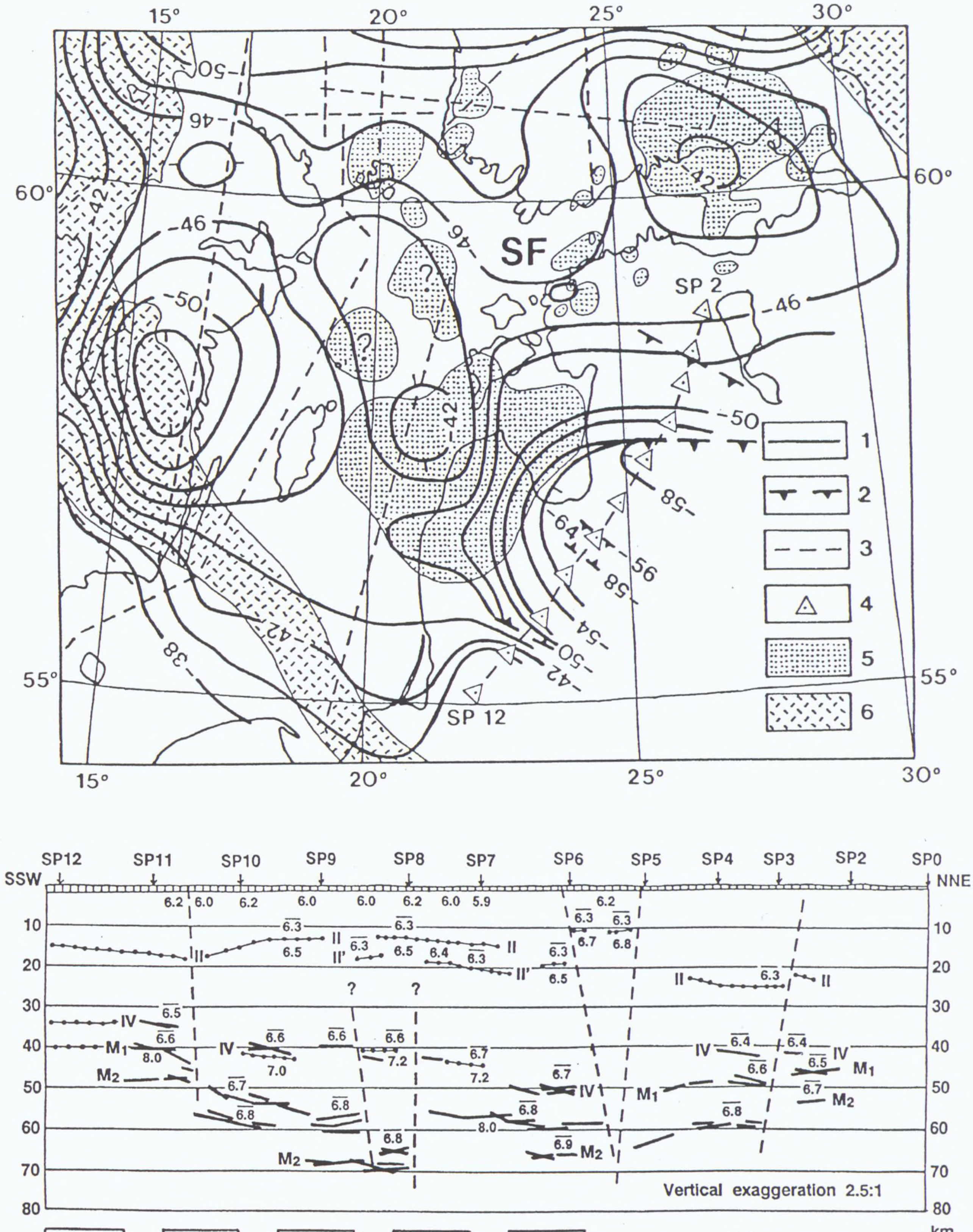


Fig. 111. Crustal structure of East Baltic; map (modified from Korja 1995) and section (from Анкудинов и др. 1994, simplified): 1 - contour on top of the Moho in kilometres; 2 - deep-seated fault; 3 - deep seismic sounding line; 4 - shotpoint on the Sovetsk - Kohtla-Järve profile; 5 - rapakivi granite and related rocks; 6 - boundary zone of the Svecofennian Domain (SF); 7 - refraction discontinuities, velocity of P waves in km/s immediately below the discontinuity; 8 - reflection discontinuities; 9 - sedimentary cover; 10 - shotpoint; 11 - average velocity of upper media. II - Conrad discontinuity, IV - discontinuity in lower crust, M - Mohorovičić discontinuity.

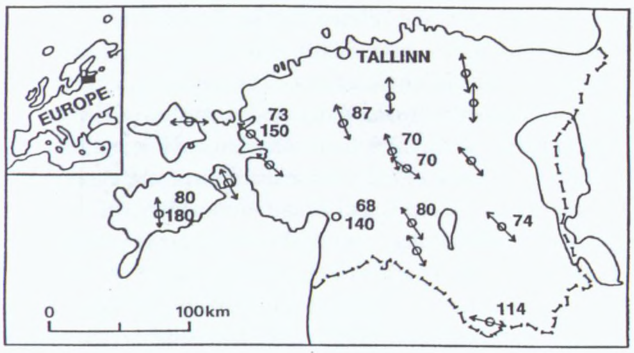


Fig. 112. Location of the magnetotelluric sites (from Андра и др. 1974) with the depth of the highly conducting regions in the mantle in kilometres. Arrows show the average polarization angles for the telluric field.

be observed in southern Finland (Koistinen 1994). The Alutaguse Zone in northeastern Estonia is probably a fragment of a large sedimentary basin presently exceeding the St. Petersburg and Novgorod areas in Russia.

Already in the early stages of investigation, it was stated that the structure of the basement in southern and southwestern Estonia differs from that in northern and northwestern Estonia (Фотиади 1958, Пуура и др.1983). The South-Estonian granulite area belongs to the Belarussian-Baltic assemblage of beltiform granulite and amphibolite facies tectonic sheets (Григялис и Пуура 1980, Пуура и др. 1984). Within the Belarussian-Baltic granulite subdomain, the origin and age of rock protoliths are probably the same as within the Svecofennian of Fennoscandia (Puura & Huhma 1993, Gorbatschev & Bogdanova 1993). The regional high-grade granulite metamorphism coupled with the regional high rock

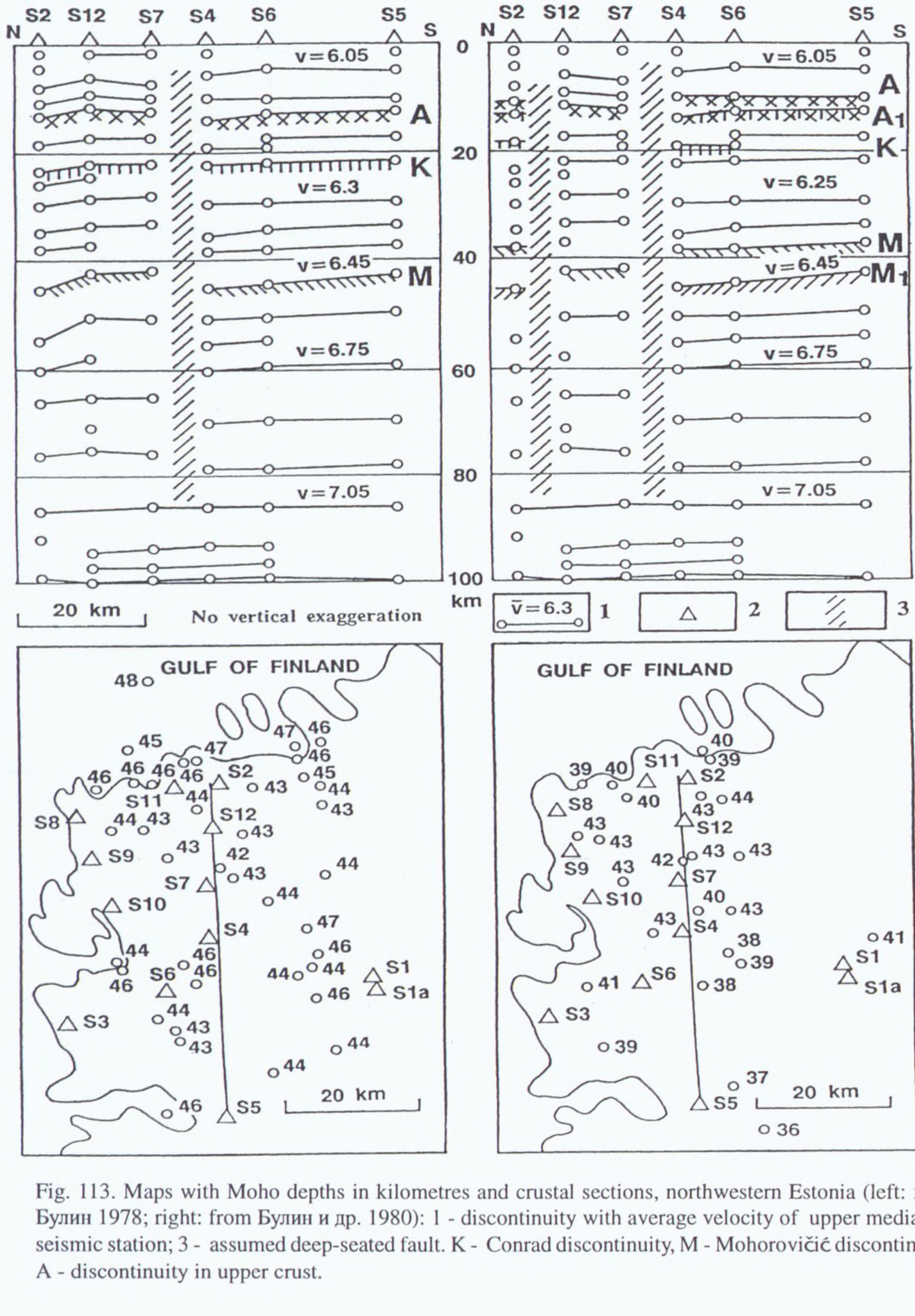


Fig. 113. Maps with Moho depths in kilometres and crustal sections, northwestern Estonia (left: from Булин 1978; right: from Булин и др. 1980): 1 - discontinuity with average velocity of upper media; 2 - seismic station; 3 - assumed deep-seated fault. K - Conrad discontinuity, M - Mohorovičić discontinuity, A - discontinuity in upper crust.

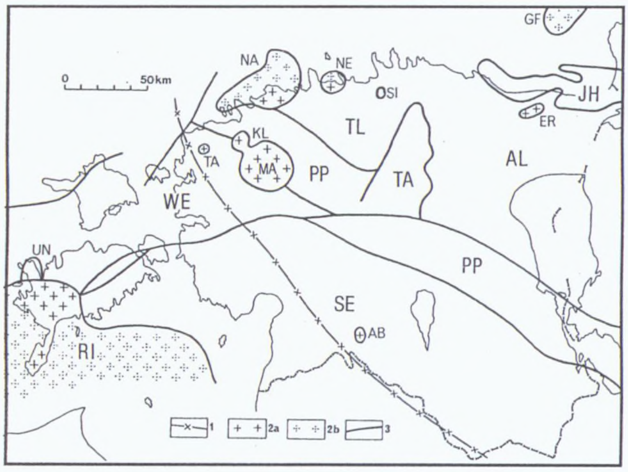


Fig. 114. Structural zoning of the Palaeoproterozoic orogenic basement and location of Palaeoproterozoic to Mesoproterozoic anorogenic igneous rocks: 1 - possible boundary of the Vyborg and Riga rapakivi subprovinces; 2 - plutons and volcanic sheets of rapakivi granite and related rocks in onshore (a) and offshore (b) areas; 3 - major faults. Zones and blocks: TL - Tallinn, AL - Alutaguse, TA - Tapa, JH - Jõhvi, PP - Paldiski-Pskov, SE - South Estonia, WE - West Estonia. Igneous structures of rapakivi and related rocks: GF - Gulf of Finland, ER - Ereda, NE - Neeme, NA - Naissaar, MÄ - Märjamaa, KL - Kloostri, SI - Sigula, TA - Taebla, AB - Abja, RI - Riga, UN - Undva.

density and magnetization in the area called the Baltic Geophysical High (Фотиади 1958, Puura & Huhma 1993) conceal the geological and potential field patterns characteristic for the proper Svecofennian in that area. Thus, the high-grade, prevailingly metavolcanic rocks in southern Estonia probably belong to the late Svecofennian stacked tectonic sheets.

Within the Svecofennian Domain, a number of orogenic fault systems can be distinguished. The undulating shear zones, shown on the map of the Precambrian basement (Koistinen 1994) in southern Finland, follow the direction of the Tampere subduction zone (Korja 1993) and are considered as being of early orogenic age. The curved but somewhat more straight fault zones, which coincide with the boundaries of the granulite belts of the Belarussian-Baltic Province, are probably of late orogenic origin (Григялис и Пуура 1980, Koistinen 1994). The general fault pattern of the Svecofennian Domain, including the Estonian basement, conforms with the deep crustal features, such as changes of Moho depth (Fig. 111) or the position of conductivity zones, or both (Korja 1993).

Like in the shield area, the structural zones of the basement, well observable in gravity and magnetic anomaly patterns, differ in assemblages of metasedimentary or metavolcanic rocks. The most clearly observable structural patterns of the orogenic crystalline rocks in drill cores are their prevailingly steep tilting of schistosity or, randomly, original bedding, and very variable stage of migmatization (or charnockitization). They occur everywhere in metasedimentary and metavolcanic rocks, but are of less intensity in early orogenic plutonic rocks. Their texture and composition are dealt with in more detail in Chapter III. The Paldiski-Pskov Zone occurs as a major marginal block assemblage of the Belarussian-Baltic granulite province (Gorbatschev & Bogdanova 1993).

The rapakivi and related magmatic structures formed 1.65-1.54 Ga, are about 200-300 Ma younger than the Svecofennian orogenic rocks. The rapakivi-anorthosite magmatism was related to and coupled with basalt underplating of the thick Svecofennian continental crust resulting in thinning of the crust in the area of the Vyborg Pluton (Elo & Korja 1993) and Åland Pluton (BABEL... 1993), but also in the northern part of the Baltic proper (Ostrovsky *et al.* 1994) and in northwestern Estonia (Fig. 114). Near the Åland Pluton, crustal signatures typical for continental rifts have been reported (BABEL... 1993). Probably, the crustal structure in the northern and western parts of Estonia (Fig. 111) carries traces of the above rifting, basaltic underplating and Subjotnian Rapakivi magmatism.

The Fennoscandian Rapakivi Province consists of four subprovinces (Koistinen 1994, Puura & Floden 1996). In all subprovinces of Subjotnian rapakivi, major polyphase bimodal (granite-anorthosite) volcano-plutonic complexes take a dominant position evident in deep seismic sounding and gravity field signatures (Elo & Korja 1993). The rapakivi-age diabase dike swarms and minor granite stocks occur around the main plutons designating the areas of different subprovinces. In Estonia, the fault-controlled Sigula offite gabbro body probably belongs to the Subjotnian diabase dike series.

The oldest, 1.62-1.67 Ga, Vyborg Subprovince has an eastern central position in the Province. The southern satellite body of the Vyborg Pluton in the central eastern part of the Gulf of Finland, and 7 rapakivi stocks in Estonia belong to the Vyborg Subprovince (Fig. 114, Table 4)). Individual plutonic bodies in Estonia measure some 300-1000 km^2 in diametre and have distinct specific geophysical signatures.

The 1.54-1.58 Ga Riga-Åland Subprovince in the central southwestern part of the province is represented in Estonia (Gulf of Riga, Ruhnu and western Saaremaa, central eastern part of the Baltic Proper) by the norheastern and northern wing of the Riga Pluton, largest in the province. Near the northern border of the Riga Pluton, the rapakivi-related local Undva volcanic sheet has been penetrated by a drill-hole.

During and after the rapakivi magmatism, faulting of the crust was of great importace. Positioning of plutons and minor stocks, as well as dike swarms indicates the prevailing northwest-southeast and northeast-southwest systems of faults. Major Svecofennian deep faults became reactivated as it was on the Åland - Märjamaa direction. However, the formation of new fault lines was a most frequent feature of rapakivi-time deformation. The Central-Estonian (Saaremaa - Mustvee) east-west striking fault zone probably belongs to this group (Пуура 1979). Tectonic block and fault movements were expressed on the earths surface as structural displacements of hundreds and, possibly, thousands of metres.

In areas adjacent to Estonia, in the Late Precambrian time span of 1.5-0.6 Ga at least five tectonic events, some of them coupled with mafic dike magmatism have been documented. There is no information on their occurrence in Estonia, as yet.

As a whole, a pattern of different-age fault systems, intersecting each other, and partly reactivated fault systems is a characteristic feature of the basement structure (Пуура 1979, Побул и Сильдвээ 1975). However, the majority of fault zones observed in geophysical fields are still difficult to date.

Much more stable geodynamic environments were established in the Svecofennian Domain after the Subjotnian rapakivi igneous activity. However, the results of the first proper planation occurred only in the beginning of the Jotnian sedimentation ca. 1.4 Ga, which became the first evidence of a stable craton. However, the final peneplanation of the region was succeeded just before the Late Vendian inundation and sedimentation.

The late pre-Cambrian break in sedimentation coinciding with the development of the pre-Vendian peneplain, resulted in the formation of the major angular disconformity in the geological sequence of Estonia which usually serves as a basis in structural studies of the sedimentary cover. Under the sedimentary cover, the upper part of the basement is weathered. In drill cores, clayey minerals as evidences of weathering have been determined in crystalline rocks at a depth 1-100 m below the basement surface.

Cover structure

The sedimentary bedrock of Estonia is divided into three **tectonic stages**, named mostly by geologists of the Baltic States (Сувейздис 1979): Baikalian (Vendian and Early Cambrian, Lontova Stage incl.), Caledonian (post-Lontova Cambrian, Ordovician, Silurian and Early Cambrian, Tilžė Stage incl.) and Hercynian (post-Tilžė Devonian). The tectonic stages are separated with regional unconformities. The solid bedrock is covered with a blanket of unconsolidated Quaternary deposits. The tectonic style of the sedimentary bedrock is to a certain degree laterally variable.

The most prominent structure of cratonic type is the **Estonian Homocline** (Пуура 1974) which extends from the Gulf of Finland to northern Latvia (Fig.115). The sedimentary bedrock strata have a very gentle (6 to 18') regional southward dip here (Пуура и Мардла 1972). Local variations in dip are well traceable in the meridional and, particularly, in the latitudinal section (Fig. 115). In southwestern Estonia, there is no distinct boundary between the Estonian Homocline and the Baltic Syneclise. Provisionally, it may be the contour line of -550 m on top of the basement.

In the southeast, the Estonian Homocline borders on the **Võru Saddle** (Вахер 1972) which is anticlinal in E-W section and synclinal in the perpendicular section. In the middle part of the saddle the basement lies at the level of -500 m, descending eastwards towards the Moscow Syneclise, and westwards towards the Baltic Syneclise. It rises northwards towards the Estonian Homocline, and southwards towards the Valmiera-Lokno Uplift. The buried Võru Saddle is clearly visible in the Baikalian and Caledonian rocks, but the unconformably overlying Hercynian strata have a fairly regular amount of dip in the general southern direction. This overlying structure is called the Estonian-Latvian Homocline (Сувейздис и др. 1977, Брангулис и др. 1984).

The **Valmiera-Lokno Uplift** in the Estonian-Latvian border zone is 20 to 30 km wide and 200 km long. It consists of four minor uplifts (Fig. 116): Valmiera, Smiltene (both in Latvia), Mõniste (in Estonia), and Haanja-Lokno (partly in Estonia, partly in Russia). On the crest of the highest, Mõniste Anticline, the basement lies at the level of -230 m, descending northward to -500 m, and southward to -1,000 m. There are deep-seated faults in the northern and southern flanks of the composite uplift (Fig. 115). The distribution of Vendian and Lower Palaeozoic rocks in southeastern Estonia shows that this structure was formed principally in the Late Silurian and has a long history (Вахер и др.1980a). In various parts of the major uplift, the basement is covered by sedimentary rocks of different age: Vendian on the Haanja-Lokno Uplift, Cambrian on the Valmiera Uplift, and even Devonian on the crest of the Mõniste Uplift. As the composite uplift began and developed in the Late Silurian and Early Devonian, from the crest at Mõniste over 200 m of Lower Paleozoic rocks were eroded. Figure 116 shows the lateral variety in the extent of the erosion at that time. A further uplift of the structure in the Middle Devonian was weaker. On the crest of the Haanja-Lokno Uplift (Fig. 117), the contact between the Middle and Upper Devonian lies at a level of 130 m, descending northward to 90 m and southward to -50 m. The anticline in the Upper Devonian strata is a result of post-Devonian uplift. The axis of the recent uplift, as established by repeated levelling data, is shifted to the north, and found straight above the northern slope of the old Mõniste Anticline (Fig. 118).

The Estonian Homocline, the largest structure in Estonia, is complicated by many minor structural features of various kind described in detail in northeastern Estonia (Вахер и др. 1962, Пуура 1986, 1987). The oldest of them are placanticlines - brachyanticlines or domes of sedimentary strata formed over pre-Vendian monadnocks of the basement. Most of the linear structures of various trend are probably of Late Silurian - Early Devonian age, but some of them possibly formed already in Vendian and Cambrian-Ordovician transition times or, otherwise, after Devonian sedimentation. The diapir-type near-surface structures in northeasternmost Estonia are probably of Pleistocene age. Meteorite impact structures may have any age. Small isometric depressions in the oil shale basin are undated yet.

Dome-like plain-type folds (**placanticlines**, Пуура и Кала 1978), some 1 to 6 km in diameter and 30 to 130 m in height on top of the basement (Table 23), have been found so far only in northern Estonia (Fig. 115). The net result of the folding is a relative local elevation without a corresponding depression. Thus, there are only brachyanticlines and domes rising above the regional dip. The folds become more pronounced with depth, the lowermost strata wedge out on the flanks of the folds, and there is a thinning of the strata above the crests of the folds (Table 24).

The best known, Uljaste Placanticline was discovered during the exploration for oil shale in 1930 (Reinwald 1935), and studied by boreholes finished in the basement in 1960-62 (Вахер и др. 1962, 1964) and 1975-77 (Пуура и Кала 1978). On the flanks of the structure (Fig. 119), not only all Vendian Strata but also the Lower Ordovician Pakerort and Varangu stages and the Mäeküla Member of the Billingen Stage wedge out (Вахер и др. 1964). Three layers of conglomerate, containing quartzite pebbles transported from the crest of the basement monadnock, were found in the Vendian (Fig. 119 - *47*) on the southern flank of the fold. In the claystone of the Dominopol' Stage the amount of glauconite and particles of silt and sand size increases toward the crest. Above the crest, the Lontova Stage is 12 to 18 m thinner, and the Lower Ordovician rocks show specific facial changes. In places, basal beds of the Billingen Stage contain rounded fragments of the underlying graptolite argillites (Fig. 119 - *49*) and pebbles of *Obolus* sandstones (Fig.119 - *45, 46, 49*), both of the Pakerort

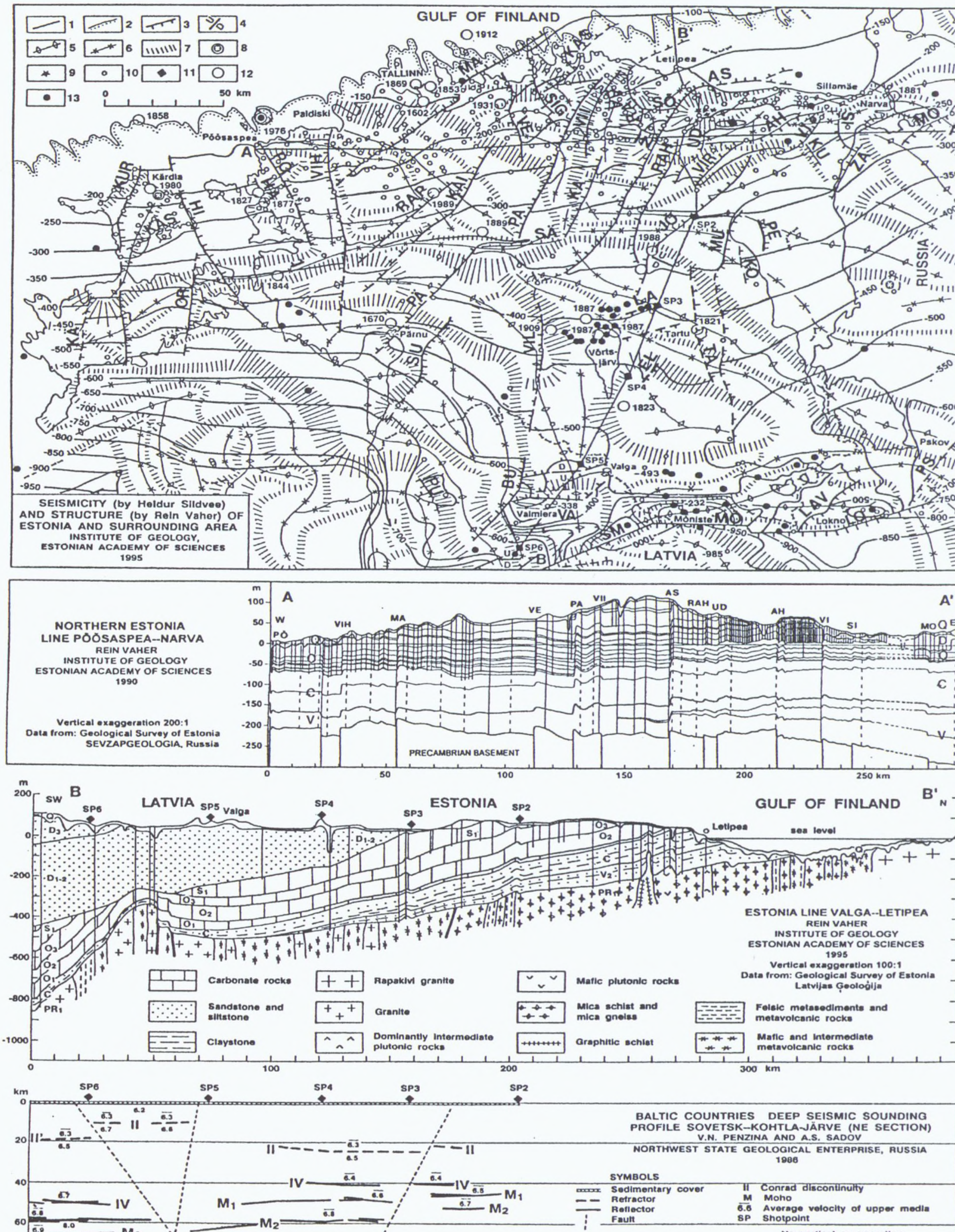


Fig. 115. Seismicity and structure of Estonia and surrounding area; map (After Sildvee & Vaher 1995, improved) and sections: 1 - contour on top of the Precambrian basement; 2 - updip limit of the sedimentary bedrock; 3 - flexure above a basement fault; 4 - deep-seated fault: U - upthrown side, D - downthrown side; 5-7 trend of Bouguer anomaly : high (5), low (6), gradient (7); 8 - crater; 9 - dome; 10 - borehole finished in the basement; 11 - shotpoint; 12, 13 - earthquakes: macroseismic (12), instrumental (13). Q - Quaternary; D - Devonian; O - Ordovician; Є - Cambrian; V - Vendian; PR - Palaeoproterozoic basement. Zones of disturbances: AA - Aaspere, AH - Ahtme, AS - Aseri, BU - Burtnieki, EL - Elva, HA - Haapsalu, HI - Hilleste, JÕ - Jõgeva, KA - Kahala, KAI - Kaiu, KAS - Kassari, KO - Kokora, KU - Kuremäe, KUR - Kurisu, KÄ - Kärla, LA - Laeva, LO - Lokno, MA - Maardu, MO - Monastyrek, MU - Mustvee, MÕ - Mõniste, OR - Orissaare, PA - Paide, PE - Peipsi, PS - Pskov, PU- Puikule, PÕ - Põõsaspea, PÄ - Pärnu, RA - Rakvere, RAH - Rahkla, RAP - Rapla, SIG - Sigula, SI - Sirgala, SM - Smiltene, SU - Surju, SÕ - Sõmeru, SÄ - Särevere, TA - Tartu, UD - Udriku, VA - Valmiera, VE - Vetla, VIH - Vihterpalu, VI - Viivikonna, VII - Viitna, VIL - Viljandi, VIR - Virunurme, ZA - Zagriv'e. For legend see also Fig. 111.

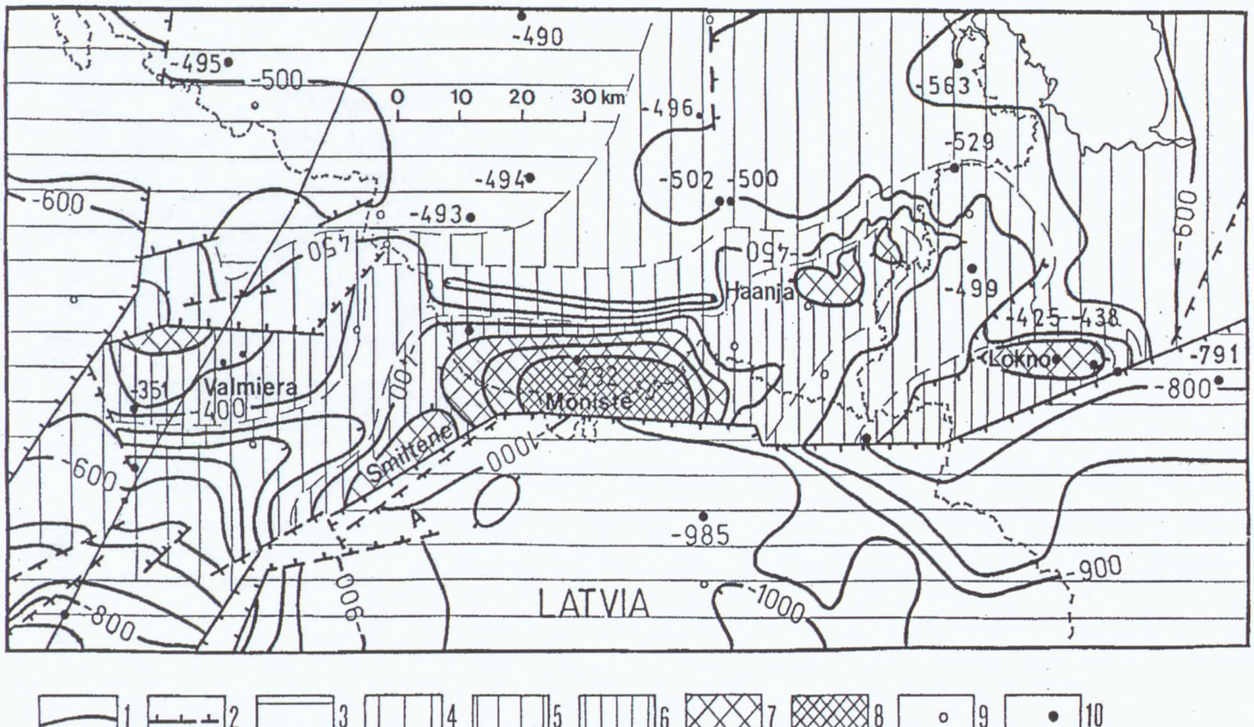


Fig. 116. Valmiera-Lokno Uplift (by Rein Vaher): 1 - contour on top of the Precambrian basement; 2 - flexure above a basement fault; 3-8 - pre-Devonian geology: Silurian (3), Upper Ordovician (4), Middle Ordovician (5), Lower Ordovician (6), Cambrian and/or Vendian (7), Palaeoproterozoic (8); 9 - borehole finished in Silurian, Ordovician, Cambrian or Vendian rocks; 10 - borehole finished in the basement.

Stage. Iron oolithes were found in an exceptional place: in the lowermost strata of the Saka Member of the Volkhov Stage.

The above structures remind monadnocks described in the seabed of the western central Baltic (Flodén 1984) and plain-type folds in the Mid-Continent Area (Clark 1932) of the North American Craton. Afanasyev and Volkolakov (Афанасьев и Волколаков 1981) classified the Uljaste Placanticline as a supratenuous fold generated by unequal compression of sediments over the buried basement hill. However, it must be admitted that the principle has certainly been reconsidered because the above facial changes in the Lower Ordovician cannot be explained by this mechanism. According to Vaher *et al.* (Вахер и др. 1980a), an uplift took place during deposition in the Vendian, the crest area being a sourceland from time to time. Mens *et al.* (1981) showed that the repeated events of uplift took place during regional interruptions of Vendian and Cambrian sedimentation accompanied by significant erosion. Movements were renewed in the Early Ordovician and after the Ordovician, possibly in the Early Devonian. Thus, the total folding of the covering rocks is due to both differential compaction and repeated uplift.

Narrow, some 1-to-4-km-wide linear **zones of disturbances** intersect both the sedimentary cover and the crystalline basement. These zones divide the homocline into a number of blocks with different sizes. In Estonia, serving as a transition zone between the Fennoscandian Shield and the Baltic Syneclise, the amplitude of block displacement does not exceed some 50 m (Пуура 1986). In the central part of the Baltic Syneclise (Latvia) a few similar blocks show displacements of the basement surface up to 600 m (Сувейздис 1979).

The zones of disturbances have been studied in detail in northeastern Estonia (Вахер и др. 1962, Пуура 1986, 1987) which is the best-known portion of Estonia because of extensive exploration for oil shale and phosphorite (more than 10,000 boreholes per 2900 km^2). Figure 120 shows a typical network of boreholes. As a rule, the zones of disturbances

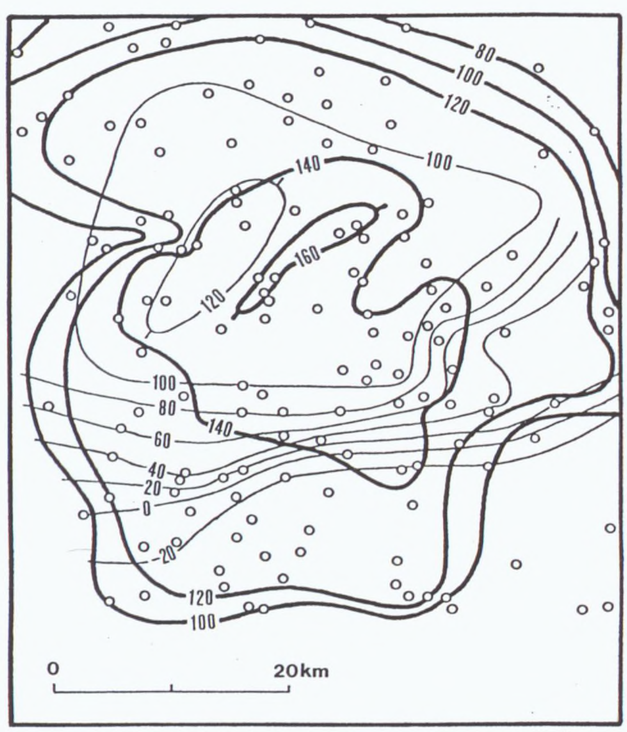


Fig. 117. Structure (thin lines) and topography (thick lines) of the bedrock of the Haanja Heights, southeastern Estonia. (Raukas *et al.* 1988). Contour interval 20 m. Small circles denote boreholes.

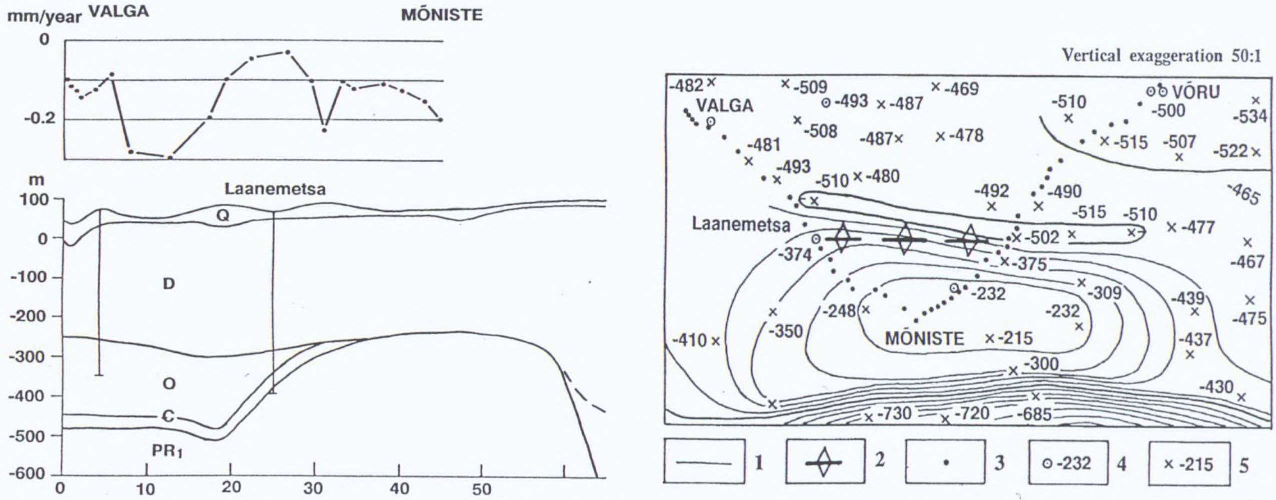


Fig. 118. Mõniste Uplift, southeastern Estonia; map and section (Sildvee & Vaher, 1995): 1 - contour on top of the Precambrian basement; 2 - axis of the recent uplift; 3 - repeated levelling set-up; 4,5 - height of the basement in metres above mean sea-level: in borehole (4), at depth point of refraction survey (5). Q - Quaternary, D - Devonian, O - Ordovician, Є - Cambrian, PR_1 - Palaeoproterozoic crystalline basement.

Table 23. Data on placanticlines

	Placanticline				
	Assamalla	Nüri	Sonda	Uljaste	Vana-Sonda
Lateral dimensions, km	2x5	1.4x3.1	1.0x1.7	1.2x2.8	0.7x1.3
Height (m) on bottom of the:					
Kukruse Stage	24	11	11	23	6
Toila Formation	24	11	12	23	7
Lükati Formation	23	11	11	27	6
Lontova Formation	40	11	17	40	6
Height (m) on top of the Precambrian basement	110	67	70	130	38

Table 24. Thickness of the Cambrian and Vendian rocks on a profile through the Uljaste Placanticline

	Borehole									
Formation	231	230	47	45	48	49	188	229	80	228
Tiskre	19	19	10?	16	16	12	18	21	8?	17
Lükati	13	11	24?	14	10?	16	11	12	20?	11
Lontova	67	65	56	59	64?	59	64	66	72	70
Voronka	24	20	4	0	0	2	5	20	20	23
Kotlin	17	8	4	0	0	0	2	16	14	16
Gdov	46	9	8	0	0	0	3	21	40	36

occur as a flexure above a basement fault combined with anticline in the upthrown side and with syncline in the downthrown side (Вахер и др. 1978). Total vertical amplitude (A_t) in the sedimentary bedrock consists of components, known as anticlinal (A_a), monoclinal (A_m), and synclinal (A_s), as shown in Figure 121. There are fracture zone(s) on the more steeply dipping limb and, in places, fault(s) in the fracture zone. Main parametres of the zones of disturbances are given in Table 25.

The folds in the zones of disturbances are in general very gentle and not high, usually 5 to 10 m and never more than 50 m. Anticlinal belts are ascribed mostly to fault movements in the basement because they are very asymmetric in shape. The Viivikonna Anticline (nose in Fig. 120) is an excellent example of that kind of structure. The beds in its southeastern limb have a very gentle average dip of 20' (max. 55') towards southeast, in the northwestern limb up to 7° towards southwest (Fig. 121A). In the 125-m-wide fracture zone the intensity of jointing increases toward the centre, and some oil shale beds are substituted by karst clay. In the 40-to-50-m-wide part of the structure, known as the shatter zone, the carbonate rocks are so intensively cracked and mixed with karst clay that it is difficult to tell whether there are faults in the zone or not. The uplifts are often called monoclines, because the opposite flank is nothing but a flexure (Fig. 121B).

Alterations of deformed rocks within the zones of disturbances show a long and multistep development of the zones, and fluid migration in the upper crust (Вахер и др. 1962, Пичугин и др. 1976), including metasomatic dolomitization of limestones, sulphide lead and zinc mineralisation, karst, *etc.* (Fig. 122).

The Aseri Zone was studied in 1975-77 at Viru-Nigula (Fig. 123) by a profile of six boreholes finished in the basement (Вахер 1983, Пуура 1987). If the thinning of some Upper

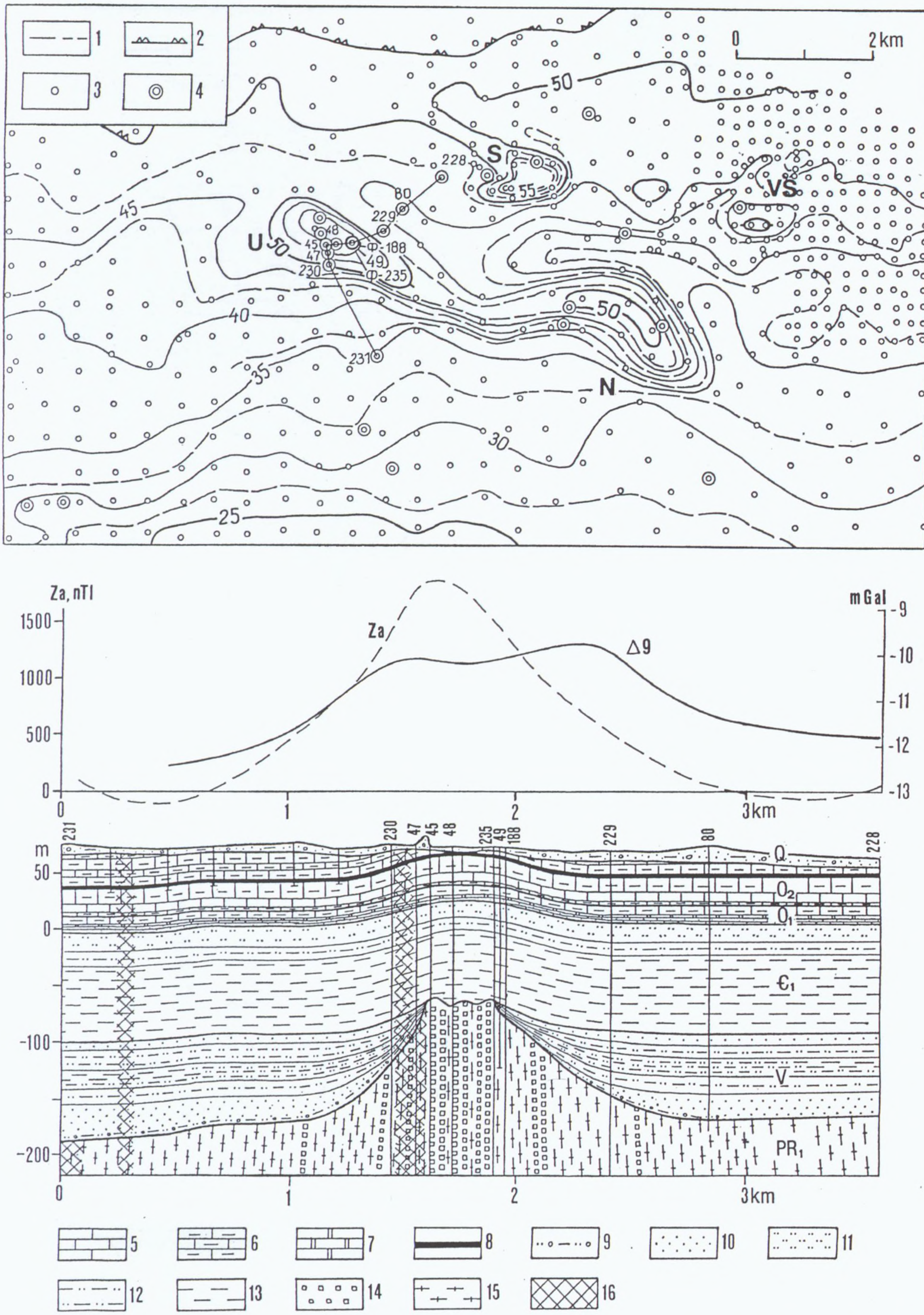


Fig. 119. Placanticlines of the Sonda-Uljaste area, northeastern Estonia; map (from Пуура 1987) and section (from Вахер 1983): 1 - contour on bottom of the oil-shale seam; 2 - updip limit of the oil-shale seam; 3 - borehole finished in the sedimentary bedrock; 4 - borehole finished in the basement; 5 - limestone; 6 - argillaceous limestone; 7 - dolomite; 8 - oil-shale seam; 9 - mixtite; 10 - sandstone; 11 - siltstone; 12 - alternation of claystone, siltstone and sandstone; 13 - claystone; 14 - quartzite; 15 - gneiss; 16 - shatter zone. Placanticlines: N - Nüri, S - Sonda, U - Uljaste, VS - Vana-Sonda.

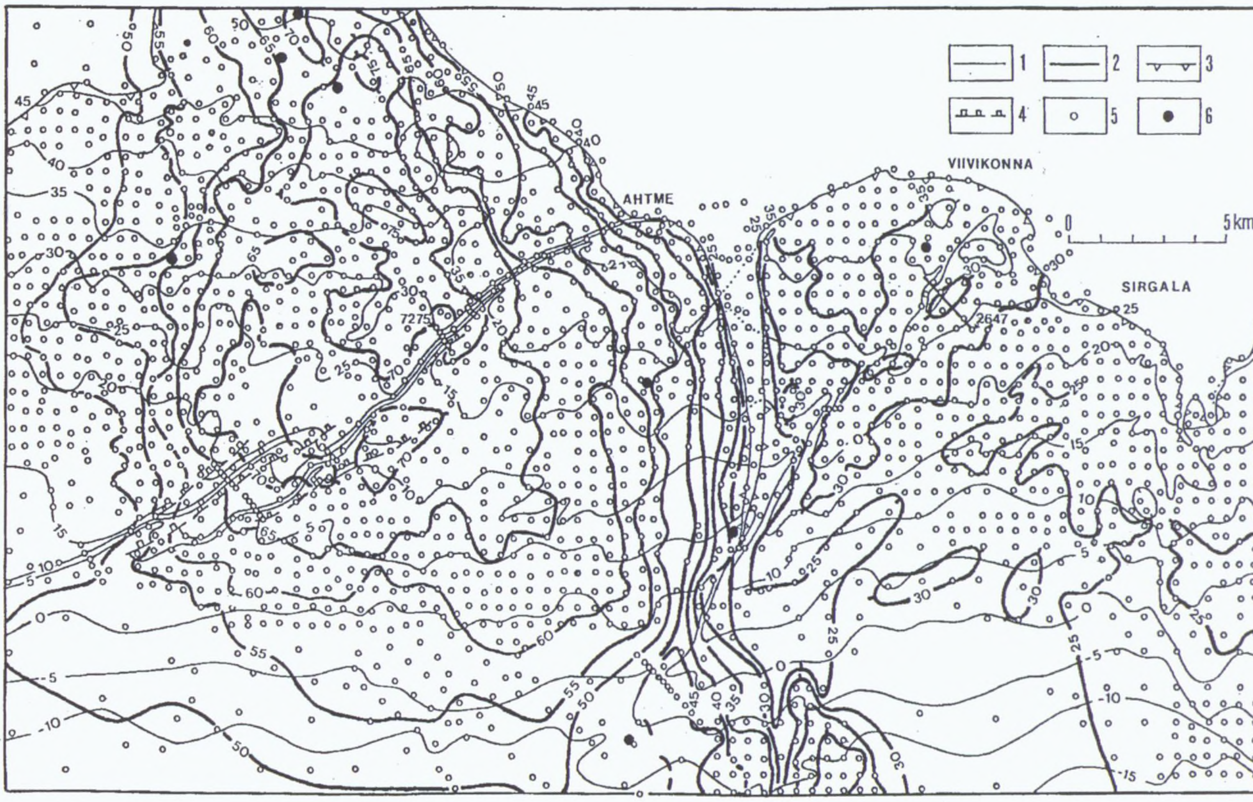


Fig. 120. Structure and topography of the sedimentary bedrock of the oil-shale mining area, northeastern Estonia. (After Вахер 1983, Raukas *et al.* 1988, Puura *et al.* 1987, Туулинг 1988а,б): 1 - contour on bottom of the oil-shale seam; 2 - contour on top of the bedrock; 3 - updip limit of the oil-shale seam; 4 fault; defined (solid), assumed (dashed); 5 - borehole finished in the sedimentary bedrock; 6 - borehole finished in the basement.

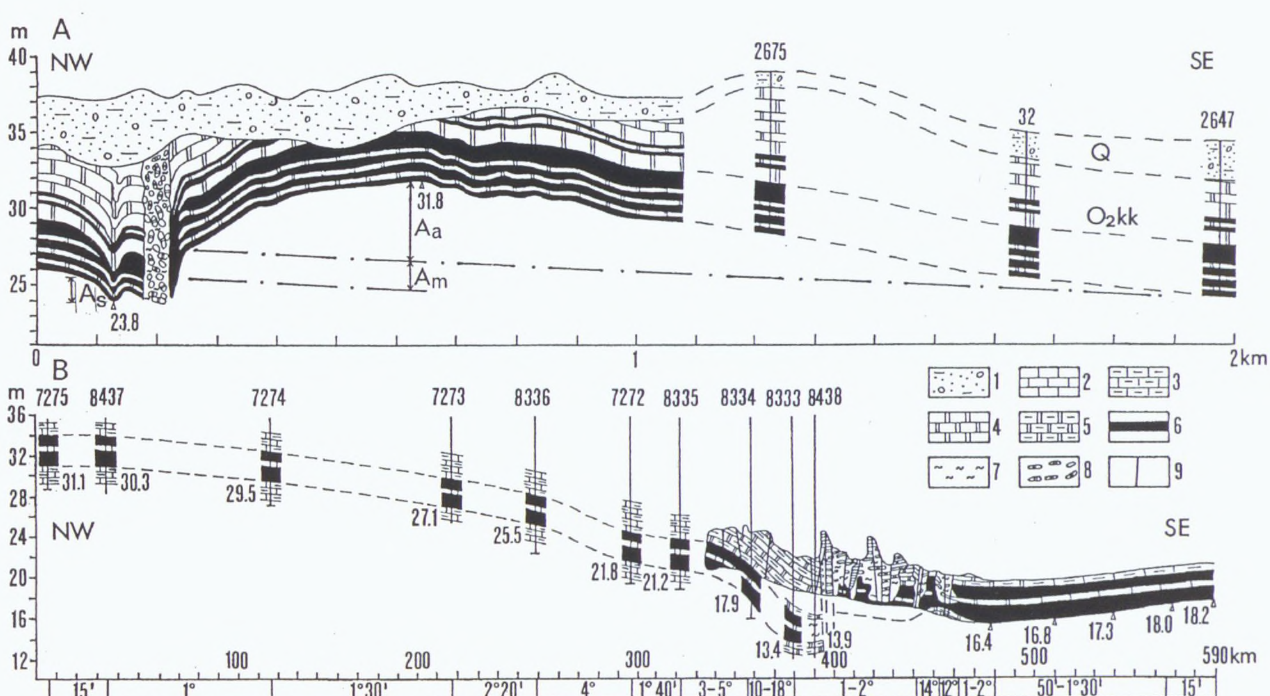


Fig. 121. Zones of disturbances: (A) Viivikonna (modified from Вахер 1983), (B) Ahtme (from Каттай и Вингисаар 1980): 1 - Quaternary deposits (Q); 2 - limestone; 3 - argillaceous limestone; 4 - dolomite; 5 - argillaceous dolomite; 6 - oil shale (kukersite); 7 - karstic clay; 8 - fragments of carbonate rocks; 9 - fracture. O_{2kk} - Kukruse Stage. Components of the vertical displacement: A_a - anticlinal, A_m - monoclinal, A_s - synclinal.

Table 25. Data on zones of disturbances

Zone	Marker	Length km	Width km	Vertical displacement, m Aa	Am	As	At
Aaspere (AA)	Kukruse Stage	18	?	9	?	?	13
Ahtme (AH)	Kukruse Stage	50	0.5-2	0-9	8-16	1-3	21
Aseri (AS)	Kukruse Stage	>150	1-4	0-7	15-19	1-3	26
	Basement surface			10	22	18	50
Elva (EL)	Basement surface	25	4	?	?	?	55
Haapsalu (HA)	Volkhov Stage	>20	?	?	?	?	19
Hilleste (HI)	Jaani Stage	>60	?	?	?	?	8
Jõgeva (JÕ)	Juuru Stage	>60	?	?	?	?	14
Kahala (KA)	Juuru Stage	45	2-4	8-9	3-6	3-4	15
Kaiu (KAI)	Juuru Stage	>50	3-4	<10	<10	1-5	20
Kassari (KAS)	Juuru Stage	25	<5	?	?	?	11
Kokora (KO)	Narva Stage	>20	?	?	?	?	20
Kuremäe (KU)	Kukruse Stage	11	2-3	?	?	?	8
Kurisu (KUR)	Juuru Stage	>100	?	?	?	?	10
Kärla (KÄ)	Rootsiküla Stage	>35	?	?	?	?	13
Laeva (LA)	Narva Stage	>20	<5	?	?	?	15
	Basement surface						26
Lokno (LO)	Pļaviņas Stage	30	<10	28	?	?	>75
	Basement surface			>100	?	?	>400
Maardu (MA)	Volkhov Stage	>120	2-6	4-15	4-5	1-8	22
Monastyrek (MO)	Volkhov Stage	22	1-2	?	?	?	11
Mustvee (MU)	Narva Stage	>30	?	?	?	?	20
Mõniste (MÕ)	Basement surface	50	15	260	?	?	>750
Orissaare (OR)	Jaani Stage	>30	?	?	?	?	15
Paide (PA)	Kukruse Stage	>90	3-4	10-17	6-16	3-5	28
Peipsi (PE)	Narva Stage	15	2-4	<5	15	<5	20
Põõsaspea (PÕ)	Volkhov Stage	>50	2-3	?	?	5	27
Pärnu (PÄ)	Juuru Stage	>60	?	?	?	?	31
Rakvere (RA)	Kukruse Stage	>50	2-4	12-15	2-4	2-5	23
Rahkla (RAH)	Kukruse Stage	65	1-2	?	?	?	12
Rapla (RAP)	Juuru Stage	>35	?	?	?	?	16
Sigula (SIG)	Kukruse Stage	>20	?	?	?	?	5
Sirgala (SI)	Kukruse Stage	23	1-3	1-6	2-6	1-2	10
Smiltene (SM)	Basement surface	35	4-7	80	?	?	>600
Surju (SU)	Narva Stage	>40	2-3	?	?	?	12
Sõmeru (SÕ)	Kukruse Stage	13	0.5-1	0-1	8-11	?	10
Särevere (SÄ)	Juuru Stage	40	6-9	?	?	?	50
Tartu (TA)	Narva Stage	>50	?	?	?	?	10
Udriku (UD)	Kukruse Stage	>15	?	?	?	?	11
Valmiera (VA)	Basement surface	35	12	130	?	?	>200
Vetla (VE)	Kukruse Stage	>40	3-4	12	5-10	?	22
Vihterpalu (VIH)	Rakvere Stage	>100	2-8	0-10	8-30	0-4	49
Viivikonna (VI)	Kukruse Stage	30	2-5	3-6	2-3	1-3	11
Viitna (VII)	Kukruse Stage	>40	2-4	5-12	3-5	2-4	17
Viljandi (VIL)	Narva Stage	>70	?	?	?	?	9
Virunurme (VIR)	Kukruse Stage	>20	?	?	?	?	11
Zagriv'e (ZA)	Kukruse Stage	>20	1-2	1-4	8-12	2-3	16

Components of the total (A_t) vertical displacement: A_a - anticlinal, A_m - monoclinal, A_s - synclinal.

Vendian (Uusküla and Voronka) and Lower Cambrian (Kestla+Mahu) members above the crest of the anticline (Table 26) is due to vertical movements during deposition, the uplift started in the Late Vendian, ceased for a time, renewed at the end of the Vendian, and ceased again for a time. A further uplift took place in the Early Cambrian (Lontova Age). It was followed by a long-term stable period which, as found out in southern sections of the structure, runs at least to the Wenlock (Early Silurian). The age of the folds in the Ordovician and Silurian strata is probably Early Devonian. Along some of the zones of disturbances one can recognize contrasting movements in succeeding periods. Thus, it appears that the Ahtme Monocline in the Ordovician strata is due to a rise of the northwestern block, while a minor fault in the Ahtme Zone is indicative of a later rise of the southeastern block (Fig. 121B).

Probably, specific disturbances (without roots in the basement) which occur in the sedimentary cover of the northeasternmost part of Estonia are of glaciotectonic nature. According to Jaansoon-Orviku (1926), the markedly tilted (20 to 70° SSE or SSW) Lower Ordovician strata in exposures on the slopes of three hills (Pargimägi, Põrguhauamägi and Tornimägi) at Vaivara belong to erratics. Miidel *et al.* (Мийдел и др. 1969) have described folded rocks (Fig. 124a) on the western slope of the Tornimägi Hill (Photo 34) and

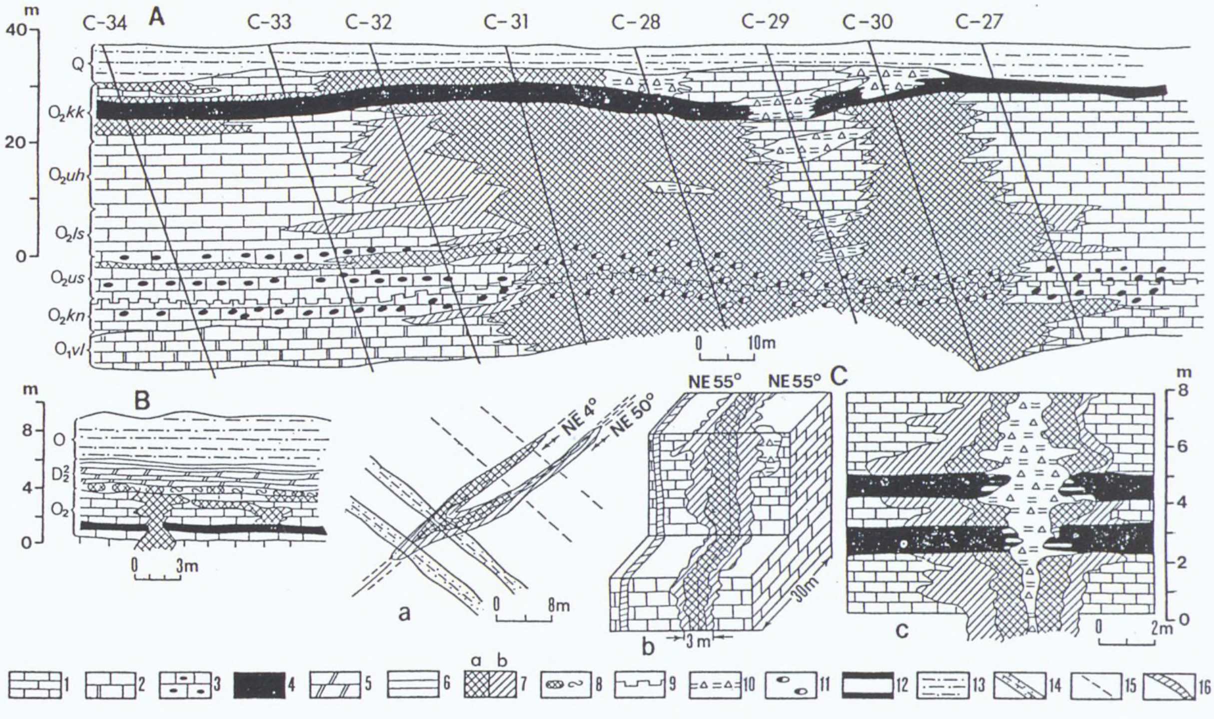


Fig. 122. Fracture zones (from Пичугин и др. 1976): profile in the sector of Sirgala quarry (A); outcrops in Sirgala (B), Väo (C,a) and Aru quarries (C,b), and in Shaft No 2 of the Slantsy deposit (C,c): 1 - limestone; 2 - dolomitic limestone; 3 - limestone with ferruginous oolites; 4 - oil-shale seam; 5 - dolomitic marl; 6 - claystone; 7 - sectors of development of metasomatic dolomitization: intensive (a), weak (b); 8 - eluvium of metasomatic dolomite; 9 - erosion surfaces during sedimentation; 10 - karst fillings (clay and fragments of carbonate rocks); 11- decomposed ferruginous oolites; 12 - oxidized kukersite; 13 - Quaternary deposits; 14 - zones of intense ferruginization; 15 - tectonic joint; 16 - calcite vein.

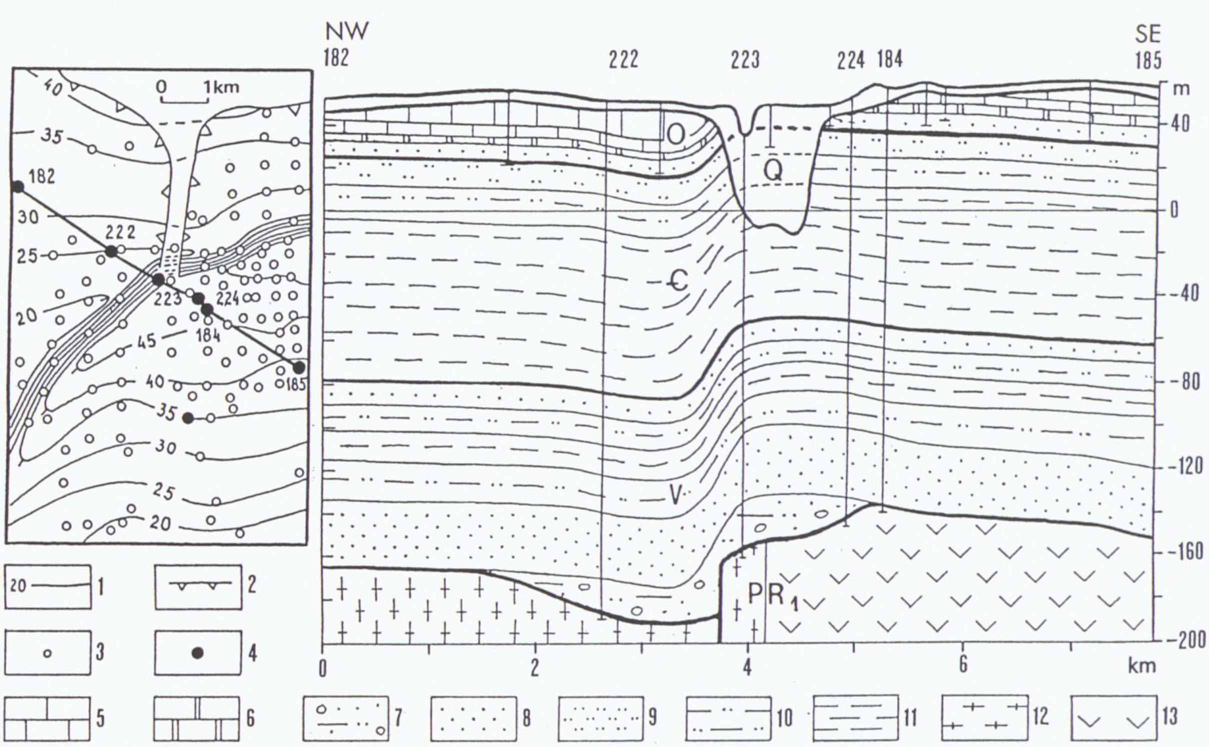


Fig. 123. Aseri Zone of disturbances, northeastern Estonia; map and section (modified from Baxep 1983): 1 - contour on bottom of Lower Ordovician carbonate rocks; 2 - updip limit of Lower Ordovician carbonate rocks; 3 - borehole finished in the sedimentary bedrock; 4 - borehole finished in the basement; 5 - limestone; 6 - dolomite; 7 - mixtite; 8 - sandstone; 9 - siltstone; 10 - alternation of claystone, siltstone and sandstone; 11 - claystone; 12 - gneiss; 13 - gabbro. Q - Quaternary, O - Ordovician, Є - Cambrian, V - Vendian, PR_1 - Palaeoproterozoic basement.

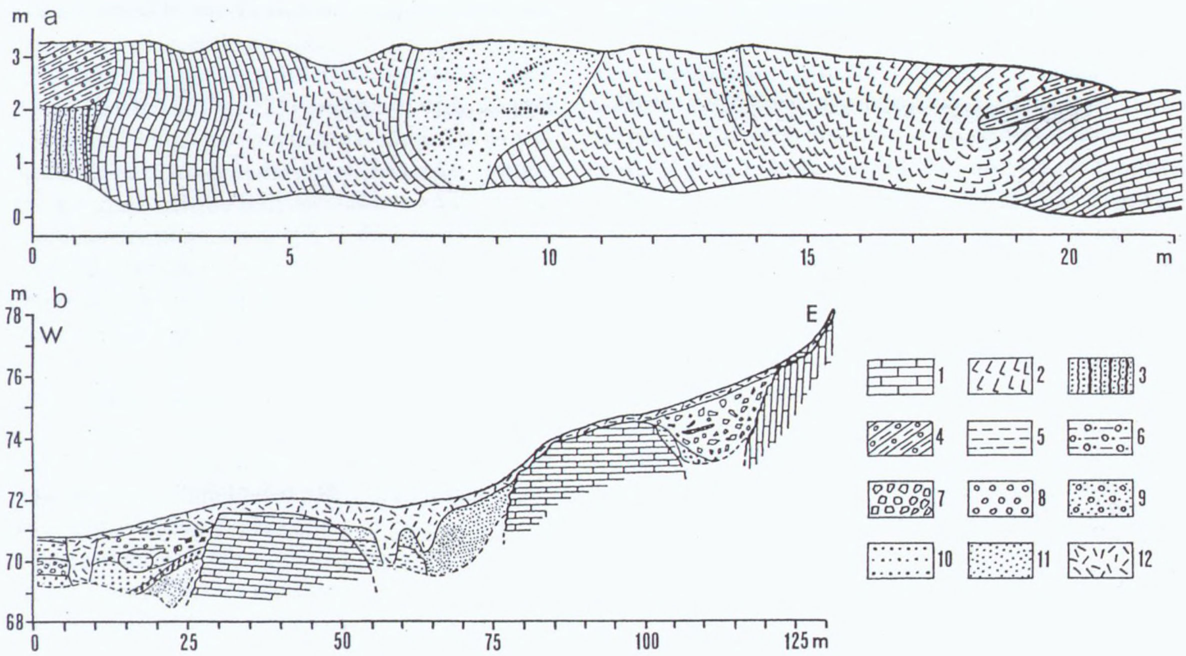


Fig. 124. Glaciotectonic structures at Vaivara, northeastern Estonia (from Мийдел и др. 1969): Tornimägi Hill (a) and Pargimägi Hill (b): 1 - limestone; 2 - shattered limestone; 3 - sandstone with argillite beds; 4 - till; 5 - green clayey till; 6 - brown sandy till; 7 - grey carbonaceous till; 8 - pebbles; 9 - glaciofluvial gravel and pebbles; 10 - sandy gravel; 11 - cross-bedded sand; 12 - soil.

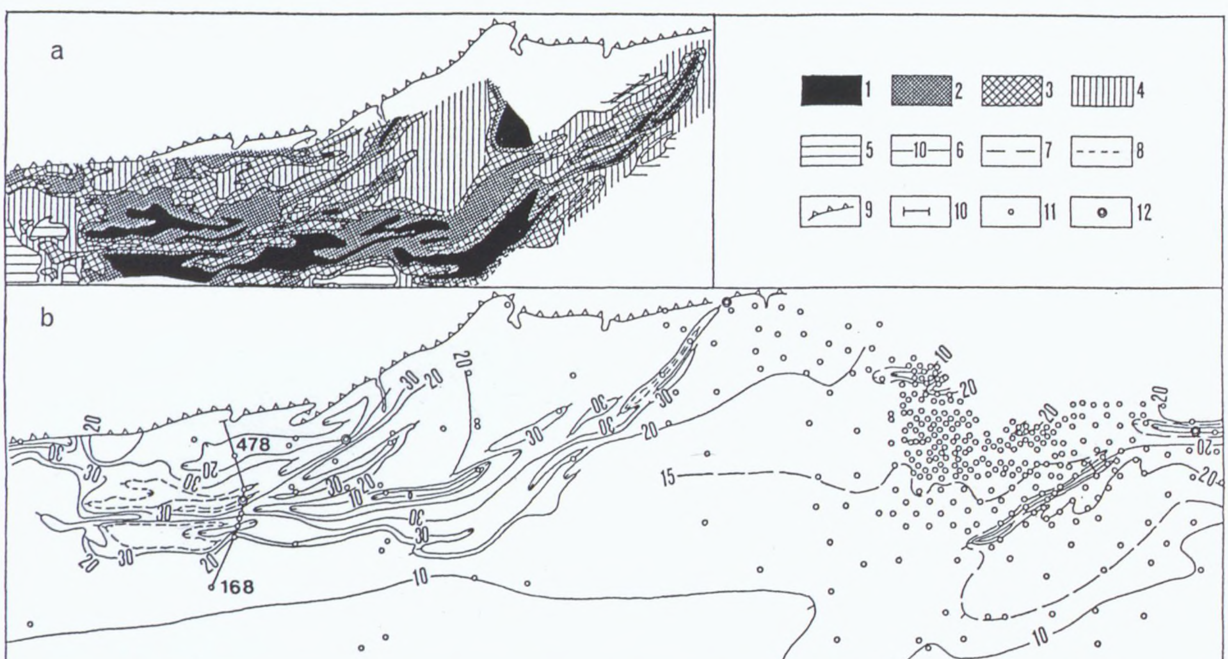


Fig. 125. Clay diapirs, northeastern Estonia; resistivity (a) and structure (b) maps (from Вахер и Мардла 1969): 1-5 – resistivity in ohm m: <50 (1), 50-100 (2), 100-200 (3), 200-400 (4), >400 (5); 6-8 – contour on top of the Pakerort Stage; 9 - klint; 10 - borehole finished in the sedimentary bedrock; 11 - borehole finished in the basement.

Table 26. Thickness of the Cambrian and Vendian rocks on a profile through the Aseri Zone of disturbances

Formation (Fm), Member (Mb)	Borehole 182	222	223	224	184	185
Tiskre Fm	14	13	0	13	12	12
Lükati Fm	13	14	0	12	12	13
Lontova Fm	75	76	50+	67	67	68
Kannuka Mb	12	10	8	8	8	7
Sirgala Mb	14	14	14	13	4	12
Meriküla Mb	10	12	12	12	12	13
Jaama Mb	6	6	6	6	5	6
Uusküla Mb	14	12	8	9	9	17
Moldova Mb	32	31	32	33	32	31
Oru Mb	0	15	23	11	1	0

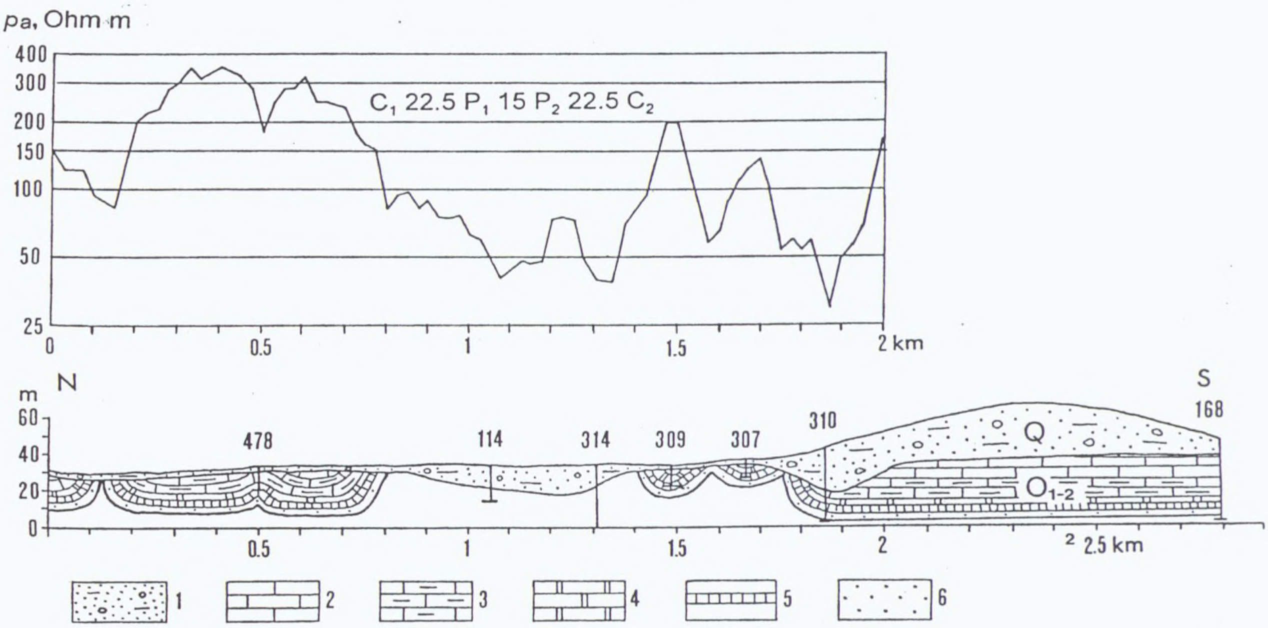


Fig. 126. Clay diapirs, northeastern Estonia; resistivity curve and cross section (from Вахер и Мардла 1969): 1 - Quaternary deposits (Q); 2 - limestone; 3 - argillaceous limestone; 4 - dolomite; 5 - argillite; 6 - sandstone. O_{1-2} - Lower and Middle Ordovician. ρ_a - apparent resistivity. Schlumberger array: C_1, C_2 - current electrodes; P_1, P_2 - potential electrodes; distance between electrodes in metres.

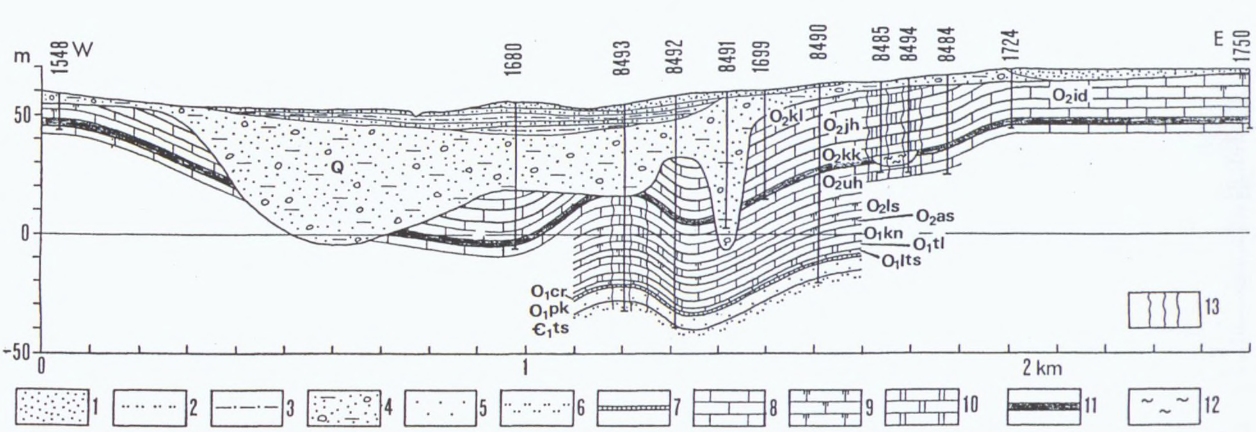


Fig. 127. Sämi Basin, northeastern Estonia (from Вахер 1983): 1 - sand; 2 - sandy loam; 3 - loam; 4 - till; 5 - sandstone; 6 - siltstone; 7- argillite; 8 - limestone; 9 - dolomitic limestone; 10 - dolomite; 11 - oil-shale seam; 12 - karstic clay; 13 - fracture zone. O_{2kl} - Keila Stage, O_{2jh} - Jõhvi Substage, O_{2id} - Idavere Substage, O_{2kk} - Kukruse Stage, O_{2uh} - Uhaku Stage, O_{2ls} - Lasnamägi Stage, O_{2as} - Aseri Stage, O_{1kn} - Kunda Stage, O_{1tl} - Toila Formation, O_{1lt} - Leetse Formation, O_{1pk} - Pakerort Stage, $Є_{1ts}$ - Tiskre Formation.

erratics (Fig. 124b) in the section of the Pargimägi Hill, the hills themselves being end moraines. In a northerly and norheasterly direction, between Sillamäe and Narva, some narrow linear structures (Fig. 125) are due to **clay diapirs** (Вахер и Мардла 1969). One anticline was studied in an artificially produced trench where the dip of the Lower Ordovician carbonate strata little by little increases toward the centre of the structure until reaching 40°. Carbonate rocks were eroded from the central part of the anticline, and the exposed Lower Cambrian (Dominopol' Stage) siltstone beds are virtually vertical in the middle of the structure. A similar structure was demonstrated by Orviku (1930b) at Narva - Kalmistu where the dip of the Lower Ordovician carbonate strata and the Lower Cambrian siltstone beds is 20 and 55^0, respectively.

In the drill core of borehole No. 314 (Fig. 126) at Sinimäe, in the centre of another anticline, the Vendian strata are practically undisturbed above the basement of normal altitude. The claystones of the overlying Lontova Stage are disturbed by numerous slickensides ever increasing upward in number. Clay- and siltstones of the Dominopol' Stage are severely disturbed, and nearly three times as thick as normal. Evidently, the claystone was locally squeezed upward along the lines of minimum resistance.

In places, the general direction of the clay diapirs coincides with the usual trend of the above linear zones of tectonic disturbances in northeastern Estonia. Thus, the diapirs might inherit the position of pre-existing disturbances.

Small depressions of unknown origin, 0.1 to 1km in diameter, have been found in oil-shale mines. The largest (1.3 to 2 km), Sämi (Kabala) depression is well-known (Вахер и др. 1962) owing to exploration for oil shale in 1950-52. This is an oval structure with a slightly wavy eastern limb (Fig. 127), in the centre of which the Ordovician strata lie 50 m below the level of surroundings. The amplitude on top of the basement is not known.

On the Island of Hiiumaa, a large impact crater is buried under Upper Ordovician carbonate rocks (see Ch. XI, 2). It is nearly 4 km in diameter and 540 m deep (Puura & Suuroja 1992). The diameter of the largest, Neugrund crater is 7 km.

Photo 34. Glaciotectonic structures on the western slope of the Tornimägi Hill, Vaivara End Moraine. *Photo by A. Miidel.*

Neotectonics and recent crustal movements

Under neotectonic movements the authors mean and deal with the Neogene-Quaternary crustal movements. Since deposits from the Upper Devonian through Neogene are absent in Estonia, the reconstructions of the neotectonic movements are to a great extent based on the analysis of the bedrock topography. The latter has been regarded as consisting of polygenetic planation surfaces of different age. It has been suggested that either Miocene-Pliocene (Можаев 1973, Šliaupa *et al.*1982), Miocene (Исаченков 1982, 1983) or Mesozoic and Palaeogene (Мещеряков 1965) planation surfaces are found in the bedrock topography in Estonia. According to Mozhayev (Можаев 1973), the total Neogene-Quaternary uplift reaches 200 m.

The formation of the bedrock topography started at the end of the Devonian, and is polygenetic in character. Besides long-term crustal movements and various continental denudational processes, its development was controlled by glacial erosion in the Pleistocene. Isachenkov (Исаченков 1982) claimed that in the northern part of the Baltic region, glacial erosion reached the amount of 80 m, which often exceeds the relative height of the bedrock surface features treated as planation surfaces by other authors. Taking into consideration remnants of erosional topography of Middle Devonian age, established in western Russia (Саммет 1961, Туулинг 1988б, 1990) and in Estonia (Puura *et al.* 1987, Kiipli 1989), the bedrock topography represents, in places, exhumed surface of very old age.

The sloping of the Earth's surface toward the present Baltic Sea is a Cenozoic feature. It is important to know when the general slope was formed. According to Karabanov *et al.* (1993), in the Late Oligocene - Quaternary, the East-European Craton south of the Gulf of Finland and east of the Tornquist's line experienced a regional inclination (inversion) from the Ukrainian-Voronezh Uplift to the Baltic Sea. According to Puura (Пуура 1980), denudational processes were activated earlier, in the Palaeogene by an asymmetric uplift of Fennoscandia, and a concurrent uplift of the southwestern part of the East-European Craton. The Baltic Sea depression as a region of minimal uplift accumulated waters from surrounding areas.

Highly disputable is the role of local differentiated movements. Rähni (Ряхни 1973) was of the opinion that a number of the bedrock surface features corresponded to blocks, mobile in the Quaternary period. During the course of the past 12,000 to 13,000 years some of those blocks rose at least 30 m, others sank 10 to 20 m. On the neotectonic map of the Soviet Baltic Republics (Šliaupa *et al.* 1982) five neotectonic uplifts (Haanja, Keila, Nuia, Pandivere and Võru) and two neotectonic depressions (Pärnu and Võrtsjärv-Mustvee) are shown, all in a good accordance with bedrock surface features.

The question now is: how many structural features of the sedimentary bedrock are actually in agreement with palaeolandforms? A special study, the morphotectonic analysis in northeastern (Вахер и Таваст 1979) and southeastern (Вахер и др. 1980a) Estonia has proved that direct morphostructures are infrequent in those areas (Raukas *et al.* 1988). Of the 18 zones of disturbances, studied in detail in northeastern Estonia, only the Viivikonna Anticlinal (Fig. 120) is expressed in the bedrock topography. On the other hand, of

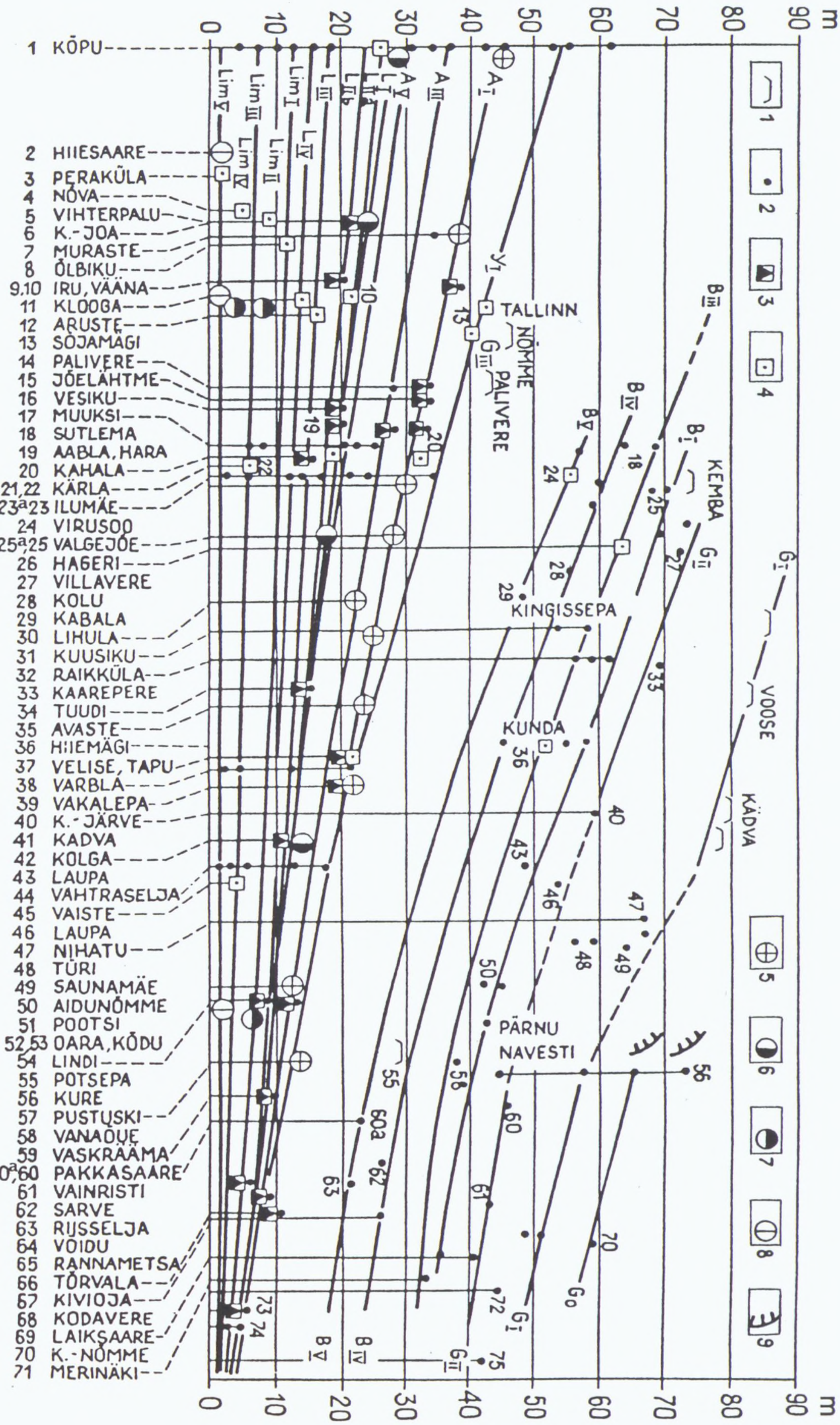


Fig. 128. Distance diagram of raised Late-glacial and Holocene shorelines of the Baltic Sea in Estonia (after Kessel & Raukas 1979): 1 - glaciofluvial delta; 2 - coastal landforms (beach ridges, bluffs and cliffs, bars, boulder fields a.o.); 3 - buried organic deposits; 4 - lagoons; 5 - Ancylus malacofauna; 6 - Litorina malacofauna; 7 - Limnea malacofauna; 8 - finds of *Mya arenaria*; 9 - margin of the glacier.

the six dome-like placanticlinales, known in northern Estonia, four are expressed in the bedrock topography (Fig. 119), one is not expressed at all, and for one there is too little information available.

The structural-topographical relations in the sedimentary bedrock are rather complicated in the Haanja Heights (Fig. 117). The southern part of the heights intersects with the steeply-dipping flank of the anticline whose amplitude accounts for more than 120 m on the bottom of the Upper Devonian Pļaviņas Stage. This flank is not expressed in the bedrock surface. The amplitude of the northerly dome on the above-mentioned bottom is some 50 m (Паасикиви 1966). Thus, the possible amplitude of relatively young movements does not exceed 50 m. It forms about one half of the relative height of the Haanja Heights in view of its bedrock surface.

The fairly large Pandivere Upland is intersected by six zones of disturbances trending mainly northwest to southeast, and with the steeper-dipping flanks of the anticlines mostly to the northwest. They divide the area into several blocks with a throw relative to one another of 10 to 20 m (Fig. 115). A very gentle southward regional dip is characteristic of the area on the whole, and no traces of major block (or fold), corresponding to the Pandivere Upland, have been found (Вахер и Таваст 1979, Карукяпп и Таваст 1985).

Disagreement between a medium-size bedrock surface feature and structural features of the sedimentary bedrock is best revealed in the Ahtme Elevation in northeastern Estonia (Fig. 120). Against a background of homoclinal dip of the Middle Ordovician Kukruse Stage, the Ahtme Zone of diturbances with the total vertical amplitude of up to 18 m, is distinctly observable. It intersects the elevation diagonally and is not expressed in the bedrock surface.

The above provides a basis for the conclusion that the formation of the bedrock topography has been but little effected by postglacial and recent blockwise differential movements. Morphotectonic analysis enables one to prove the occurrence of those movements in a few cases only.

The area of Fennoscandian continental glaciation was repeatedly subjected to crustal downsinks and uplifts caused by ice sheet's loading and unloading. The recent and present uplift of Fennoscandia, including the northwestern part of Estonia, is an evidence of glacioisostatic relaxation of the crust.

The Late-glacial and Holocene crustal movements have been studied in more detail. Based on numerous distance diagrams compiled, starting from Ramsay (1929) and ending with Kessel and Raukas (1979) (Fig. 128), the following conclusions have been reached:

(1) in the Late-glacial and Holocene, this area experienced crustal uplift and tilting;

(2) the amount and rate of the uplift decreased from the northwest to the southeast;

(3) the rate of the uplift has been decreasing from the Late-glacial to the present time.

The total uplift in Estonia without the eustatic factor is at least 85 m; however, extrapolating the glacial lake level G_1 along the 355° azimuth up to the Kõpu Peninsula (Hiiumaa Island), the total uplift reaches 115 m (actually, the highest shoreline has there an height of 60.8 m a.s.l.). The amount of the total uplift in southeastern Estonia is not known so far. It has been estimated at about 25 m (Орвику 1960в).

Most of the uplift was evidently realized at the end of the Late-glacial. The preliminary calculations, now a little out of date, suggest that in northwestern Estonia the uplift proceeded at a rate of 30 mm per year, and reached the amount of 65 m in the time interval from the end of the Middle Dryas up to the Pre-Boreal (Кессел и Мийдел 1973).

In the Holocene, the rate of uplift decreased considerably. At that time, the average rate of uplift on the Kõpu Peninsula, in Tallinn and Pärnu was 5.0, 4.0 and 1.0 mm per year, respectively.

Through the postglacial, the rate of uplift probably decreased unevenly. As is seen from the gradient curve (Fig. 129), the rate of uplift decreased during the periods lasting from 9,500 to 9,000 and 6,800 to 5,300 years ago. Thereafter, the rate stabilized to decrease again some 4000 years ago (Кессел и Мийдел 1973).

The trend surface analyses have revealed that during Ancylus time northwestern and western Estonia rose fast, but in Litorina time these areas experienced considerable sinking (Miidel 1995). Irregularities in uplift are also expressed in the change of the direction of tilting from 155° (local ice-dammed lakes) to 130-135° (recent crustal movements) (Пярна 1962, Кессел и Мийдел 1973, Kessel & Raukas 1979, Miidel 1995).

The gradient of the Baltic Ice Lake shorelines changes across the line Pärnu - Navesti River - Narva (Пярна 1962), forming a hinge line in the crustal tilting. To the northwest from the hinge line, or belt, the rate of uplift was higher than in the southwest. The belt coincides in places with tectonic dislocations in the basement and Palaeozoic rocks. It is assumed (Орвику 1960в, 1969), that in the Late-glacial the dislocations became active. Regionally, this belt partly belong to the Svecofennian deep-seated fault zone Vyborg - Pärnu - Liepaja (Пуура 1979, Раукас и Хюваринен 1992). Another hinge zone in the area of the Gulf of Finland was demonstrated by Donner (1966, 1969, 1970). The hinge zone, expressed as a change in shoreline gradients between Finland and Estonia, must be controlled by tectonics (Donner 1970).

By means of geodetic, mareographic and cartographic methods it has been established that crustal movements have continued up to the present (Желнин 1960, 1964, Zhelnin 1966, Валнер и Желнин 1975). According to the most recent map (Vallner *et al.* 1988), the major part of Estonia is rising at a rate up to 3 mm per year northwest of the line Lake Võrtsjärv - Tartu), whereas southeastern Estonia is sinking at

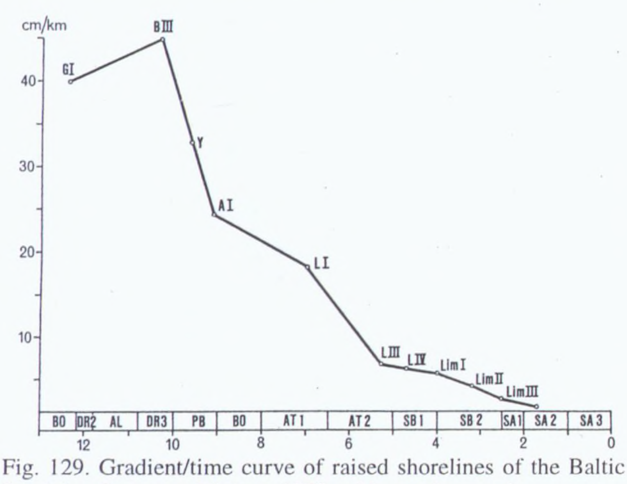


Fig. 129. Gradient/time curve of raised shorelines of the Baltic Sea (after Кессел и Мийдел 1973, with supplements).

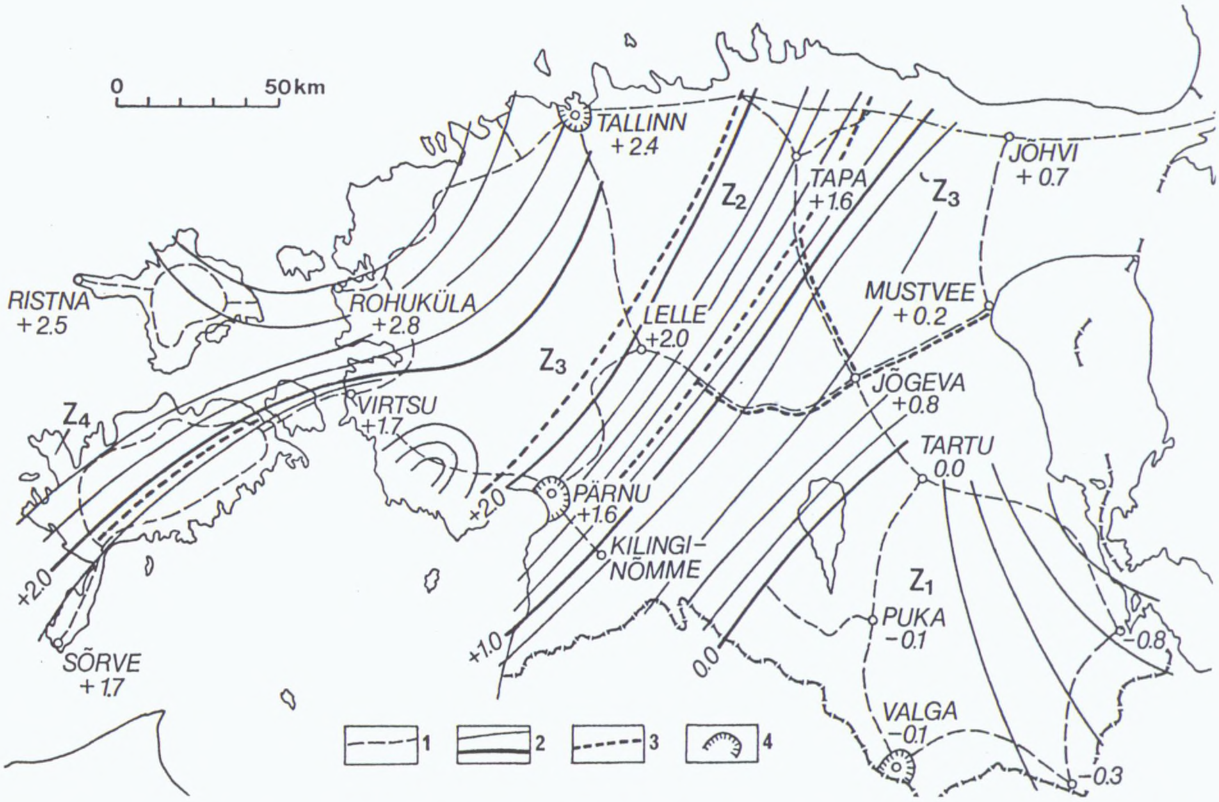


Fig. 130. Scheme of isobases and planes of annual velocities of recent vertical movements in Estonia (after Vallner *et al*. 1988): 1 - levelling netvork; 2 - isobases; 3 - boundaries of planes of annual velocities of recent vertical movements; 4 - local area of subsidence.

a rate of up to 0.8 mm per year (Fig. 130). In general outline, the isolines of the recent crustal movements coincide with the late- and postglacial ones, but reveal a great deal of irregularities. This may be due to block movements along rejuvenated, NE-SW-oriented fault zones (Сильдвээ и Мийдел 1978, 1980; Vallner *et al*. 1988). The most prominent gradient zone along the line Pärnu - Tapa - Kunda coincides with the zone of deep-seated fault in the bedrock and partly with the above-mentioned hinge line. River activity in the Pärnu River basin has been noticeably influenced by crustal movements in the zone (Мийдел 1966б, Сильдвээ и др. 1973, Miidel 1991). Recently, it was demonstrated (Miidel 1994) that the Pärnu - Kunda gradient zone may lie at northeastern border of the central part of a present regional uplift anomaly, established by Svensson (1989, 1991).

Vallner (Валлнер 1978, 1981; Vallner *et al*. 1988) divided Estonia into five annual velocity planes of vertical movements (southern, central, southeastern, southwestern, and western Saaremaa), moving in parallel or under a small angle relative to one another.

During the past 20 years, the problems of seismicity have become topical. Prior to the Osmussaar earthquake on October 25, 1976, Estonia had been considered aseismic. The unexpectedly strong Osmussaar earthquake aroused interest in seismic phenomena. As a result of numerous studies (Klaamann 1977, Булин 1978, Аботиня и др. 1988, Sildvee 1988, 1991, Nikonov & Sildvee 1992, Sildvee & Vaher 1995), the former concepts about the seismicity of Estonia have been entirely changed.

According to Sildvee and Vaher (1995), twenty four macroseismic and numerous small (M<3.5) instrumental events have occurred in Estonia since 1602 till 1991 (Fig. 115), their intensity being 3 to 7, magnitude 1.5 to 4 and depth of the focus 5 to 14 km (for local shocks about 5 km). The most powerful was the Osmussaar earthquake (magnitude 4.7, intensity 6 to 7), with a epicentre 5 to 7 km northeast of the island. Its impact was felt over an area of 191,000 sq km (Kondorskaya *et al*. 1988). According to macroseismic evaluations, the depth of its focus was about 13 km, according to instrumental studies - 10 km. After Kondorskaya *et al*. (1988) the epicentre is connected with the fault running from the Central Baltic via the northwestern part of Hiiumaa Island to the coast of Finland near Hamina. However, it is not excluded that the epicentre lies at the fault line Lahti-Porkkala (Сильдвээ и Мийдел 1980, Раукас и Хюваринен 1992).

The distribution of the earthquakes in Estonia is not random. Their occurrence must be, to a certain extent, correlated with geology and tectonics. The majority of macroseismic epicentres are located in the northwestern part of Estonia, *i. e*. in the region where the rate of the uplift is highest. A considerable number of macroseismic events occur in the Paldiski - Pskov gravity and aeromagnetic anomaly zone which is related to a deep-seated fault (Sildvee & Vaher 1995). The epicentres of small instrumental events, concentrated in the northern part of Lake Võrtsjärv and its surroundings, may be located in a minor cross fault. It is of interest that they are situated in a gradient zone of recent crustal movements.

VIII FORMATION OF THE TERRITORY

FORMATION OF THE EARTH'S CRUST

Formation of main features of the continental crust in the region

The development of the Earth's crust and its main parameters in the northern Baltic and adjacent areas (Puura & Flodén 1996) was controlled by two processes: 1) creation of the crust during the Palaeoproterozoic Svecofennian orogeny, and 2) changes in the juvenile continental crust caused by bimodal rapakivi-anorthosite magmatism (Palaeo- to Mesoproterozoic).

Recent basic achievements in geophysical and geological studies of the Earth's crust have changed the general views on the formation of the earliest, Precambrian continental crust.

First, the applicability of the plate tectonic fundamentals to the Proterozoic crustal growth was veryfied in our crustal domain, on the example of the Palaeoproterozoic Svecofennian crust in the northern part of the Gulf of Bothnia (BABEL... 1990). It became evident that also the Precambrian structures in the Baltic basement should be interpreted in terms of plate tectonics.

Second, according to the general time trends in crustal thickness, the Proterozoic crustal domains have the thickest crust (40-55 km) compared with the domains formed earlier (27-40 km in the Archean) or later in the Phanerozoic (Durrheim & Mooney 1994). Consequently, the extraordinarily thick crust of the Palaeoproterozoic Svecofennian Domain (50-65 km) fits well to the global regularities.

Third, in the areas subjected to crustal shortening due to compression during the orogenic stage, lateral displacements of deep crustal slabs into the upper crust are common. In the overview of the Precambrian crust in the Fennoscandian Shield and adjacent Russian Platform (Gorbatschev & Bogdanova 1993), the model of crustal stacking as the mood of formation of the beltiform Belarussian-Baltic, granulite terrain (including southern and western Estonia) was introduced.

After the formation of the juvenile continental crust, substantial changes may have taken place in its composition and thickness. One of the most significant processes of the crustal deformation accompanied by crustal thinning is rifting and the related magmatism in tensional stress fields. Formation of the rapakivi and related rocks of Fennoscandia has long been interpreted as a result of magmatism in tensional tectonic environments. The bimodal composition of magmas was due to the melting of both the lower crust and uppermost mantle (Rämo 1991). The bulk of felsic magmas intruded into the uppermost crust and extruded on its surface. The mafic magmas only in places reached the Earth's surface (anorthosite plutons and diabase dikes and sheets), while the bulk of their volume remained in the low levels of the crust. Crustal thinning due to basaltic underplating was stated around the Vyborg Rapakivi Pluton (Elo & Korja 1993) and in the area of the Åland Pluton (BABEL... 1993). The deep crustal signatures of the rifting type were identified by seismic sounding near the Åland Rapakivi Pluton in the southernmost part of the Gulf of Bothnia. The whole area of crustal thickening down to 50-40 km in the central part of the Svecofennian orogenic domain was interpreted as a result of the bimodal anorthosite-rapakivi magmatism in tensional environments (Puura & Flodén 1996) in the time span of 1645-1540 Ga (Rämö *et al.* 1996).

Thus, the thick crust in southeastern Estonia is considered as a survived primary Svecofennian crust, while the considerably thinned crust of northern and western Estonia is due to rapakivi-time reworking, including basaltic underplating.

Formation of the continental crust during the Svecofennian orogeny

A principal question is what kind of conditions existed in the region before the Svecofennian orogeny. By means of Sm-Nb isotopic studies of typical Svecofennian rocks of southwestern and central Finland (Huhma 1986) and granulites of central and southern Estonia (Puura & Huhma 1993) it was established that there were no Archean crustal remnants in the internal areas of the Svecofennian orogen. Thus, the Svecofennian crust was formed as a juvenile feature in an area of oceanic crust. It can be supposed that oceanic environments existed in an area larger than the present Svecofennian crustal domain, because the latter was formed during the subsequent process of substantial crustal shortening.

However, little is known about the crustal processes before 1.9 Ga. In the border zone of the Karelian Archean continental domain, ancient rift structures older than 2 Ga have been documented. It was concluded that they originate from the times of opening of the pre-Svecofennian ocean. The ophiolite series of the Outokumpu area in the Karelian-Svecofennian transition zone has been considered to be a fragment of the pre-Svecofennian oceanic crust of an age ca. 2.02 Ga which was obducted on the edge of the Karelian Craton during the Svecofennian orogeny (Koistinen 1981).

Plutonic, volcanic and sedimentary orogenic suites, isotopically dated up to now in the internal parts of the Svecofennian Domain, show ages of magmatism and metamorphism in the time span of 1.9-1.8 Ga (Huhma *et al.* 1991). It means that no direct evidences on pre-Svecofennian rocks have been found there so far. However, at least two circumstances might point to the existence of the sialic crustal compounds before the final Svecofennian orogeny. The SHRIMP dating of detrital zircons from clastic metasedimentary rocks of the Svecofennian interior (Huhma *et al.* 1991, Claesson *et al.* 1993) has shown that at least in Finland and Sweden there existed source rocks with an age of *ca.* 2.1-1.9 Ga. Basing on the geochemical and isotopic studies of the South-Estonian granulites, it was concluded that original volcanic rocks were created in the crust having sialic signatures (Puura & Huhma 1993). As a whole, the data obtained suggest probable existence of a considerably thick sedimentary, or volcanic layer, or both, on top of the mafic ocean floor before the island arc magmatism started. Somehow it reminds recent geological environments transitional from oceans to continents.

Basing on deep seismic (BABEL... 1990) and magnetotelluric soundings (Korja *et al.* 1993) and on the interpretation of petrological data of igneous rocks (Hietanen 1975), the fossil volcanic island arc structures have been determined in the northern part of the Svecofennian Domain. The Vihanti - Pyhäsalmi and Tampere zones in Finland belong to the kind. Considering the petrological data, the early orogenic metavolcanic and -plutonic rocks in the Baltic area carry the same signatures.

The fossil island arc systems with deep (up to 80 km) fossil roots of subduction zones have been identified in the northern part of Svecofennia (BABEL... 1990). Taking into consideration the isotopic signatures, the early orogenic igneous rock bodies in the internal parts of Svecofennia have been interpreted as of Paleoproterozoic juvenile origin (Huhma 1986).

It has been stated that in the Fennoscandian Shield area, the thick Svecofennian continental crust was created during a short time span between about 1.9 and 1.77 Ga (Huhma 1986, Gorbatschev & Bogdanova 1993, Koistinen 1996). The early stage of the Svecofennian island arc plutonic and volcanic activity occurred during 1.90-1.88 Ga: intrusive Svecofennian rocks in Finland have zircon ages 1890+/-10 Ma which include early-, syn-, late- and post-orogenic rocks of central Finland. Younger Svecofennian rocks lie within the potassium granite-migmatite area in southern Finland and probably in Estonia. Potassium migmatite-granites dated at 1840-30 Ma evidence of a large secondary west-east-striking zonal magmatism within the young (1850 Ma) orogen. In the same zone, post-tectonic granitoids at 1820-1770 Ma occurred (Koistinen 1996). In Estonia, the possible age analogues are Virtsu and Taadikvere plutons (Fig. 5).

Judging by the isotopic data concerning the formation of the Svecofennian crust in Finland, the Estonian basement probaly belongs to an area where both early orogenic at 1.890+/-10 Ma island arc and late orogenic within-crustal zonal granite magmatism occurred. The accretion of Svecofennia to the present SW border of the Late Archaen Karelian crustal domain and the late orogenic tectonic, metamorphic and magmatic transformations came generally to an end at ca. 1.8 Ga.

The crust of Svecofennia, the thickest one compared with older and younger crustal domains of the East-European Craton (Platform), was created during several stages. After the pre-orogenic oceanic igneous activity and sedimentation, the overwhelming magmatic and sedimentary activity of the island arc (subduction) stage created rock assemblages together with the interarc sedimentary basins available for later observations. Intraorogenic deformation and metamorphism together with within-crustal partial melting (granitoids) changed the primary rock assemblages. Several phases of deformation and metamorphism have been documented and dated in the time span of 1.9-1.8 Ga.

The crustal shortening and corresponding thickening started with the formation of subduction zones and island arcs above them. The above early orogenic igneous rocks and the first stages of deformation, metamorphism and partial melting (migmatisation) were related to extraordinarily intense processes of subduction contemporarily occurring in several island arc systems within Svecofennia (Gorbatschev & Bogdanova 1993). Erosion of island arcs became a source for sedimentary depressions between the arcs, the largest of which in Estonia is the Alutaguse Zone.

The next stage of crustal shortening occurred as compression and fragmentation of fossil island arcs and sedimentary basins, with the second stage of general folding and metamorphism. In the late orogenic period, other phases of faulting and infracrustal granitoid magmatism occurred, the latter being represented by the potassium granite intrusions and migmatites. The step-wise cratonisation of Svecofennia finished with the intrusion of the so-called post-orogenic granitoids at 1.82-1.77 Ga.

Already before the rapakivi magmatism at about 1.65 Ga, the young crust was deeply eroded up to the levels of mainly amphibolite, in places to granulite grade. Starting already from the early orogenic island arc stage at about 1.9 Ga, single mountain chains in marine surroundings became subjects of intense erosion. In this early stage, depositories of sediments located in the between-arc space. Afterwards the whole orogenic domain uplifted at about 1.87-1.85 Ga and turned into a continent with mountain chains. Intense erosional processes planated the original complex mountain relief. Erosional debris was removed outside the whole domain and their depositories are not known yet.

Crustal changes during the Meso- to Neoproterozoic

The above principal reconstruction of the Svecofennian crust during rifting and rapakivi and related magmatism at 1.65-1.54 Ga was coupled with one more tectonic blocking of the crust and differentiated movements of the blocks reaching kilometres (Koistinen 1996). The volcanic complexes with volcanic topography above the rapakivi plutons diversified the mountain environments. The concurrent and subsequent erosion planated the Earth surface composed of both rapakivi and pre-rapakivi rocks. Before the first epicontinental Jotnian sedimentation at about 1.4 Ga, the first perfect peneplain was worked out in the Svecofennian Domain, as it can be concluded from the survived fragments of Jotnian clastic sedimentary basins in the Bothnian Sea - the Satakunta and Lake Ladoga areas (Koistinen 1996).

However, the post-Jotnian (about 1.2 Ga) and younger (at about 1.1-0.9 and 0.6 Ga) events of mafic dike magmatism evidence about the repeated blocking of the crust. The gentle depressions, diversified by faults with horsts and grabens identified in the Jotnian sedimentary cover, carry information on the reaction of the crust on both tensional and compressional stresses in the region. The latter data have been gathered from the neighbouring areas of Fennoscandia and the Baltic Sea (Puura & Flodén 1996, Koistinen 1996).

The pre-Late Vendian/Cambrian peneplain has been worked out upon the complex target composed of Svecofennian orogenic, rapakivi-time anorogenic, and Jotnian and later epicratonic sedimentary and within-cratonic plutonic rocks.

Crustal changes during the Phanerozoic

The pre-Late Vendian/Cambrian peneplain marking the pre-Phanerozoic Earth's surface has fully survived in Estonia and surrounding areas under the sedimentary cover. As shown above, the total thickness of the survived Palaeozoic sediments is less than 800 m. According to some estimations, the total rate of erosion of the Phanerozoic sedimentary cover reached 500-800 m. During the Late Caledonian compressional phase, local vertical displacements in zones of disturbances did not generally exceed 50 m in Estonia. As an exception serves the Riga-Pskov fault zone on the

Estonian-Latvian border where it reached 600 m. All the other tectonic phases have caused weaker deformations.

As a summary, the crustal conditions during the Phanerozoic can be classified as most stable cratonic ones. Since the opening and spreading of the North-Atlantic Ocean, the Fennoscandian-Baltic region as the northwestern part of the East-European Craton was situated in a most quiet tectonic regime in terms of plate interactions which promoted isostatic levelling of the crust. During the Cenozoic, the Svecofennian Domain became a morphostructure with the slightly elevated transitional zones and subsided (locally even submerged) interior. The primary Svecofennian crust is thick and light. The secondary crust, reworked by rapakivi-anorthosite magmatism, in the Svecofennian interior is thinned and denser. The buoyancy differencies between these crustal sections are considered as having been responsible for the formation of the Baltic Sea drainage basin with the Baltic Sea in the centre (Puura & Flodén 1997).

The late Cenozoic glaciations caused repeated glacioisostatic downwarping and uplift of the Earth's surface in and around the Fennoscandian centre of glaciation. At present, the upper crust is in a stage of postglacial uplift. The glacioisostatic crustal movements were superposed to the Cenozoic Baltic depressional morphostructure and, thus, diversified the morphology of the latter. They have been more localised and short-term compared with its larger and long-term features.

After the formation of the Palaeo-Mesoproterozoic crust, a long-lasting continental period followed. A deep denudation shear on the bedrock marks this important discontinuity in the pre-Vendian geological history of the region. The topmost 0.5 to 150 metres of the basement have been changed due to weathering.

In general outline, the bedrock succession in Estonia is divided into three parts (in ascending order): the mainly terrigenous part of Vendian-Tremadoc age, the Ordovician-Silurian carbonate formation and the Devonian, predominantly terrigenous rocks.

The Vendian-Tremadoc rocks occur all over the East- European Platform, particularly in its western and central parts. In Estonia, they form a complex with a thickness of up to 220 m and are non-existent only in the area of the Lokno-Mõniste-Valmiera Uplift (Fig. 131).

Despite the stratigraphic incompleteness and several minor or major hiata, the general stability and regional uniformity of the succession is remarkable. To provide a better understanding of Estonia's geological development during the Vendian-Tremadoc, the palaeogeographical situation in much larger area than Estonia will be discussed (Figs. 131-137, Rozanov & Lydka 1987).

The sedimentation of the lower, terrigenous part of the bedrock in the area under consideration was discontinuous due to the cyclically repeated transgressions and regressions related to the tectonic evolution of the platform in general. In the northwestern part of the area, seven evolutionary stages of deposition may be distinguished. The sedimentary complexes formed during these stages differ in the distribution, structure, thickness, facies composition, lithological characteristics, number of hiata and several other features (Fig. 132). In ascending order they correspond to the following stratigraphic intervals (Tables 5,6): Valdai Subseries; Baltic, Liivi, Aisčiai and Deimena series, late Middle Cambrian - early Late Cambrian and late Late Cambrian - Tremadoc (Fig. 132). The former four complexes are characterised by easily defined cyclic structure and regular alternation of lithotypes reflecting the transgressive-regressive nature of sedimentation. Three upper complexes occur in Estonia as remains of

Fig. 131. Distribution pattern of basal deposits of the sedimentary cover in Estonia: 1 - Valdai; 2 - Baltic; 3 - Liivi; 4 - Aisčiai; 5 - Middle Cambrian; 6 - Devonian.

the complexes, most completely developed in the west and east. These complexes are mostly represented by sandy deposits formed under shallow-water conditions and complicating the interpretation of the development.

Valdai evolutionary stage

Widespread typical platform facies in the northwestern part of the East- European Platform are known from the second half of the Vendian. The subsidence of the Earth crust at the beginning of the Vendian embraces only the narrow linear depressions, the so-called aulacogenes, seated deep in the pre-platform basement far from Estonia.

The early Vendian volcanic, sedimentary-volcanic and sedimentary rocks of continental origin deposited during the second half of the Drevljan Stage and are confined to the extensive Orsha Depression which partly reached the area under consideration (Fig. 133A).

The Valdai complex, representing the Upper Vendian in the East-European Platform, is of wider distribution and similar in character throughout the platform. Its deposits overlie transgressively older accumulations of the platform bedrock cover or occur immediately on the Palaeo-Mezoproterozoic basement.

The distribution of the Valdai complex testifies to the emplacement of a new structural framework on the East-Europen Platform (Craton) which existed throughout the terminal Neoproterozoic and initial Palaeozoic. Its principal characteritic feature is an extensive eastern transgression resulting in the formation of the large Moscow Basin, which influenced upon sedimentation in the northeastern part of the platform, including Estonia.

The Valdai section can be divided into two great sedimentary cycles: Redkino (below) and Kotlin (above).

Redkino Age. The Redkino transgression was discontinuous with intermittent regressions. Evidence is derived from the rhythmic development of the strata. Each new transgression started with a basal coarse-clastic sandy unit and graded into a clay unit corresponding to a submerged maximum. Deposits of the regressive phase have not usually preserved.

The Redkino rocks, not known in Estonia, are distributed immediately beyond its boundaries - in the Leningrad Region and eastern Latvia (Fig. 133B). According to the specific features, such as the occurrence of soft-bodied *Metazoans*, glauconite, early diagenetic pyrite and phosphates, they accumulated under typical marine conditions.

The end of the Redkino accumulation is marked by an uplift in an extensive area. It changed sedimentation conditions and broke the link with the ocean.

Kotlin Age. The rocks of the Kotlin Stage are the only representatives of the Valdai evolutionary stage in Estonia. The uplift of the northwestern part of the platform at the end of the Redkino Age led to a slight erosion and subsequent subaerial weathering of the topmost beds of the Redkino Stage. This short-term continental period was followed by an extensive Kotlin transgression. The development of the Moscow Basin proceeded from the structural framework of the preceeding Redkino Age, being the major subsidence zone in the centre of the platform. Changes in the framework can be

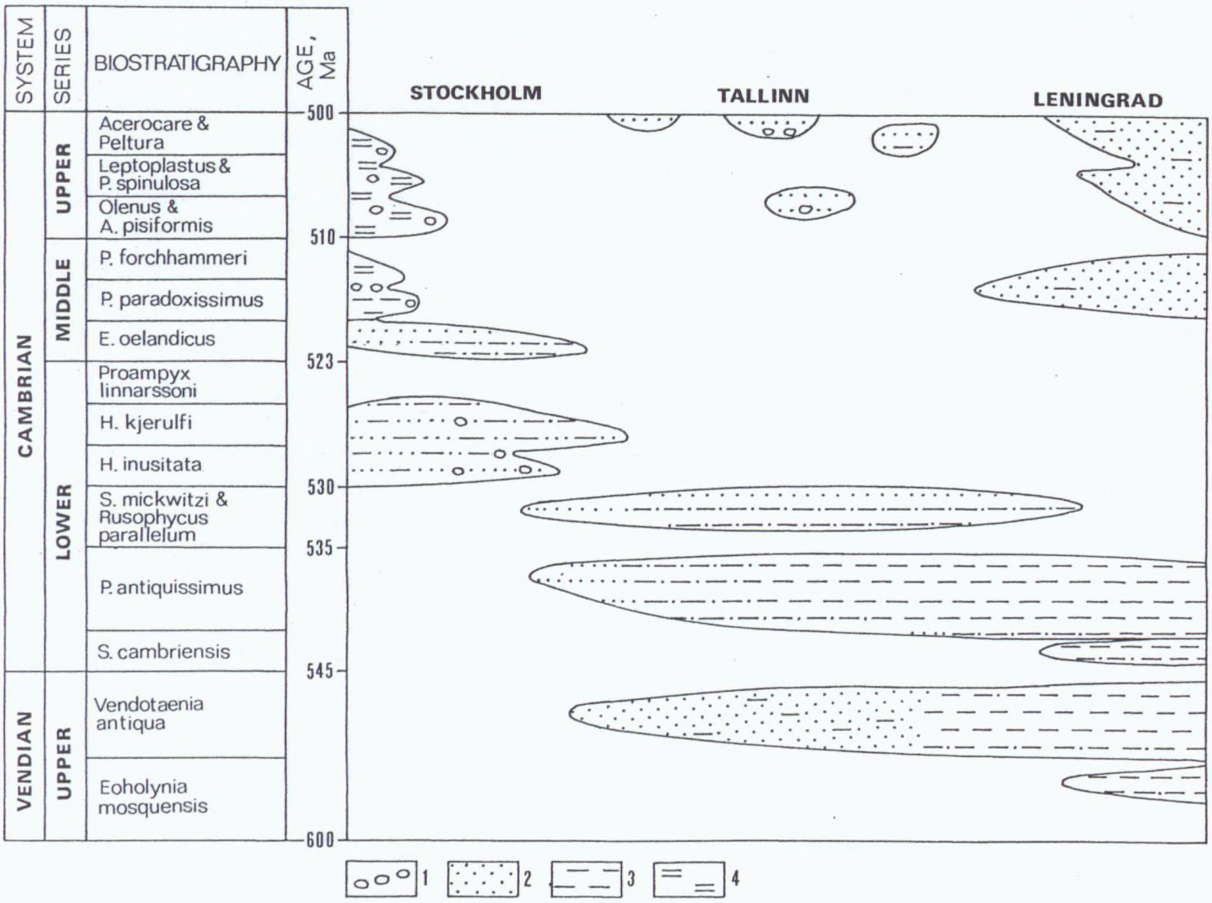


Fig. 132. Distribution of Vendian and Cambrian rocks and hiata along the Närke (central Sweden) - Kostov (Leningrad Region) transect and their relationship with biozones: 1 - conglomerate, gritstone; 2 - sandstone; 3 - argillaceous rock; 4 - black shale (Alum Shale). White areas indicate hiata.

seen in the northwest where the sedimentation spread extensively westwards and most of Estonia was submerged (Fig. 133 C).

In Estonia, the rocks of the Kotlin Stage rest directly on the weathered crystalline basement. At the beginning of the Kotlin transgression, the mellow surface of the weathered basement was partially destroyed and redeposited in a basin as follows from the composition of the basal Oru Member. Upwards the share of the clastic material derived from the weathered crust decreases. Nevertheless, the enchanced content of feldspars, micas and kaolinite and their poor roundness in rest of the Gdov Formation (Table 6) indicate low-grade reworking of the clastic material inherited from the surrounding areas of igneous or metamorphic, or both, rocks. The grain-size decreased constantly upwards until the formation of a thick stratum of the "Laminarite clays" (Kotlin Formation). The clay component of this argillaceous rock is low in kaolinite and high in illite-chlorite (Пиррус 1970, 1983), which may be partially due to the redeposition of Redkino deposits into the Kotlin basin (Пиррус 1980, Pirrus 1987).

The assemblage of authigenic minerals, total lack of the body fossils, weakly developed varved structure of clays and low content of boron (Битюкова и Пиррус 1979) suggest accumulation of the argillaceous deposits of the Kotlin Age in a basin similar to ice-dammed lake. The above-mentioned deposits wedge out abruptly without transitional facial belts along the meridian in central Estonia (Fig. 133C). A similar kind of accumulation is observed in the recent arctic sedimentation regime which is in accordance with palaeospatic reconstructions of Late Proterozoic time (Scotese *et al.* 1979).

The terminal phase of the Kotlin clay accumulation in Estonia is marked by some shallowing of the sedimentary basin and the appearance of interbeds of sand- and siltstones (Laagna Member).

The succeeding uplift of Estonia's territory is marked by the weathered uppermost part of the Kotlin Formation (Менс и Пиррус 1969, 1970). The short-term (small thickness of the weathering crust, absence of kaolinite-zone) subaerial development was followed by the formation of the shallow-water Voronka basin, the variegated silty-clayey deposits of which (Sirgala Member) are upwards gradually replaced by well-sorted quartzose sandstones (Kannuka Member). More likely, this part of the Vendian sequence represents a small separate sedimentary cyclithe of regressive origin.

The present limits of the Kotlin Stage are erosional, except the northwesternmost part (Fig. 133C) and cannot, therefore, serve as real boundaries of the Late Vendian sedimentary basin.

Baltic evolutionary stage

Like in the preceeding Vendian, the Baltic sedimentary stage was influenced by the sinking of the Moscow Depression. After the uplift of the entire Vendian basin area at the

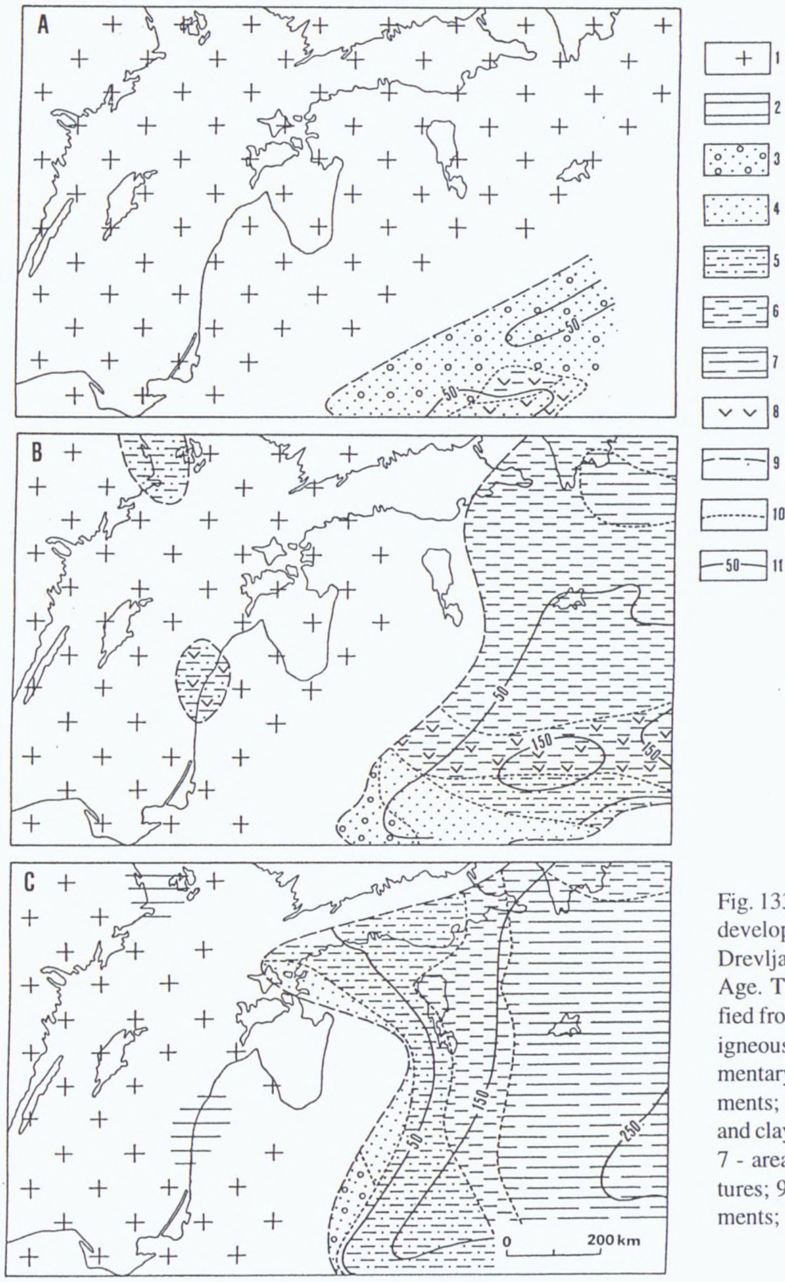


Fig. 133. Sketch-map summarizing the palaeoenvironmental development of Estonia and surrounding areas: A - during Drevljan Age; B - during Redkino Age; C - during Kotlin Age. The reconstruction of the adjacent territory is modified from Rozanov & Lydka (1987): 1 - land, represented by igneous and metamorphic rocks; 2 - land, covered with sedimentary rocks; 3 - area of dominating coarse-grained sediments; 4 - area of sandstones; 5 - area of alternating sandy and clayey sediments; 6 - area of 50-75% argillaceous rocks; 7 - area of >75% argillaceous rocks; 8 - volcanic admixtures; 9 - boundary of the present-day distribution of sediments; 10 - boundary of facies belts; 11 - isopachs.

end of the Vendian, accompanied with the denudation or weathering, or both, processes, the marine basin again revolved into the northwestern part of the platform. The conditions of sedimentation changed considerably from the beginning of the Cambrian. The configuration of the transgression remained unchanged, but the rock composition is indicative of a distinct rearrangement in the hydrochemical regime of the basin. Despite the continuing accumulation of sandy-clay deposits, a typical marine assemblage of authigenic minerals (glauconite, phosphates, early diagenetic pyrite, *etc.*) appears. This shows that by the time of the Early Cambrian accumulation, the sedimentary basin had a link with the ocean and was subject to long-term normal marine regime.

The reconstruction of the palaeobasin of the Baltic evolutionary stage is complicated enough, due to intensive denudation processes prior to the accumulation of younger, Cambrian deposits.

Rovno Age. The Rovno deposits are absent in Estonia, but distributed immediately east of its boundaries. The sequence of the Rovno Stage starts, as a rule, with grit- or coarse-grained, or both, sandstones followed by an accumulation of argillaceous rocks. In some sections, a distinct bipartite structure — two smaller scale cyclithes (rhythms) can be recognized which differ in palaeontological characteristics (Пашкявичене 1986). If the upper cyclithe is absent, the topmost beds of the lower cyclithe are usually weathered.

The lack of coarse-grained rocks within the present-day distribution area provides a basis for the supposition that during the Rovno Age the basin might have been larger and the area of Estonia could have been submerged, at least partially. This is in good accordance with the present western and northern limits of the Rovno Stage which are characterised by argillaceous rocks accumulated in the inner part of the basin (Fig. 134A).

The character of the Rovno/Lontova boundary beds in the Leningrad Region and eastern Latvia shows continuous deposition which was not the case in the other sections. In this area, the deposits of the Baltic sedimentary stage form a unique cyclithe, the basal part of which is formed by Rovno deposits.

Lontova Age. The Lontova sedimentation in the second half of the Baltic evolutionary stage took place during a marine transgression which advanced from the east and was more extensive than the previous one. The Lontova rocks in the northwest are of much wider areal extent than the underlying Rovno rocks.

Most of the present-day Estonia was submerged in the Lontova Age (Fig. 131, 134B). A relatively small amount of coarse clastic material and the small proportion of sandstone in the basal beds of the stage point to a slow transgression and smooth topography of both the basin bottom and surrounding source area.

The Lontova sequence is dominated by clays. Huge accumulation of argillaceous deposits is typical of the Lontova Age over the whole East-European Platform. The prevailing illite with an admixture of chlorite in the clay component and a relatively mature character of clastic grains provide a basis for the supposition that the Vendian siliciclastic rocks of surrounding areas were the main source of sandy-silty material of the Lontova Stage.

The Lontova sedimentation is characterised by relatively smooth changes in facies environment. Evidence is derived from the gradual decrease in sand and increase in clay content from west to east (Fig. 134B). The changes in sedimentation were caused mainly by changes in the depth of the basin (Кала и др. 1981б). In Estonia, two distinct lithofacies belts may be distinguished: prevailingly sandy facies in the west (Voosi Formation) and clayey facies in the east (Lontova Formation). The former is interpreted here as marginal shallow-water deposits formed not far from the shoreline, the latter as deposits which accumulated under quiet hydrodynamic conditions farther from the coast (Менс и Пиррус 1986). On the basis of clay content, some detailed zones can be distinguished within the limits of the clay facies belt (Fig. 134B).

The recent areal extent of the Lontova Stage, except the northwestern part of Estonia, is erosional as is suggested by the lithofacies distribution. Probably, the primary northern boundary of the basin was situated farther in the north. The occurrence of lithologically and palaeontologically similar rocks in the vicinity of Torneträsk (Bergström & Gee 1985,

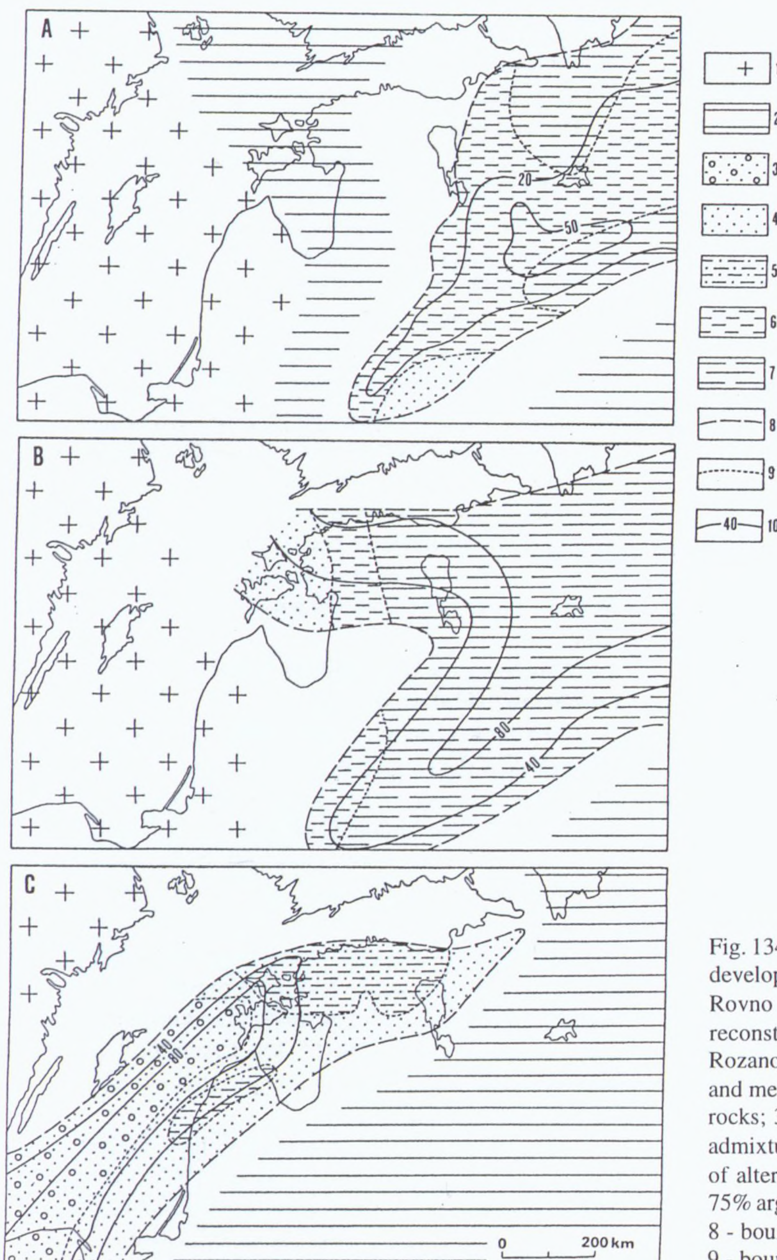


Fig. 134. Sketch-map summarizing the palaeoenvironmental development of Estonia and surrounding areas during: A - Rovno Age; B - Lontova Age; C - Dominopol' Age. The reconstruction of the adjacent territory is modified from Rozanov & Lydka (1987): 1 - land, represented by igneous and metamorphic rocks; 2 - land, covered with sedimentary rocks; 3 - area of dominating sandy sediments with gravel admixture on several levels; 4 - area of sandstones; 5 - area of alternating sandy and clayey sediments; 6 - area of 50-75% argillaceous rocks; 7 - area of >75% argillaceous rocks; 8 - boundary of the present-day distribution of sediments; 9 - boundary of facies belts; 10 - isopachs.

Vidal & Moczydłowska 1996) suggests that this part of the platform was submerged by the basin having depositional conditions similar to those in the Lontova basin in Estonia. Probably, a passage may have existed across the recent Fennoscandian Shield connecting the sedimentary basin in the north with the Lontova Sea in the south (Пуура и др. 1987, Келлер и Розанов 1980).

The Lontova rocks in southeastern Estonia are partly or completeley reduced (Кала и др. 1981б). This is due to the change in the structural framework on the pre-trilobite/trilobite Cambrian boundary and tectonic activity in the Lokno-Valmiera uplift area (Mens 1981b). The well-preserved crust of weathering in the uppermost part of the Lontova Stage in southeastern Estonia shows that erosional processes occurred mainly before the Liivi evolutionary stage.

The end of the Baltic evolutionary stage was marked by the regression of the sea and extensive changes in the structural framework.

The Moscow Depression, which so far functioned as the main subsiding area, underwent an uplift and stayed for a long time (up to the Middle Cambrian) above the sea level. In the western part of the platform, a new subsiding area—the Baltic Basin formed. It was opened to the west and had a direct connection with the ocean.

Liivi evolutionary stage

Dominopol' Age. The beginning of the trilobite-bearing Cambrian was associated with a considerable reconstruction of the structural framework of the East-European Platform. The onset of the Liivi transgression was related only to the westernmost margin of the platform where a number of longitudinal gulf-like depressions were formed. The Baltic Basin, as one of those depressions, was situated at the southern slope of the Fennoscandian Shield (Fig. 134C).

The rocks of the Liivi evolutionary stage belong to the Dominopol' Stage and are represented by alternating arenaceous and argillaceous deposits, accumulated in a shallow-water basin, and providing favourable conditions for various organisms. The flooding events with intervening water low-stands up to short-term uplifts can be established on the grounds of the composition of the Dominopol' sequence. Its lower part (Sõru Formation) is represented mainly by arenaceous rocks with several interbeds of coarse clastic sediments, while claystones are less frequent. The succeeding part of the sequence (Lükati Formation) consists of argillaceous rocks and is terminated by well-sorted fine-grained sandstones (Tiskre Formation). All the above-mentioned lithounits are separated by hiata on boundary level, which in some places are marked with conglomerate lenses. The occurrence of phosphorite nodules and phosphatized pebbles on the lower boundaries of lithounits and the presence of strong bioturbation on several levels in the middle part of the succession indicate slow sedimentation with periods of non-deposition even regardless of the high content of clay material in the Lükati Formation.

The Liivi evolutionary stage in Estonia terminates with a gradual uplifting from the south as evidenced by weathering traces in the topmost Lükati Formation.

Aisčiai evolutionary stage

Ljuboml' - Vērgale Age. Arenaceous-argillaceous strata corresponding to the Ljuboml' and Vērgale stages have deposited during the first half of the Aisčiai evolutionary stage. The Rausve and Kybartai stages are missing in Estonia probably due to the postsedimentation denudation prior to the Middle-Late Cambrian. The formations of the Aisčiai evolutionary stage overlie rocks of different age down to the crystalline basement.

The Aisčiai transgression, like the preceeding Liivi transgression, advanced from west to east but its extent was more spacious (Fig. 135A).

The composition and palaeontological data of these rocks refer to a generally shallow basin. The presence of interformational breaks, generally associated with ferruginous oolith interbeds, an abundance of glauconite and phosphates on the bedding planes (Пиррус 1986) and intense bioturbation of argillaceous beds indicate slow sedimentation with periods of non-deposition during the whole interval.

The mineral composition of the clastic component suggests that the surrounding unflooded areas were mainly covered by sedimentary rocks. The occurrence of immature minerals is indicative of the uplifts of the crystalline basement and the existence of small islands within the basin.

The well-preserved crust of weathering on the underlying argillaceous rocks and the low content of coarse clastic material in the basal beds suggest a relatively slow marine transgression and a peneplaned topography of the bottom of the sedimentary basin and the source area.

The subsidence of the Baltic Basin in the course of the Aisčiai evolutionary stage was not simultaneous; it began in the north-west, spread southwards and in a stable phase (Vērgale Age) of the transgression, the whole Baltic Basin was submerged (Vidal & Moczydłowska 1996). According to the facial distribution of rocks, the eastern coastal limit of this basin was probably not far from the recent distribution area of the Ljuboml' and Vērgale rocks. The northern boundary of the basin must have been north of the present limit of the rocks embracing most of mainland Sweden where synchronous rocks occur as small patches.

Deimena and late Middle Cambrian - early Late Cambrian evolutionary stages

The Middle Cambrian sketch-map is a generalization of the geological development of Estonia (Fig. 135B), although it presents data only on two evolutionary stages - Deimena (older) and the late Middle Cambrian - early Late Cambrian (younger), which is represented only by the rocks of the *Paradoxides paradoxissimus* Zone in Estonia. For the early Late Cambrian part of this unit see Figure 135C.

The Middle Cambrian is represented by incomplete successions over the whole northeastern part of the East-European Platform. The most incomplete erosional sections are in the East Baltic where the Middle Cambrian rocks are in places missing and the highly complicated evolution of the area is difficult to follow.

At the Lower/Middle Cambrian boundary, the sedimentation was relatively continuous (Eklund 1990, Mens 1981a, Hagenfeldt 1989a,b). At the beginning of the Middle Cambrian the deposition continued in the limits of the Aisčiai sedimen-

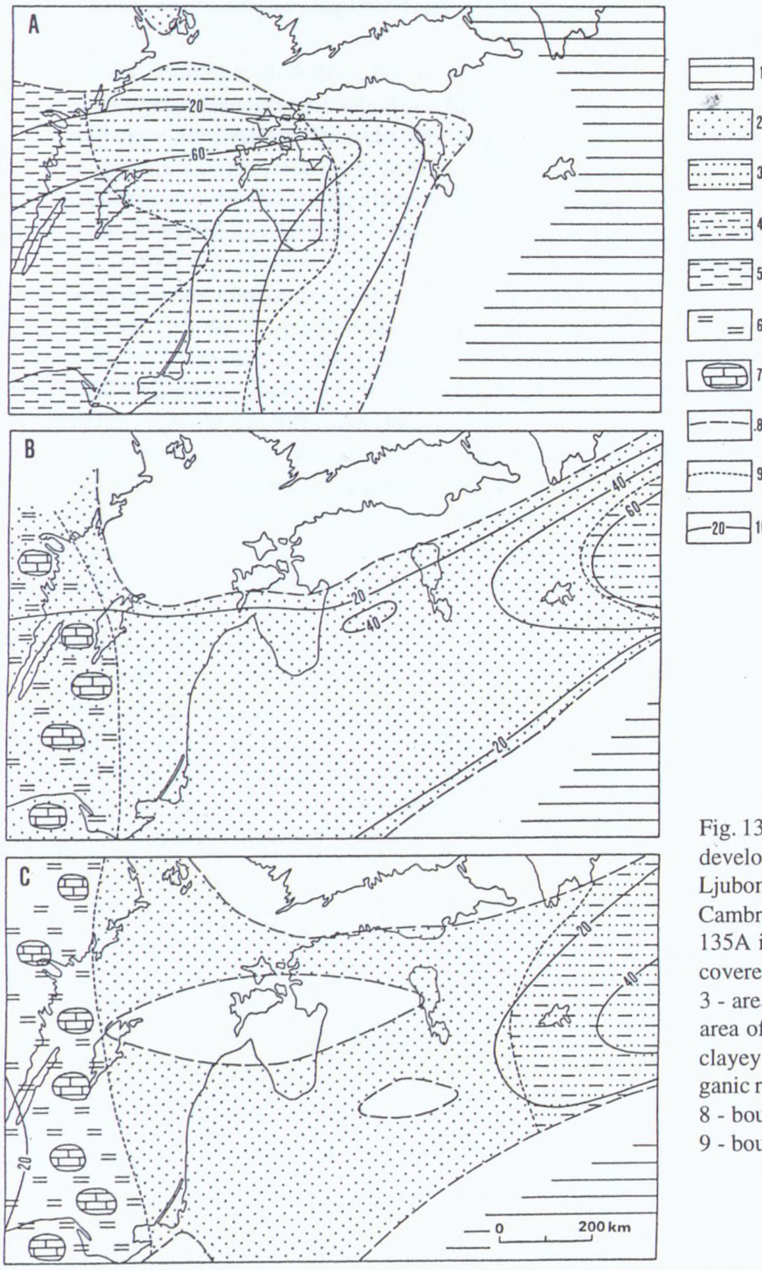


Fig. 135. Sketch-map summarizing the palaeoenvironmental development of Estonia and surrounding areas during: A - Ljuboml'-Vērgale time; B - Middle Cambrian; C - early Late Cambrian. The reconstruction of the adjacent territory in Fig. 135A is modified from Rozanov & Lydka (1987): 1 - land covered with sedimentary rocks; 2 - area of sandy sediments; 3 - area of sandy sediments with few clayey interlayers; 4 - area of alternating sandy and clayey sediments; 5 - area of clayey sediments with few sandy interlayers; 6 - area of organic rich muds (Alum Shale facies); 7 - lenses of limestone; 8 - boundary of the present-day distribution of sediments; 9 - boundary of facies belts; 10 - isopachs.

tary stage. During the *Eccaparadoxites insularis* time (Table 6) the regressive deposits of the Kybartai Stage accumulated, but they have not preserved in Estonia.

After the short post-Aisčiai continental period, Estonia, at least its southern part, was again submerged. The Paala and Ruhnu formations, which in Estonia consist of sandy rocks, are Middle Cambrian and separated from the Early Cambrian rocks by hiata. In a limited area in southeastern Estonia, the Middle Cambrian rocks rest directly on the crystalline basement (Fig. 131).

The Ruhnu Formation spreading in southwestern Estonia is considered as belonging to the *Ptychagnostus praecurrens* Zone (Mens *et al.* 1990). Two sedimentary areas with varying lithofacies can be distinguished in the western part of the area under consideration. The easternmost area is characterised by the accumulation of well-sorted fine- to medium-grained quartzose sandstones yielding a few thin interbeds of brownish claystone. Evidently, they formed under high hydrodynamic conditions not far from the shoreline in a shallow-water basin expanding from the west. The mineral composition, sorting and roundness of grains suggest that the sandstones originate from the sedimentary rocks east of the basin.

Westwards, on the islands of Gotland and Öland, the above-mentioned sandstones are replaced by grey or greenish, or both, shales with intercalations of silt- and sandstones designated to the *Eccaparadoxides oelandicus* Zone, like the Faludden Sandstone on Gotland (Hagenfeldt 1994). Towards the west, the role of silt- and sandstones decreases and interlayers of dark grey shales appear.

The *Eccaparadoxides oelandicus* deposition is followed by a general regression. Evidence is derived from abrupt changes in faunas and commonly also in the lithology (Bergström & Gee 1986). The subsequent flooding was of the largest areal extent. Thus, in the middle of the Middle Cambrian (*Paradoxides paradoxissimus* time) sedimentation was renewed in the Moscow Depression. Estonia with the adjoining areas in the west and east served as a seaway between the two intensively subsiding areas: the Moscow Depression in the east and the "Alum Shale Basin" in the west.

The latter was opened towards the Iapetus Ocean bordering the East-European Platform from the west and supplying the depositional basin with nutrient-rich water.

Rocks of the *Paradoxides paradoxissimus* time in Estonia are represented by the Paala sandstones, disseminated in its southeastern part. Probably, these deposits accumulated in the offshore conditions, and in that case the advance of the basin was from the east. This viewpoint is supported by the decrease of the share of argillaceous rocks and palaeontological evidence of coeval deposits in a easterly direction.

Late Cambrian - Tremadoc evolutionary stage

The uppermost part of the Vendian - Tremadoc clastogenic succession from the middle of the Upper Cambrian to the base of the Arenig forms a unitary epoch in the history of the evolution of the Baltoscandian sedimentary basin. The basin was opened to the west - towards the Iapetus Ocean. Eastwards it extended as far as the Moscow Syneclise.

The completeness of stratigraphical sequences of this interval varies greatly. The sediments of the Late Cambrian - Tremadoc usually rest on the eroded surface of Lower or Middle Cambrian rocks. The subsequent erosion removed large parts of the succession, and in Estonia only minor remains of the Upper Cambrian - Tremadoc rocks are preserved. The sequence is dominated by clastic rocks represented by varigrained sand- and siltstones and black shales (alum shales - kerogenous argillites). They contain rare interlayers of the brachiopod coquina ("*Obolus conglomerata*") and greenish or greyish clay.

On the basis of the spatial distribution, thickness and number of hiata, three major depositional settings (facies belts) may be recognized in the northwestern part of the East-European Platform: a black shale facies in the west, a common siliciclastic marine sandy-silty-clayey facies in the east and a transitional facies between them including Estonia's area (Mens *et al.* 1993).

The western facies belt was characterised by an accumulation of fine muds rich in organic matter (Alum Shale Formation) (see Thorslund 1960, Martinsson 1974, Andresson *et al.* 1985, Bergström & Gee 1986, Dworatzek 1987).

As is generally acknowledged, the Alum Shale Formation was deposited in anoxic shallow-water conditions, influenced by upwellings which brought nutrient-rich organically productive water from the bordering Iapetus Ocean in the west. The condensed black shale succession during a long-lasting period is indicative of the great stability in the sedimentary environment.

The Upper Cambrian of the eastern facies belt is dominated by sandstones, but both the share of argillaceous rocks and stratigraphic continuity increase rather quickly eastwards (Mens *et al.* 1990). The presence of glauconite and phosphates and faunal evidence suggest deposition in the marine basin with changeable hydrodynamic conditions.

The Upper Cambrian - Tremadoc succession of transi-

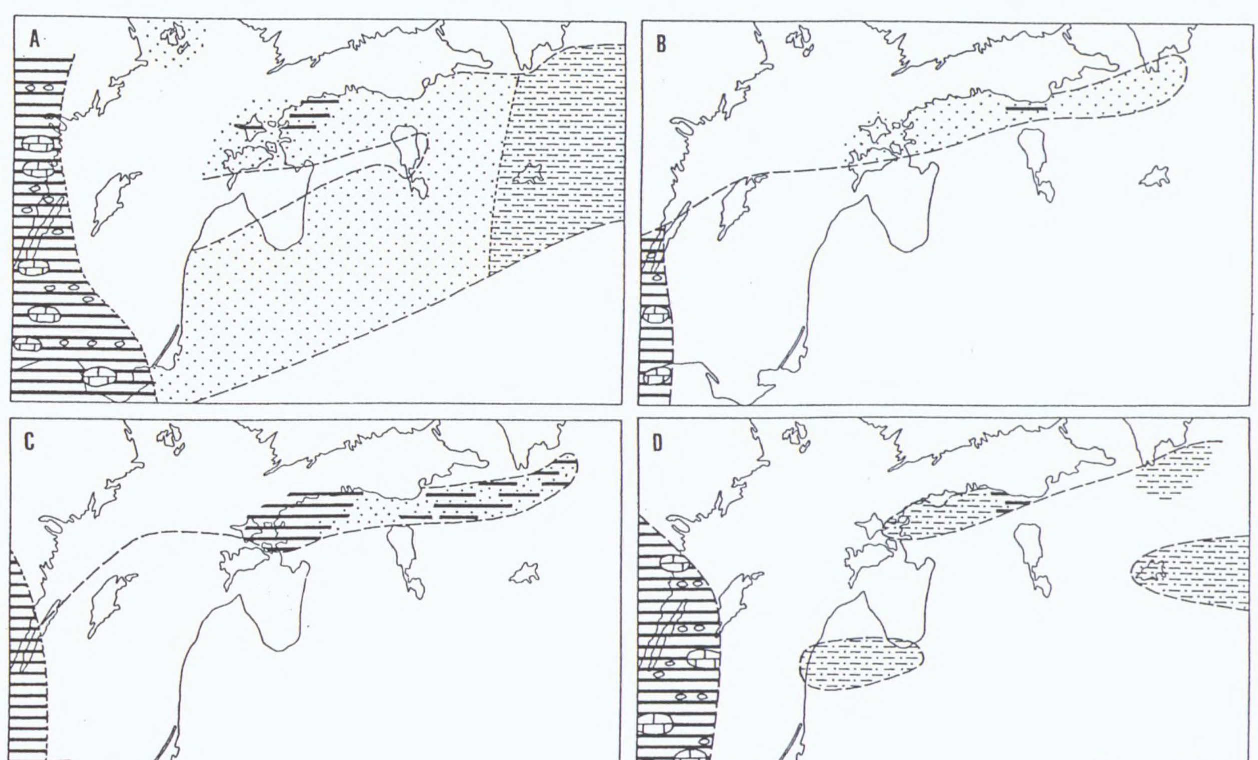


Fig. 136. Sketch-maps summarising the palaeoenvironmental development of Estonia and surrounding areas during the latest Late Cambrian and Tremadoc: A - the latest phase of the Late Cambrian; B - the beginning of the Early Tremadoc; C - Middle Tremadoc; D - Late Tremadoc: 1 - area of sandy sediments; 2 - area of alternating sandy and clayey sediments; 3 - areas of organic-rich muds (Alum Shale facies); 4 - lenses of limestone; 5 - lenses and/or interlayers of conglomerate; 6 - boundary of the present-day distribution of rocks; 7 - boundary of facies belts.

tional facies belt is very condensed and interrupted by several hiata of different duration. The preserved deposits belong to those of the deepest part of the sedimentary basin. The deposits of more elevated areas were presumably reworked and redeposited during the subsequent period. A levelling of the sea floor took place over a large area in the East Baltic. The Upper Cambrian - Tremadoc together with the late Middle Cambrian (from the *P. paradoxissimus* Zone) form the two latest evolutionary stages of the Vendian - Tremadoc clastogenic sedimentation (Figs.135C, 136). The Late Cambrian part of the former includes the rocks from the *A. pisiformis* till the middle *P. spinulosa* zones, represented by the Petseri and Ülgase formations (Table 6), indicating comparatively stable depositional environments and flattened sea floor (Fig. 135C).

The terminal evolutionary stage is characterised by the increasing instability of the depositional environment. The lithological composition and palaeontological evidence of the Tsitre and Kallavere formations demonstrate the deposition under shallow-water conditions and uneven bottom topography complicated by a lot of mobile sand bars. Within the terminal evolutionary stage four phases are distinguished (Fig. 136). They are determined by the rapid changes in the sedimentation environment, characteristic to the end of the clastogenic sedimentation. The deposits of the latest phase, coinciding with Varangu time, were of wider distribution than the underlying sandy deposits (Fig. 136D). The distribution pattern of the pre-Arenig bedrock (Fig. 137) was shaped by the post-Tremadoc erosion.

The lithofacies development of the Tremadoc - Arenig boundary indicates a transition from clastogenic to carbonate sedimentation, accompanied by intensive glauconite formation.

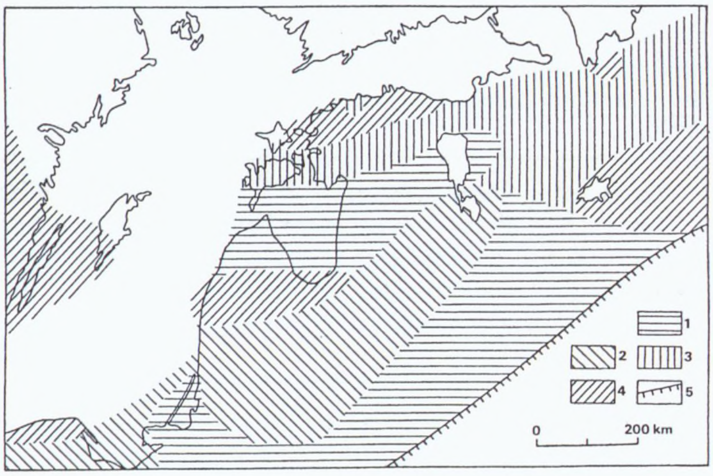


Fig. 137. Sketch-map of the pre-Arenig bedrock in Estonia and surrounding areas. Rocks of: 1 - Middle Cambrian; 2 - Upper Cambrian; 3 - Lower Tremadoc; 4 - Upper Tremadoc; 5 - boundary of the present-day distribution of the Upper Cambrian - Tremadoc rocks.

During the Ordovician and Silurian from the Arenig to the end of the Přidoli and even at the very beginning of the Devonian (Lochkovian), Estonia was part of the northern flank of a shallow cratonic sea in which carbonate and fine-clastic sediments accumulated (Fig. 138). In the earlier stages of development this sea extended from Norway to the Volga area, and from the Finnish to the Belarussian-Mazurian Precambrian massif. During the final stages of development, the basin was restricted to the Baltic Syneclise in the East Baltic area and North Poland. The nuclear part of the basin in the Baltic area is treated here as the Palaeobaltic Basin s. s.

Männil (Мянниль 1966) published a monograph on the development of the Baltic (Baltoscandian) Basin (s.l.) in the Ordovician. Kaljo and others composed a set of Silurian lithofacies maps for Estonia (Кальо и др. 1970) and for the East Baltic area (Кальо и Юргенсон 1977). Nestor and Einasto (Нестор и Эйнасто 1977) created a facies-sedimentary model for the Silurian Palaeobaltic Basin, further developed by Einasto (Эйнасто 1986). A brief summary on the basin development and facies models was published by Nestor (1990a).

Tectonical and facies settings

Two main structural elements of the Baltic Ordovician Basin (*s.s.*) were defined by Männil (Мянниль 1966): 1) the marginal or Estonian-Lithuanian Confacies Belt, 2) the central or Swedish-Latvian Confacies Belt (= Livonian Tongue, Fig. 138). A transitional zone between the above-mentioned belts has also been distinguished. The first belt roughly corresponds to the southern slope of the Fennoscandian Shield and to the northwestern slope of the Belarussian-Mazurian Anteclise (Massif), the second belt - to the Baltic Syneclise. The marginal confacies belt was dominated by shallow-water carbonate sediments with a lot of discontinuity surfaces, while the relatively deeper-water central belt comprised predominantly clayey sediments. The present-day Estonia is situated mostly within the marginal confacies belt and transitional zone. Typical facies of the central belt reach only the southernmost part of Estonia. During the basin development, the limit of the main confacies belts gradually shifted southwestwards.

In the Early and Middle Ordovician, the western part of the East-European Platform as far as the Moscow Syneclise was slowly subsiding and covered by a shallow, epicontinental sea with a comparatively weak bathymetric differentiation and extremely slow sedimentation rate. At the end of the Ordovician, since the Late Caradoc and especially during the Silurian, the upheaval of the northwestern margin of the craton got dominance in connection with the closing of the Iapetus Ocean. At the same time, on the southwestern margin of the craton, belonging to the sphere of influence of the Palaeo-Tethys, the subsidence of the basin floor intensified. As a result of different tectonic movements, a comparatively deep, "starved" of sediments, intracratonic basin depression was formed within the limits of the central (axial) confacies belt

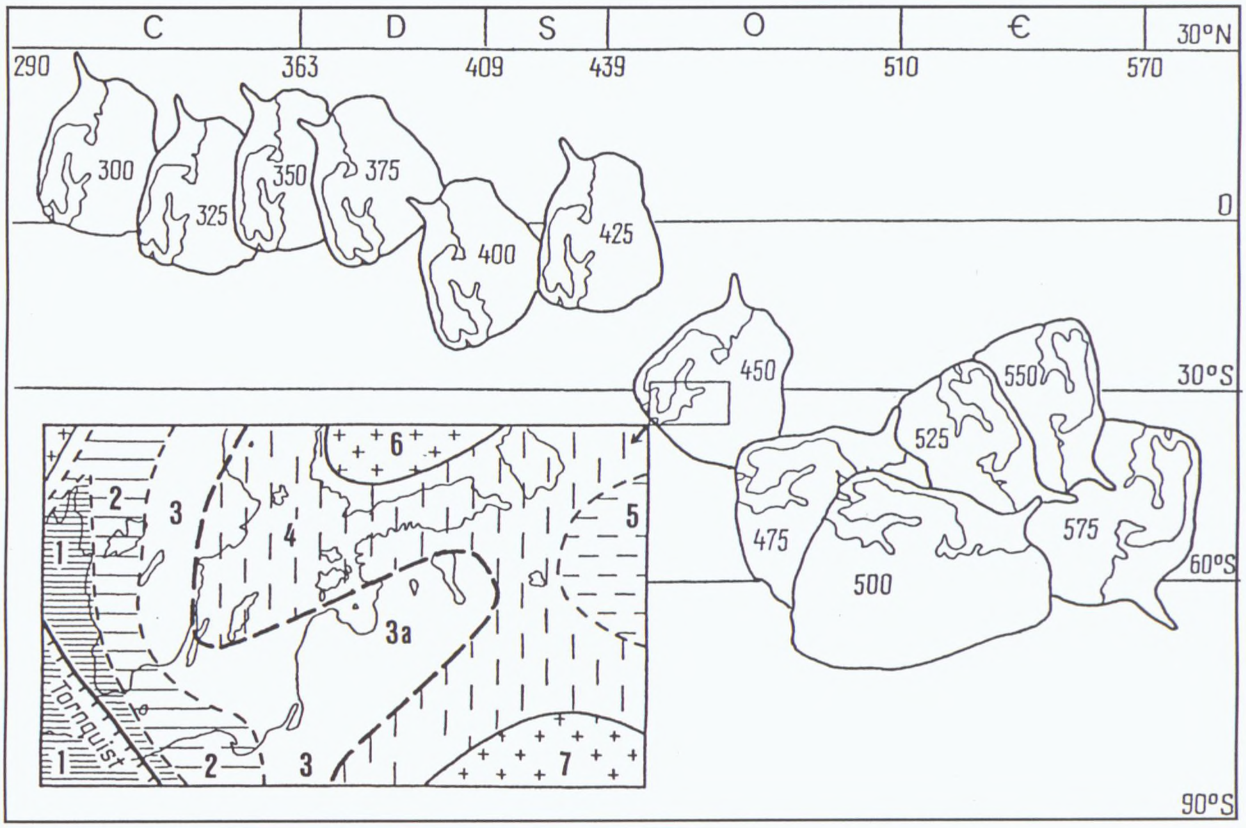


Fig. 138. The main structural elements of the Baltoscandian Ordovician and Silurian basin (after Männil 1966 and Jaanusson 1973a) and drift of the Baltica Continent during the Palaeozoic Era (after Torsvik *et al.* 1992): 1 - Skanian Confacies Belt; 2-3 - central or Swedish-Latvian Confacies Belt including Livonian Tongue or Baltic Syneclise (3a); 4 - marginal or Estonian-Lithuanian Confacies Belt; 5 - Moscow Syneclise; 6 - Fennoscandian Shield; 7 - Belarussian-Mazurian Anteclise (Massif).

Table 27. Facies belts and environments in the East Baltic Silurian Basin (modified after Нестор и Эйнасто **1977)**

Main geomorphic units	Shelf (Carbonate platform)			Deeper basin	
Type of sedimentation	Carbonate			Fine - terrigenous	
Facies belts	I Tidal flat/lagoon	II Shoal	III Open shelf	IV Transition	V Depression
Lithogenetic types of sediments	Dolomicrites	Sparitic calcarenites	Micritic calcarenites	Calcareous mudstones	Mudstones
Hydrodynamic zones	Near-shore quiet-water	Agitated-water (high-energy, turbulent)	Quiet-water to storm-agitated (subturbulent)	Calm-water	Stagnant-water
Depositional environments	Tidal (mud) - flats, lagoons, restricted shelf	shoals, reef belts, banks	Open shelf (platform)	Basin slope, shelf/basin transition	Basin depression
Characteristic rocks	Argillaceous dolomites and domerites (massive, laminated, bioturbated); dolomitic limestones, bioturbated marls	Skeletal, oolitic oncolitic pelletal grainstones, coquinoid bio- and lithoclastic rudstones, boundstones	Nodular skeletal pack- and wackstones; skeletal calcareous marls; micritic limestones	Argillaceous marls; calcareous mudstones, argillaceous micritic limestones	Mudstones, argillites, clays
Characteristic fossils	Burrows Stromatolites Eurypterids Agnathans Leperditids Scolecodonts Gastropods Lingulids	Stromatoporoids Corals Pelmatozoans Calcareous algae Oncolites Brachiopods Conodonts Vertebrates Bryozoans Bivalves	Brachiopods Pelmatozoans Ostracodes Burrows Corals Stromatoporoids Bryozoans Chitinozoans Conodonts Molluscs	Trilobites Pelmatozoans Burrows Chitinozoans Ostracodes Brachiopods Molluscs Conodonts Graptolites	Graptolites Chitinozoans Cephalopods Burrows Lingulids

in western Latvia, western Lithuania, the Kaliningrad District and northern Poland where hemipelagic argillaceous deposits accumulated. At the same time, the sea gradually retreated from the northwestern and central parts of the East-European Platform and the basin evolved from an epicontinental to a gulf-like pericontinental sea. During the Silurian, the influx of the fine-clastic material progressed from the Scandinavian Caledonides and it gradually infilled the starved depression.

During the Ordovician and Silurian, drastical climatic changes took place. The Baltica Craton drifted from the southern high latitudes to the tropical realm (see Scotese & McKerrow 1990, Torsvik *et al.* 1992, *etc.*). It induced the growth of the sedimentation rate of carbonates and development of such types of deposits which are characteristic of the arid and tropical climate (*e.g.* sedimentary dolostones, pelletal and oolitic deposits, coral-stromatoporoid reefs, *etc.*). These characteristics were totally lacking during the Early and Middle Ordovician when the Baltic Basin was situated definitely in the temperate climate zone (Jaanusson 1973a). The first signs of warm climate (appearance of tabulate corals, stromatoporoids and reefs) became evident in the Middle Caradoc (Oandu Age), but it was not until the very end of the Ordovician (Porkuni Age) that they got prevalence.

In a sedimentary basin, the facies distribution may be generalized by means of facies models presenting a lateral succession of facies along the reconstructed bathymetric profile. According to the facies model worked out by Nestor and Einasto (Нестор и Эйнасто 1977, *etc.*), five main facies belts can be differentiated in the Baltic Silurian Basin: tidal flat/lagoonal, shoal, open shelf, transitional (basin slope) and a basin depression (Table 27). The first three facies belts formed a carbonate shelf or carbonate platform and the latter two - a deeper pericratonic basin with fine-clastic deposits.

Later on, Einasto (Эйнасто 1986) elaborated two modifications of the basic Silurian facies model: one for the periods of weak and another for the periods of intense supply of terrigenous material. In the first case, the pure lime muds were widespread all over the open shelf and transitional belt; in the second case, argillaceous muds diluted carbonate sedimentation on the open shelf and periodically even in the shoal belt. It also became evident that the Silurian models are not applicable to the Early and Middle Ordovician which climatically and tectonically differed considerably from the Late Ordovician and Silurian. During the Early and Middle Ordovician, sediments accumulated under moderate climatic conditions. The main source of the carbonate component was skeletal material and its production was extremely slow. Therefore, the amount of loose skeletal material on the sea floor was limited. On the other hand, the effect of waves on the bottom was weak, because they subsided gradually in the shallow epeiric sea and lost their energy before reaching the shore. Due to these two circumstances, presumably there did not form notable accumulations of winnowed skeletal sands (grainstones) characteristic of the shoal belt of the basic Silurian facies model. Typical lagoonal carbonate sediments were also lacking, because evaporation was very weak. The position of the lagoonal and shoal facies belts was probably occupied by the belt of nondeposition, represented by a hardground (discontinuity surface). As mentioned above, in the Early and Middle Ordovician, a deeper, sediment-starved axial depression was not developed in the Baltic Basin, and the basin floor formed an evenly and weakly tilted ramp. Accordingly, only three facies belts: a nondepositional belt, an upper and a lower ramp were distinguished in the Early and Middle Ordovician facies models of the Baltic Basin (Nestor 1990a, Кыртс и др. 1991). In the development of the Baltic post-Tremadoc Ordovician and Silurian basin, five stages were differentiated (Fig. 139):

1. The transgression stage (Arenig - Llanvirn). In the marginal part of the basin the deposition was very slow and with many breaks. Within the limits of the upper ramp, micritic skeletal calcarenites accumulated, sometimes containing silt, scattered pebbles and abundant glauconite grains, goethite and francolite ooids and impregnated hardgrounds. On the lower ramp, coinciding with the central confacies belt, mainly red-coloured calcareous-argillaceous deposits (argillaceous limestones and marls) were formed. The deposits of the lower ramp were 2 to 10 times as thick as those in the coeval upper ramp.

2. The unification stage (Llandeilo - Early Caradoc). Along the whole extent of the bathymetric profile (ramp) grey calcareous - argillaceous sediments (argillaceous limestones and marls) accumulated although the general trend towards an increasing clay component and decreasing content of bioclasts in the offshore direction remained. Deposits of the upper ramp contained an admixture of light-brown kukersite kerogen and also pure kukersite interlayers. Interbeds of volcanic ash (metabentonite) were also characteristic of that stage of the basin evolution.

3. The differentiation stage (Late Caradoc - Middle Llandovery). A deeper axial depression of the basin was formed and facies zonation, typical of the Silurian developed. The supply of the basin with clastic material was periodically extremely low, and comparatively pure calcareous muds were deposited during those periods on the open shelf and in the transitional belt, while condensed dark graptolitic shales formed in the basin depression. Also an agitated-water shoal belt with pelletal and skeletal sands and a lagoonal belt with dolomicritic deposits (dolomitic marls, argillaceous dolomites) developed. During this stage, the periods of low terrigenous mud influx cyclically alternated with the phases of more intensive supply with the fine clastic material. In the latter case, deposition followed the basic Silurian facies model characteristic of the stage of stabilization of the basin development.

4. The stabilization stage (Late Llandovery - Early Ludlow). Moderate influx of the fine-clastic material, which partly deposited in the lagoonal and open shelf belts but mostly in the transition belt, resulting in side-filling of the "starved" depression with deposit wedges (lenses) and gradual progradation of the carbonate shelf margin was characteristic of this stage. Facies zonation was clear and the basic Silurian facies model (Table 27, Fig. 139d) reflects the situation during that stage of evolution. However, the facies belts were not equally developed during the different phases of the basin development. The deeper-water facies were widespread during the transgressive phases (the end of the Llandovery and the beginning of the Wenlock), while shallow, marginal-marine facies were best developed during the regressive phases (the end of the Wenlock and the beginning of the Ludlow) of the basin development.

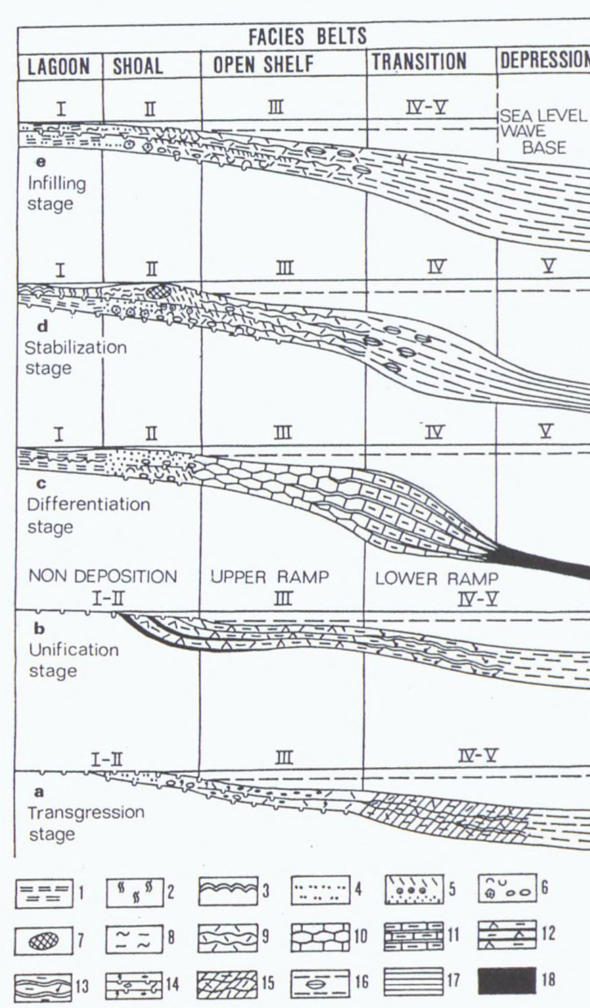


Fig. 139. Facies models for different stages of development of the post-Tremadoc Ordovician and Silurian Palaeobaltic Basin. Rock types: 1 - laminated argillaceous dolomites; 2 - bioturbated argillaceous dolomites; 3 - stratiform stromatolites; 4 - terrigenous silt- and sandstones; 5 - skeletal, oolitic and pelletal grainstones; 6 - coquinoid, oolitic and oncolitic rud- and floatstones; 7 - bioherms, carbonate mounds; 8 - bioturbated clayey wackestones; 9 - nodular biomicritic limestones (skeletal pack- and wackestones); 10 - wave-bedded micritic limestones; 11 - interbedded clayey micritic limestones and kerogenous graptolitic mudstones; 12 - biomicritic limestones with kukersite interlayers; 13 - interbedded argillaceous limestones and marls; 14 - limestones with glauconite, ferruginous ooids, lithoclasts and hardgrounds; 15 - purple argillaceous limestones; 16 - marlstones with limestone nodules; 17 - grey graptolitic mudstones; 18 - dark-brown kerogenous graptolitic shales.

5. The infilling stage (Late Ludlow - Pridoli). Intense influx of terrigenous material filled the basin depression and also diluted carbonate sedimentation on the open shelf where bioclastic limestones were replaced by bioclastic marls. Even in the shoal belt skeletal sands interlayered with marls.

Generally, certain facies models were typical only of certain stages of basin evolution, however, in some cases different types of sedimentation could also alternate. For example, during the Late Caradoc to the Middle Landovery, the periods of low and moderate supply with terrigenous clastic material alternated cyclically and sedimentation followed the models charcateristic of the differentiation and stabilization stages, respectively.

In the following, the evolution history of the basin will be presented with an emphasis on the situation in Estonia. A distinct cyclicity in the basin evolution was summarized by Einasto (1995) and is figured in the present work (Fig. 140). Nine high-rank macrocycles of eustatic origin have been established in the evolution of the Baltic Ordovician and Silurian Basin. They are separated from each other by subregional sedimentation breaks of different duration, increasing in onshore direction. Lower-rank, meso- and microcycles have also been distinguished which, besides the fluctuation of the sea level, were also induced by climatic cyclicity and pulsatory supply with terrigenous material. A meridional cross section of the Ordovician and Silurian rocks across Hiiumaa and Saaremaa islands and the Kuramaa Peninsula (Fig. 141) shows spatial and temporal relations of the main facies commented in the text below.

Transgression stage of development

This, the first stage in the evolution of the Baltic carbonate sedimentation basin coincided roughly with the Arenig and Llanvirn (*s.s.*), *i.e.* with the time interval from the Hunneberg to Lasnamägi ages. However, the upper limit of the stage was quite transitional and felt in the middle of a high-rank eustatic macrocycle (see Fig. 140). It is defined by the disappearance of goethide and francolite ooids in the marginal confacies belt and marine red-beds in the central confacies belt, by the appearance of kukersite kerogen and by an abrupt increase in the content of terrigenous clay material, characteristic of the next - unification stage of development.

At the beginning of the transgression, the general structural setting of the post-Tremadoc Ordovician deposition basin developed. Two main facies (sedimentation) belts: an upper (inner) and a lower (outer) ramp were formed within the Baltic Basin *sensu stricto* (Fig. 139a). The former coincides with the marginal or Estonian-Lithuanian Confacies Belt by Männil (Мяннилъ 1966), the latter corresponds to the central, Swedish-Latvian Confacies Belt (Мяннилъ 1966) or the Livonian Tongue by Jaanusson (1973a).

The marginal, upper ramp deposits were represented by comparatively pure bioclastic carbonate muds with an admixture of glauconite in the lower and goethite or francolite ooids (pseudo-ooids) in the upper part of the sequence. A lot of discontinuity surfaces (hardgrounds), corresponding to shorter or longer sedimentation breaks, but also scattered lithoclasts and skeletal grainstone interlayers are characteristic of the deposits of the most peripheral part of the basin. In the deeper-water, central confacies belt (Livonian Tongue) much thicker deposits of purple calcareous-argillaceous muds with rare discontinuities accumulated. These were treated as the deposits of lower (outer) ramp. Outside the Baltic area, on the western margin of the East-European Platform (Scanian Confacies Belt), the purple calcareous argillaceous muds were replaced by the deepest-water black kerogenous graptolitic muds (see Мяннилъ 1966, Figs. 50-55).

The stage of transgression may be subdivided into two substages of development. The first substage roughly corresponds to the Arenig (Hunneberg to Volkhov ages) and represents a full eustatic macrocycle (Fig. 140). The second, Llanvirn (*s.s.*) substage (Kunda to Lasnamägi ages) embraced only half of the next – Llanvirn - earliest Caradoc macrocycle.

The **Arenig macrocycle (substage)** began with a well-known world-wide transgression. In the marginal confacies belt it started with the deposition of a very thin band of glauconitic sands and silts of the Leetse Formation (Hunneberg and Billingen regional stages). In the central confacies belt (Livonian Tongue) comparatively thick deposit of purple to greenish-grey terrigenous muds (Zebre Formation) accumulated. In the course of progressive transgression, the bulk of the carbonate component gradually increased and the area of all facies belts extended remarkably.

During the Volkhov Age, glauconitic, bioclastic calcareous muds (Toila Formation) accumulated in the marginal part of the basin. There were numerous interruptions in sedimentation marked by impregnated hardgrounds. At the same time, in the central belt of the basin the purple calcareous-terrigenous muds (Kriukai Formation) expanded their area. A wide transitional zone, including all of southern Estonia, with variegated bioclastic calcareous-argillaceous muds was developed between the marginal and central confacies belts (Fig. 142).

The Arenig macrocycle ended with an extensive eustatic regression as a result of which the sea receded from the whole upper ramp area and a karstified erosional surface with deep corroded pockets and furrows developed on top of the Volkhov Stage. At the time of the regression maximum, in the central confacies belt the purple calcareous argillaceous muds were replaced by grey calcareous muds (Šakina Formation).

The second, **Llanvirn substage** of basin development roughly corresponded to the Llanvirn *sensu stricto* (Kunda, Aseri and Lasnamägi ages). In the central belt of the basin, the deposition of purple and variegated argillaceous-calcareous muds (Baldone, Segerstad, Stirnas formations) continued. The deposits being represented by argillaceous and biomicritic limestones became only more calcareous than the earlier ones. In the marginal part of the basin, deposition of bioclastic calcareous muds was ongoing, but instead of glauconite, goethite and francolite ooids were distributed on certain levels of the Loobu, Kandle and Väo formations. The Early Ordovician transgression reached its maximum spatial extent during the Kunda Age (Saadre 1992). However, a slow deepening of the basin, especially in its peripheral part, continued until the end of the stage of transgression and during the next, unification stage as well (Fig. 140).

Unification stage of development

This stage in the basin development covered the Llandeilo and Early Caradoc interval from the Uhaku to Keila ages. It was a period of relative tectonic and eustatic stillstand on the

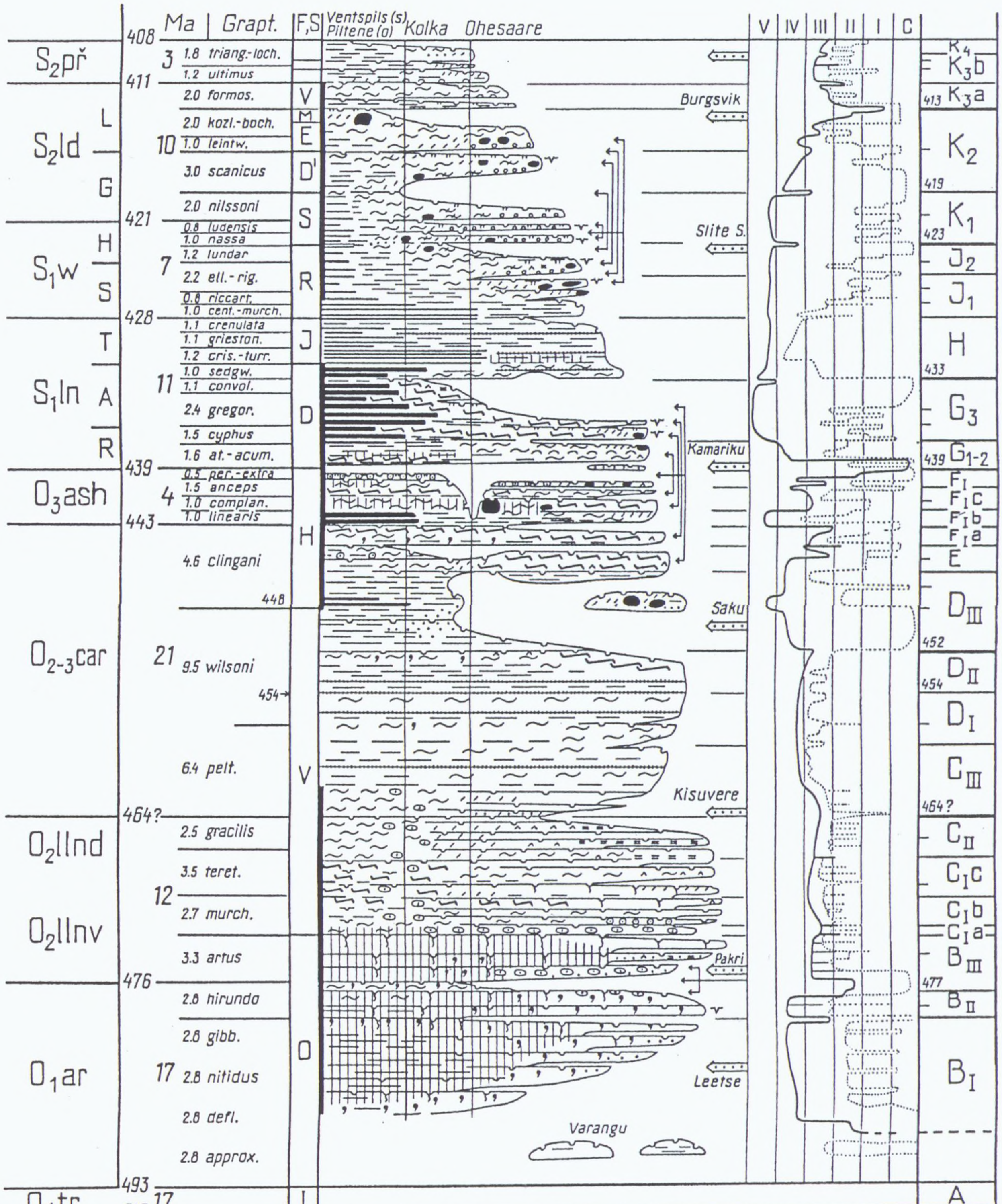


Fig. 140. A generalized sequence of the Ordovician and Silurian rocks in Estonia (by Einasto 1995) shown in time-scale (after Harland *et al.* 1990 a.o.) with eight higher rank stratigraphical breaks, separating nine eustatic macrocycles of basin development. On the right, two sea-level curves are shown - a dotted line for the marginal and a solid line for the central confacies belts. Facies belts: I - lagoonal, II - shoal, III - open shelf, IV - transitional, V - depression (see Table 27). For lithofacies legend see Fig. 141. Stratigraphical indices: *Regional Series*: I - Iru, O - Ontika, V - Viru, H - Harju; *Silurian Formations*: D - Dobele, J - Jurmala, R - Rìga, S - Siesartis, D - Dubysa, E - Engure, M - Mituva, V - Ventspils; *Silurian Standard Stages*: R - Rhuddanian, A - Aeronian, T - Telychian, S - Sheinwoodian, H - Homerian, G - Gorstian, L - Ludfordian; *Baltic Regional Stages*: B_I - Latorp, B_{II} -Volkhov, B_{III} - Kunda, C_Ia - Aseri, C_Ib - Lasnamägi, C_Ic - Uhaku, C_{II} - Kukruse, C_{III} - Idavere, D_I - Jõhvi, D_{II} - Keila, D_{III} - Oandu, E - Rakvere, F_Ia - Nabala, F_Ib - Vormsi, F_Ic - Pirgu, F_{II} - Porkuni, G_{1-2} - Juuru, G_3 - Raikküla, H - Adavere, J_1 - Jaani, J_2 - Jaagarahu, K_1 - Rootsiküla, K_2 - Paadla, K_3a - Kuressaare, K_3b - Kaugatuma, K_4 - Ohesaare.

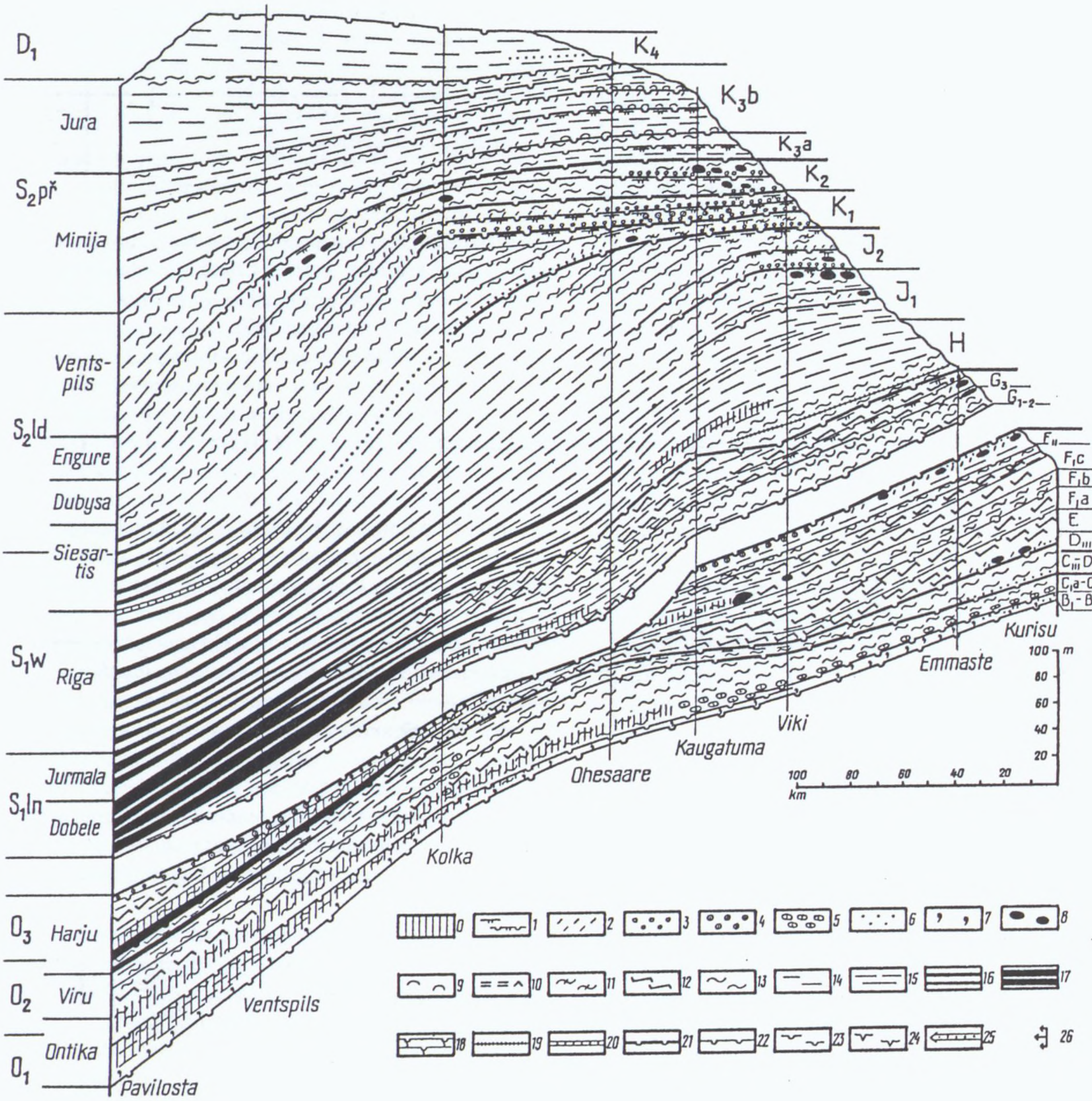


Fig. 141. A meridional cross section throughout the Ordovician and Silurian lithofacies in Hiiumaa and Saaremaa islands and the Kurzeme Peninsula by Einasto (1995), showing drastical differences in thicknesses and sedimentation rates between the Ordovician and Silurian. Lithological legend: 0 - marine redbeds; 1 - lagoonal dolomitic mud; 2 - skeletal sand; 3 - oncolitic and coquinoid gravel; 4 - calcareous ooids; 5 - ferrugenous ooids in calcareous matrix; 6 - terrigenous silt and sand; 7 - glauconitic deposits; 8 - bioherms and carbonate mounds; 9 - brachiopod and bivalve coquina; 10 - kukersine interlayers; 11 - argillaceous-calcareous muds of restricted shelf; 12 - pure lime muds of restricted and open shelf; 13 - bioclastic calcareous muds of open shelf; 14 - argillaceous-calcareous muds at shelf margin; 15 - calcareous-argillaceous muds of transitional belt; 16 - grey graptolitic muds; 17 - dark kerogenous graptolitic muds; 18 - numerous hardgrounds; 19 - metabentonites; 20 - laminated calcareous-argillaceous muds of basin depression; 21 - erosional surfaces and disconformities; 22 - significant discontinuities; 23 - brokened hardgrounds; 24 - mud-cracks, 25 - levels of silt and sand influx; 26 - symmetrical facies successions. Explanation of stratigraphical indices see Fig.140.

East-European Platform which terminated the general transgressive phase in the development of the Ordovician basin. The most characteristic features of the stage were remarkable intensification of the influx of fine terrigenous material, prevailingly from the northeast, and considerable equilization of environmental conditions in the marginal and central parts of the Baltic Basin (s.s.). All over the Baltic area mainly bioclastic argillaceous-calcareous muds accumulated. Besides lateral variation in the ratio of argillaceous and calcareous components, also a distinct lower rank cyclicity, expressed in alternation of more and less argillaceous deposits, was apparent. The latter reflects the pulsatory supply of the basin with fine terrigenous material rather than the sea-level fluctuation.

Characteristic of the earlier part of the unification stage (Uhaku and Kukruse ages) was accumulation of light-brown organic matter - kukersine which formed kukersite interlayers and dispersed admixture in calcareous sediments. During the second half of the unification stage (Haljala and Keila ages), numerous volcanic ash (metabentonite) intercalations formed giving evidence of growing volcanic activity in the adjacent Iapetus Ocean, perhaps connected with its transition from the opening phase to the closing state. It is remarkable that the

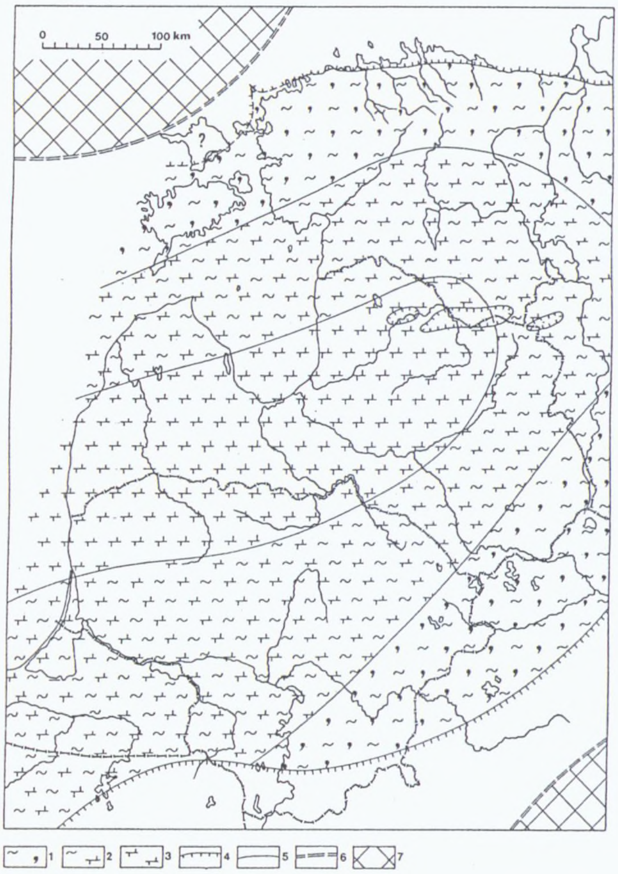


Fig. 142. Distribution of the Early Ordovician Volkhov Age sediments and facies belts in the Baltic Basin: 1 - grey glauconitic, bioclastic calcareous mud; 2 - variegated calcareous-argillaceous mud with bioclasts; 3 - purple calcareous- argillaceous mud; 4 - boundary of the present extension of rocks; 5 - facies boundaries; 6 - shoreline; 7 - presumable land.

same phenomenon, *i.e.* the presence of numerous metabentonite interlayers, is also characteristic of the highly argillaceous Late Llandovery and Early Wenlock sediments which formed during the Silurian transgression maximum. The above-mentioned two parts may be treated as substages of the unification stage.

The **Llandeilo - earliest Caradoc substage** of development (Uhaku and Kukruse ages) was a specific period of accumulation of kerogenous, kukersine-bearing deposits in northeastern Estonia and in its vicinity (Fig. 143). The depocenter of these deposits was situated near the present-day outcrop belt. It is supposed (Мяннилъ и др. 1986, Bauert 1989, Кыртс и др. 1991) that most of the kukersine was allochtonous and derived from algal-microbial mats, covering the flat sea-shore (tidal flat), from where it was transported into a broad near-shore sinking of the sedimentation basin. In rest of the marginal confacies belt and transition area, monotonous bioclastic argillaceous-calcareous muds (Kõrgekallas and Dreimani formations) were deposited (Fig.143). In the sediments of a wide transition zone (Dreimani Formation), the skeletal particles are heavily pyritized and scattered goethite ooids occur. In the axial part of the Livonian Tongue, the bioclastic muds became more argillaceous, but other characteristics remained almost the same. In northern Estonia, the Llandeilo - earliest Caradoc evolution substage terminated with a short erosional sedimentation break at the end of the Kukruse Age (Fig. 140).

The **Early - Middle Caradoc substage** of development (Haljala and Keila ages) was a complete eustatic macrocycle with a short transgressive phase at the beginning, longer stillstand period in the middle and drastical shallowing and regression at the end. It began in northern Estonia with the deposition of a thin basal layer of calcareous silt (Kisuvere Member of the Kahula Formation). The silt probably derived from the Kärdla impact crater (Пылма 1982) and was transported eastwards by longshore drift. The macrocycle in consideration is characterised by the occurrence of the most argillaceous sediments in the whole Ordovician sequence and by the presence of numerous metabentonite interlayers. Only at the beginning (Tatruse Formation) and at the end of the macrocycle (Pääsküla and Saue members of the Kahula Formation), purer bioclastic calcareous sediments occurred in northern Estonia, in the marginal confacies belt. During the middle part of the macrocycle, bioclastic calcareous-argillaceous muds accumulated in the marginal confacies belt, while in the central belt highly argillaceous bioclastic muds (Adze Formation) deposited.

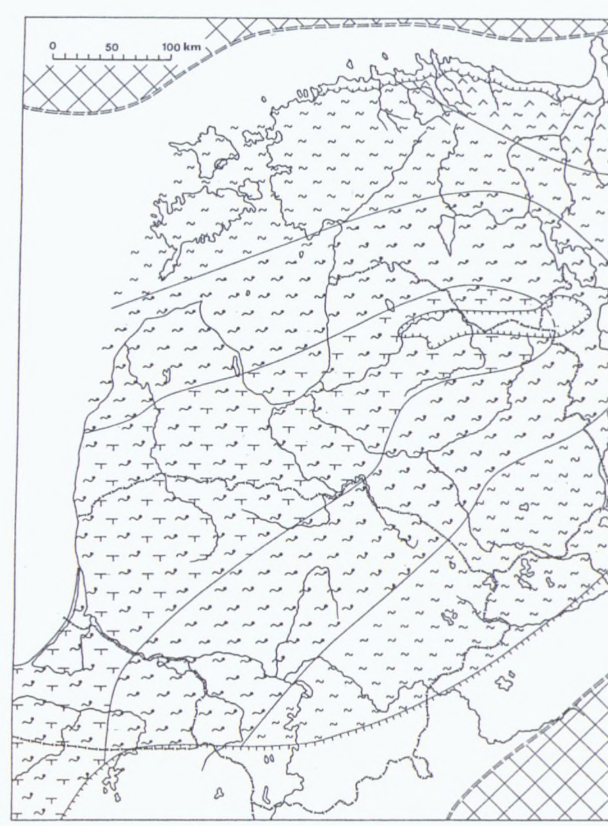


Fig. 143. Distribution of the Middle Ordovician Kukruse Age sediments and facies belts in the Baltic Basin: 1 - bioclastic calcareous mud; 2 - bioclastic calcareous mud with kukersine interlayers; 3 - bioclastic argillaceous-calcareous mud with pyritized skeleton particles; 4 - interbedded bioclastic argillaceous-calcareous and calcareous-argillaceous muds; 5 - presumable land; 6 - boundary of the present extension of rocks; 7 - facies boundaries; 8 - shoreline.

The macrocycle and the whole unification stage ended with remarkable shallowing and regression. In northwestern Estonia, in the Vasalemma area at late Keila and early Oandu times, a reef complex with bryozoan-microbial carbonate mounds and intermound pelmatozoan grainstones was formed in the shallow-water high-energy environment. Elsewhere in northern and central Estonia, a probable hiatus existed between the Keila and Oandu stages. At the same time, highly argillaceous muds of the Blidene Formation (marls and mudstones) formed in the axial part of the Livonian Tongue (Fig. 144). In the wide transitional zone, they were replaced by slightly more calcareous muds which in their most peripheral part in southwestern Estonia, easternmost Latvia and Lithuania contain fine siliciclastic material (see Ainsaar 1995). The latter deposited at the time of regression maximum when the sea supposedly withdrew from the whole upper ramp area, including northern Estonia.

Differentiation stage of development

This stage of evolution corresponds to the Late Ordovician and earliest Silurian covering the time interval from the Oandu Age up to the end of the Raikküla Age. The beginning of the evolution stage roughly coincided with the general tectonic inversion in the western part of the East-European Platform, *i.e.* transition from the transgressive to regressive phase in basin development. On the other hand, it also reflected transition from humid, moderate climatic conditions to arid subtropical-tropical climatic conditions. Evidence is derived from the appearance of the first corals and stromatoporoids, beginning of the formation of organic build-ups, pelletal, aragonitic sediments, extensive accumulation of pure lime muds, *etc.*

The most characteristic features of this evolution stage included: 1) development of considerable lateral lithological differentiation of sediments, 2) rapid increase in the sedimentation rate and thickness of deposits, 3) formation of condensed deposits of dark kerogenous muds (shales) in the central confacies belt of the basin, cyclically interbedded with thicker deposits of greenish-grey to purple muds (mudstones and marls), 4) cyclical alternation of deposits of pure lime muds (micritic limestones) with bodies of argillaceous lime muds (marls, argillaceous limestones) in the marginal part of the basin and transitional zone. The former three peculiarities were probably connected with transformation of the western margin of the Baltica Continent from passive to active state due to the beginning of the gradual closure of the Iapetus Ocean. It caused different tectonic movements in the Baltic Syneclise and its surroundings; as a result, deeper intracratonic depression began to form within its limits. A possible reason for cyclical alternation of the deposition of pure and argillaceous lime muds may be an interchange of the arid and humid climate periods characteristic of the big glacial epochs in the Earth history, including the Late Ordovician - Early Silurian glacial epoch. The arid periods caused a relatively low influx of terrigenous material and, consequently, deposition of purer lime muds, while during the humid periods more argillaceous muds were deposited.

In the evolution of the Baltic Basin, the cyclically alternating periods of low and high influx of terrigenous material are well recognizable. During the low influx ("limy") phases rather pure, light lime muds (micritic or calcilutitic limestones) deposited on the open shelf and in the transitional facies belt (see Fig. 139c). During the high influx ("clayey") phases, on the open shelf bioclastic argillaceous-calcareous muds (nodular argillaceous biomicritic limestones) were deposited, while in the transitional belt prevailingly terrigenous muds (marlstones, mudstones) accumulated (Fig. 139d). Nine of such cycles of alternating low- and high-influx phases have been recognized in the development of the Baltic Basin during Late Ordovician - earliest Silurian times.

The low-influx ("limy") phases were: 1) Rakvere (Rägavere Formation), 2) late Nabala (Saunja Formation), 3) early Pirgu (Moe and Svėdasai formations), 4) mid-Pirgu (Oostriku, Baltinava, Parovėja formations), ?5) latest Pirgu (Taučionys Formation), 6) early Juuru (Koigi, Ruja, Sturi members), 7) early Raikküla (Järva-Jaani beds, Slītere Member), 8) mid-Raikküla (Jõgeva beds, l. pt., Ikla Member), 9) late Raikküla (Mõhküla beds, Staicele Member).

The high-influx ("clayey") phases were correspondingly: 1) Oandu (Hirmuse, Lukštai formations), 2) early Nabala (Saunja, Mõntu formations), 3) Vormsi (Kõrgessaare, Tudulinna, Meilūnai formations), 4) early-mid Pirgu (Adila, Halliku, Ukmerge formations), 5) mid-late Pirgu (Ludza, Kuiļi formations), 6) Porkuni (Ärina, Kuldiga, Saldus formations), 7) middle and late Juuru (Varbola, Tamsalu formations, Rozeni Member), 8) early-mid Raikküla (Vändra beds, Kolka Member), 9) mid-late Raikküla (Jõgeva beds, u. pt., Lemme Member).

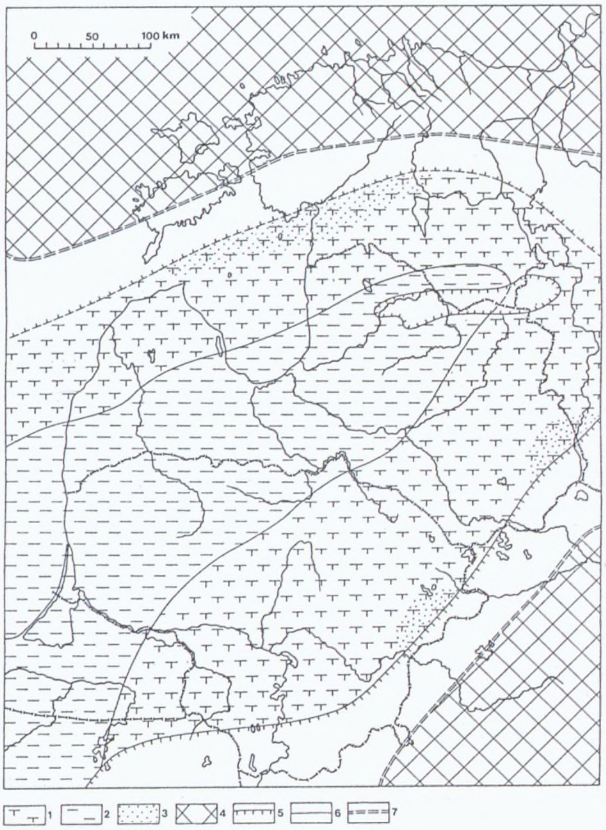


Fig. 144. Distribution of the late Middle Ordovician late Keila (top Pääsküla) sediments and facies belts in the Baltic Basin: 1 - calcareous-argillaceous mud; 2 - terrigenous mud; 3 - admixture of silt; 4 - presumable land; 5 - boundary of the present extention of rocks; 6 - facies boundaries; 7 - shoreline.

In most cases, the distribution of the pure lime muds was restricted to the open shelf and transitional facies belt (see Fig. 138c). Basinwards the lime muds were replaced by dark brown kerogenous graptolitic muds (Mossen Formation of the Rakvere Stage, Dobele Formation of the Raikküla Stage) or by purple calcareous-argillaceous muds (Jonstorp Formation of the Pirgu Stage, Remte Formation of the Raikküla Stage). However, in some cases light lime muds covered also extensive areas in the central part of the basin (Saunja Formation of the Nabala Stage, Parovėja Formation of the Pirgu Stage, Stūri Member of the Juuru Stage); more deeper-water deposits are not represented in the Baltic area. Lateral transition of the open shelf lime muds into the deposits of the near-shore agitated-water shoal belt has been established only in the Raikküla Age (Figs. 144 and 66). In this case, they were gradually replaced by fine-grained skeletal-pelletal deposits, in places containing numerous corals and stromatoporoids.

The deposition of the pure lime muds probably proceeded under somewhat specific hydrochemical conditions. The deposits of the micritic limestones (calcilutites) contain extremely sparse organic remains. Most frequently, the Late Ordovician micritic limestones comprise skeleton particles of specific dasycladacean algae *Cyclocrinites*, *Vermiporella*, *etc*. The micritic limestones of the Raikküla Stage often contain detritus of problematic dendroid graptolites. The presence of such specific biotic elements suggests that a certain generative role of the biochemical factor in the formation of pure lime muds is not excluded. The micritic limestones of the Pirgu Stage, especially those of the Moe Formation, are richest in skeleton particles of *Vermiporella* and other dasycladaceans. The latter and the adjacent Tootsi Formation also contain specific carbonate mounds (Hoitberg, Niiby, Ruunavere, Kaugatuma, Paatsalu, Võhma) which are analogous to the well-known Boda mounds in Central Sweden (see Nestor 1995b).

The differentiation stage of the basin development consisted of two similar bathymetric macrocycles, one corresponding to the Late Caradoc - Ashgill from the Oandu to the Porkuni stages, and another to the Early-Middle Llandovery from the Juuru to the Raikküla stages. These macrocycles began with a slow, gradual deepening of the basin and finished with a rapid shallowing, regression and, in places, with intensive erosion of the older deposits.

The **Late Caradoc-Ashgill macrocycle** began with a short transgressive episode in the Oandu Age when for the first time in the post-Tremadoc history of the basin development typical anoxic depression facies - dark-brown kerogenous graptolitic muds (Mossen Formation) were formed in the Baltic Syneclise. At the slope of the Belarussian-Mazurian Anteclise the Oandu deposits transgressively overlap the older strata and in northern Estonia they also disconformably overlie the Keila deposits.

After a rapid initial deepening during the Oandu Age, there followed a relative stabilization and levelling of sedimentation conditions culminating during late Nabala time when monotonous pure lime muds (Saunja Formation) expanded over the whole East Baltic area. A new deepening impulse and bathymetric differentiation followed in the Vormsi Age when deposition of dark kerogenous muds with graptolites (Fjäcka Formation) was restored in the central depression of the Baltic Syneclise (Fig. 145).

The Pirgu Age was a variable period in the basin evolution with two distinct episodes of accumulation of pure lime muds. A general regressive trend of basin development is recognizable in the Pirgu Age. During the first, early Pirgu phase of deposition of lime muds (Moe Formation) purple terrigenous muds (Jonstorp Formation) were deposited in the central confacies belt, while during the second, mid-Pirgu phase of lime mud deposition (Oostriku, Baltinava, Parovėja formations), monotonous pure lime muds spread all over the Baltic Basin. The Pirgu Age ended with an obvious sedimentation break of different duration in different places. Extensive local sedimentation gaps with considerable erosion of lower-lying strata have been established in the areas of the Lower Nemunas and Irbe structural elevations (see Мяннилъ 1966, Kaljo *et al.* 1988b). In these and some other places, the Porkuni rocks rest disconformably on the Pirgu strata showing that the Porkuni Age began with a certain transgression event. During the first half of the Porkuni Age a diverse complex of shallow, agitated-water carbonate sediments (skeletal sand and silt, coral-stromatoporoid bioherms, bioclastic calcareous muds of the Ärina Formation) formed in the marginal confacies belt. In the central belt bioclastic calcareous-argillaceous muds (Kuldiga Formation) deposited at the same time. An abrupt shallowing took place in the middle of the Porkuni Age and deposition of shallow-water calcarenitic sediments (bioclastic, oolitic, lithoclastic sand, silt and gravel) with remarkable admixture of siliciclastic silt and sand shifted into the central confacies belt, forming high-energy shoal deposits of the Salduse Formation (Fig. 146). In the peripheral part of the Baltic Basin, including northern Estonia, subaerial conditions existed at that time. In the mid-Estonian transitional zone some 15-to-30-m-deep erosion channels (Tootsi, Jõgeva, Ruskavere) were recently discovered (Perens 1995, Ainsaar 1995). The end-Ordovician drastical shallowing event (or events) in the Baltic Basin was definitely connected with the global glacio-eustatic drop of the ocean level (Кальо и др. 1991), perhaps combined with certain tectonical upheaval, especially at the end of the Pirgu Age.

The **Early-Middle Llandovery macrocycle** began with a glacio-eustatic rise of the sea level and deposition of pure lime muds (Stūri, Rūja, Koigi members) on wide areas of central and eastern East Baltic. Only in the deepest-water residual depression (South Estonia - Kurzeme) the lime muds were replaced by calcareous-argillaceous muds (Õhne Formation). During the Juuru Age, a bathymetric differentiation and development of regular facies zonation, interrupted by the end-Ordovician hiatus, denudation and levelling of the sea floor topography, were gradually restored, but it was not until mid-Raikküla time (Ikla/Jõgeva beds) that a deep-water central depression with dark kerogenous graptolitic muds was finally restored and since then, until middle Ludlow time, it functioned as a sediment-starved depression where deposition rate could not keep pace with the subsidence of the sea floor. A full set of five main facies belts (Table 27) was completely established and a shelf-type sedimentation finally replaced the ramp-type sedimentation which prevailed during most of the Ordovician. The most characteristic feature of the shelf-type sedimentation was formation of thick deposit wedges in the transitional facies belt which led to the side-filling of the basin depression and gradual progradation of the carbonate shelf edge. A gradual side-filling effect is well visible in the cross-section of the Llandovery rocks in western Estonia (Figs. 66, 141).

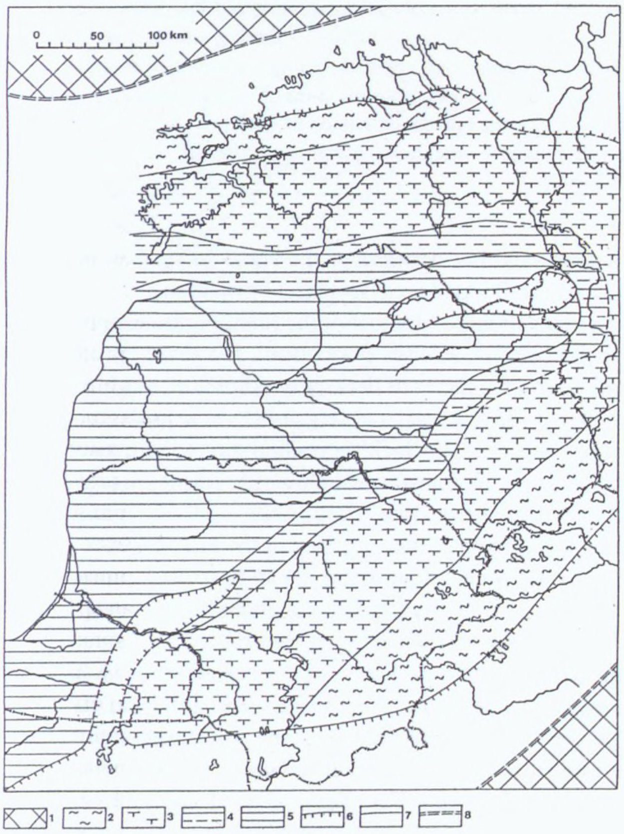


Fig. 145. Distribution of the Late Ordovician Vormsi Age sediments and facies belts: 1 - presumable land; 2 - bioclastic calcareous mud; 3 - calcareous-argillaceous mud; 4 -grey terrigenous mud; 5 - dark kerogenous argillaceous mud with graptolites; 6 - boundary of the present extension of rocks; 7 - facies boundaries; 8 - shoreline.

The Early-Middle Llandovery macrocycle ended with extensive local sedimentation breaks and denudation of earlier deposited strata in the marginal parts of the basin. One area of denudation was situated in western Estonia where the north-westwards increasing erosional hiatus cut the older strata down to the Järva-Jaani beds. Still more remarkable erosional break was developed on the opposite flank of the Baltic Basin, in eastern Lithuania, where the Early and Middle Llandovery deposits were subject to denudation all over the carbonate shelf as far as the eastern limit of the Baltic Syneclise with depression facies of dark graptolitic shales of the Dobele Formation (Fig. 147). The local (subregional) character of the nondeposition and changeable extent of deposition breaks suggest the tectonic nature of the upheaval, most probably induced by the beginning of the collision of the Laurentia and Baltica continents.

Stabilization stage of development

The stabilization stage of basin evolution embraced the main part of the history of the Baltic Silurian Basin from the beginning of the Late Llandovery (beginning of the Adavere Age) up to the end of the Middle Ludlow (end of the Paadla Age). The most characteristic features of the stabilization stage of evolution were: 1) a moderate influx of the fine terrigenous material, 2) the presence of a comparatively deep, starved axial basin depression with continuous sedimentation of dark-grey argillaceous deposits (mudstones and shales) with graptolites, 3) a perfect shelf-type facies zonation beginning from marginal-marine, lagoonal dolomitic muds and ending with dark -grey graptolitic muds in the depression belt (see Table 27 and Fig. 139d). At the beginning of the stabilization stage, during the transgressive phase of the basin development, deeper-water facies had a wide distribution. During the regressive phase, in the second half of the stabilization stage, shallow near-shore facies were well developed.

In the marginal part of the basin, a general regressive trend of evolution was characteristic to the stabilization stage and it led to the transformation of the Baltic Basin into a gulf-like pericontinental sea. On the background of the general shallowing trend, smaller-scale sea level fluctuations were characteristic of the basin evolution. The stabilization stage may be subdivided into two eustatic macrocycles of unequal duration and completeness. The first, perfect deepening-shallowing macrocycle corresponds to the Late Llandovery - Middle Wenlock (Adavere to Jaagarahu ages) and the second, uncomplete one to the Late Wenlock and Early Ludlow (Rootsiküla and Paadla ages).

The **Late Llandovery - Middle Wenlock macrocycle** began in the Baltic Basin with a rather long transgressive (deepening) phase, lasting from the beginning of the Adavere Age up to the mid-Jaani time. The transgression proceeded in

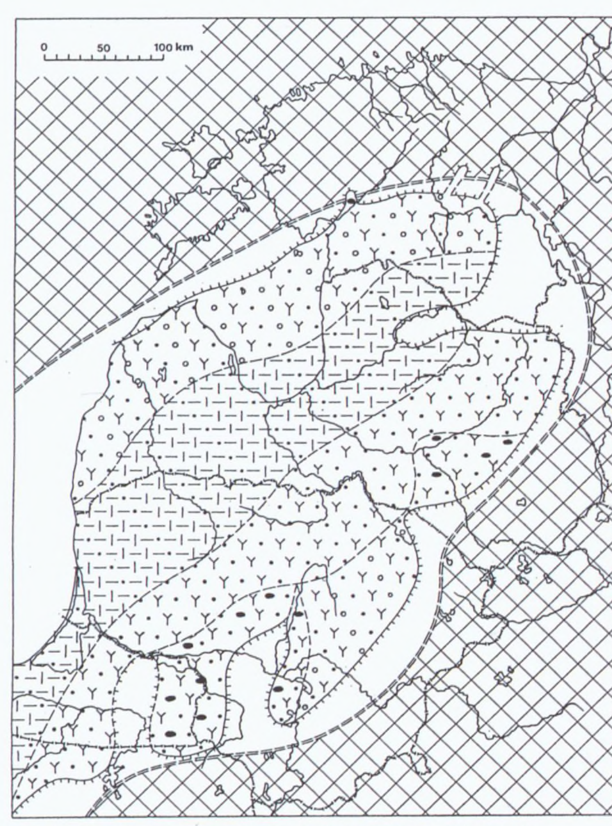


Fig. 146. Distribution of the end-Ordovician late Porkuni (Saldus Formation) sediments and facies belts: 1 - presumable land; 2 - calcareous sand with oolites; 3 - calcareous sand with siliciclastic admixture; 4 - microlayered silty calcareous and terrigenous muds; 5 - conglomerate; 6 - boundary of the present extension of rocks; 7 - boundaries of sediment types; 8 - shoreline; 9 - erosion channel.

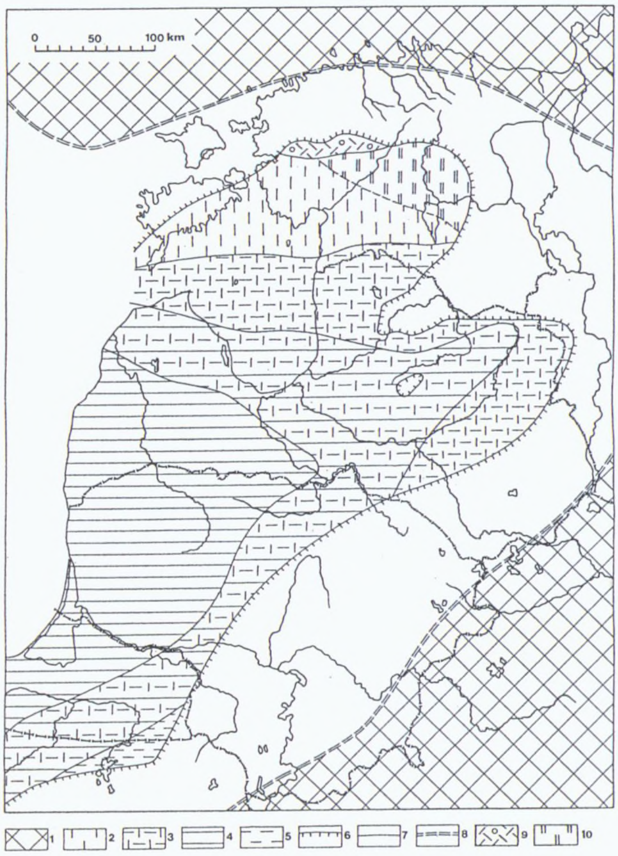


Fig. 147. Distribution of the Llandovery mid-Raikküla (Ikla Member) sediments and facies belts: 1 - land; 2 - pure lime mud; 3 - interbedded lime mud and dark kerogenous mud with graptolites; 4 - dark kerogenous graptolitic mud; 5 - grey graptolitic mud; 6 - boundary of the present extension of rocks; 7 - facies boundaries; 8 - shoreline; 9 - pelletal-skeletal sand and silt; 10 - dolomitized lime mud.

two steps (Нестор 1972). It began with an extensive deposition of bioclastic argillaceous-calcareous muds with coquinite accumulations of the brachiopod *Pentamerus oblongus* (Rumba Formation) in the the open shelf facies belt, transgressively overlying different strata of the Raikküla Age. In the central depression of the basin, condensed dark-brown kerogenous graptolitic muds of the Dobele Formation continued to deposit. During the next step of deepening, corresponding to late Adavere (Velise) time, argillaceous sediments of the depression (Jurmala Formation) and transitional facies belts (Velise and Švenčionys formations) covered the whole East Baltic area. The transgression was especially extensive in eastern Lithuania where marly deposits of the Švenčionys Formation disconformably overlie the Ordovician strata. The Late Llandovery - Early Wenlock transgression expanded even into the Moscow Syneclise (Решения... 1987). During the late Adavere time, purple terrigenous muds accumulated in the western part of the Baltic Basin (Kurzeme, Sõrve, Gotland areas).

Deposition of the deeper-water, highly argillaceous sediments continued also at the beginning of the Wenlock, Jaani Age (Tõlla beds of the Rīga Formation, Mustjala Member of the Jaani Formation, Sutkai beds of the Paprieniai Formation). The highly argillaceous sediments of the Late Llandovery and the earliest Wenlock contain numerous thin metabentonite interlayers. It leads to the supposition that the Late Llandovery - Early Wenlock sea-level high-stand was probably induced by the acceleration of the sea-floor spreading accompanied by intensification of volcanic activity in the Iapetus Ocean.

In the middle of the Jaani Age, an abrupt shallowing developed in the peripheral part of the Baltic Basin. Reefs and skeletal sand bar deposits of the shoal facies belt began to form in the Gotland area (Högklint Formation), northwestern Saaremaa (Ninase Member of the Jaani Formation) and eastern Lithuania (Jačionys Formation). However, in the basin depression the facies of dark-grey graptolitic muds even expanded its area at that time (Fig. 148). As a result, extremely rapid early-mid Wenlock regression along the perimeter of the basin finally transformed the Baltic Basin s.s. into a gulf-like pericratonic sea (the "Baltic Gulf"). The most drastic upheaval and regression took place in the Scandinavian part of the sea due to the progressing rise of the Caledonides.

During the Jaagarahu Age, the gradual shallowing continued at the margins of the basin, interrupted by short deepening episodes. This resulted in a cyclical alternation of high-energy shoal deposits (winnowed skeletal-pelletal sand and silt, coral-stromatoporoid banks and reefs) with different lagoonal-littoral dolomitic muds (*Eurypterus*- and bioturbated pattern-dolomites). The *Eurypterus*-dolomites probably formed in a brackish-water environment (Эйнасто 1968). Three of such shallowing-up mesocyclithes have been recognized in the Jaagarahu Formation and are treated as the Vilsandi, Maasi and Tagavere beds. Basinwards these intercalating shoal and lagoonal deposits of the Jaagarahu Formation were successively replaced by the bioclastic argillaceous-calcareous muds (Riksu Formation), greenish-grey calcareous-argillaceous muds (Jamaja Formation) and finally by dark grey terrigenous muds with graptolites (Rīga Formation). During late Jaagarahu time, a long sedimentation break and erosion of the earlier deposits took place all over the shelf plateau (Nestor & Nestor 1991). At the moment of the maximum shallowing, at the end of the Jaagarahu Age, a band of thinly interbedded marls and limestones (Ančia Member) was formed in the central depression of the basin, marking the end of the Late Llandovery - Middle Wenlock macrocycle in the basin development.

The **Late Wenlock - Middle Ludlow macrocycle** corresponds to the Rootsiküla and Paadla ages. It is represented by the cyclical alternation of winnowed skeletal-pelletal grainstones with sedimentary argillaceous dolostones (*Eurypterus*-, pattern- and microlaminated dolomites). Unlike the Jaagarahu cycles, the shoal and lagoonal facies of the Late Wenlock - Middle Ludlow mesocycles extended over a very wide area of southwestern Estonia, covering the whole levelled shelf plateau. Southwards these marginal-marine facies were rapidly replaced by the dark-grey graptolitic muds of the Siesartis and Dubysa formations which shows that at that time a rather steep gradient existed between the shelf plateau and basin depression. The open shelf and transitional facies belts were heavily reduced (Fig. 149). It means that a platform-type sedimentation with a very wide belt of marginal-marine facies was typical of the Late Wenlock - Middle-Ludlow macrocycle of basin development.

On the background of the cyclical sea-level fluctuations a

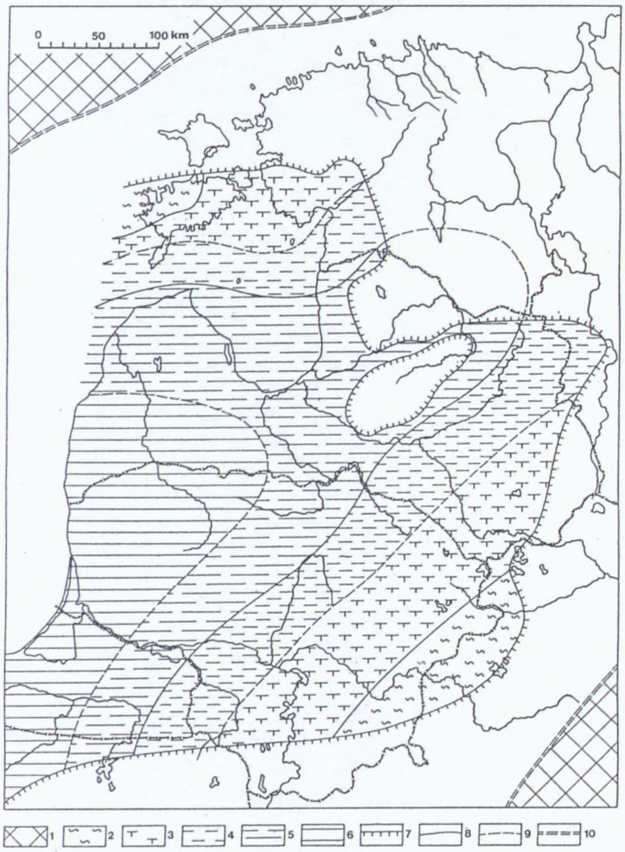


Fig. 148. Distribution of the Wenlock mid-Jaani (base of Ninase Member) sediments and facies belts: 1 - land; 2 - bioclastic calcareous mud; 3 - argillaceous-calcareous mud; 4 - green terrigenous mud; 5 - grey terrigenous mud with graptolites; 6 - dark kerogenous graptolitic mud; 7 - boundary of the present extension of rocks; 8 - main facies boundaries; 9 - boundary of sediment types; 10 - shoreline.

faintly expressed regressive-transgressive trend is perceivable in the basin evolution. The break point, *i.e.* the regression maximum was probably reached by the end of the Vesiku time (mid-Rootsiküla) and further a very slow, gradual transgression followed. It is likely that several sedimentation gaps existed during the time-interval in consideration, but direct evidence is still lacking and only a probable hiatus has been revealed between the Rootsiküla and Paadla stages (Table 8).

Infilling stage of development

It was the final epoch in the evolution of the Ordovician-Silurian carbonate sedimentation basin at the western margin of the East-European Platform which corresponded to the time interval from the Late Ludlow (Kuressaare Age) up to the earliest Devonian (Tilžė Age). An abrupt increase in the supply of the basin with terrigenous material, coming from the Scandinavian Caledonides, was a characteristic feature of the infilling stage. The previous side-filling of the basin depression was changed by the total infilling and shallowing all over the central part of the Baltic Syneclise where the deposition of terrigenous graptolitic muds was now replaced by the accumulation of the greenish-grey calcareous-argillaceous muds with benthic biota. The formation of the graptolitic muds of the depression belt migrated to the platform margin in north-eastern Poland where a thick clayey-silty complex of distal turbidites (Siedlce beds) was formed. Even in the shelf area, corresponding to the open shelf and shoal environments, the accumulation of the terrigenous material was so heavy that it diluted the carbonate sedimentation. As a result, the formation of bioclastic marls got dominance all over the shelf area. However, from time to time rather thick (3 to 5 m) deposits of crinoidal grainstones, coquinite banks with *Atrypoidea prunum*, thickets of rugose and tabulate corals were formed. Such thick deposits of crinoidal grainstones, containing cross-bedding, ripple marks and other signs of the agitated-water environment mark the end of several shallowing-up mesocycles. They are most typically developed in the Kaugatuma Stage where four of such cyclithes (Lower and Upper Äigu beds, Lower and Upper Lõo beds) have been distinguished. A clear facies zonation was characteristic of the moments of the formation of crinoidal grainstones in the shoal belt (Fig. 150), while during the rest of time with intensive influx of clayey component, the lateral differentiation of facies was much poorer.

The infilling stage began with a brief transgressive event during the Kuressaare Age. This Late Silurian transgression

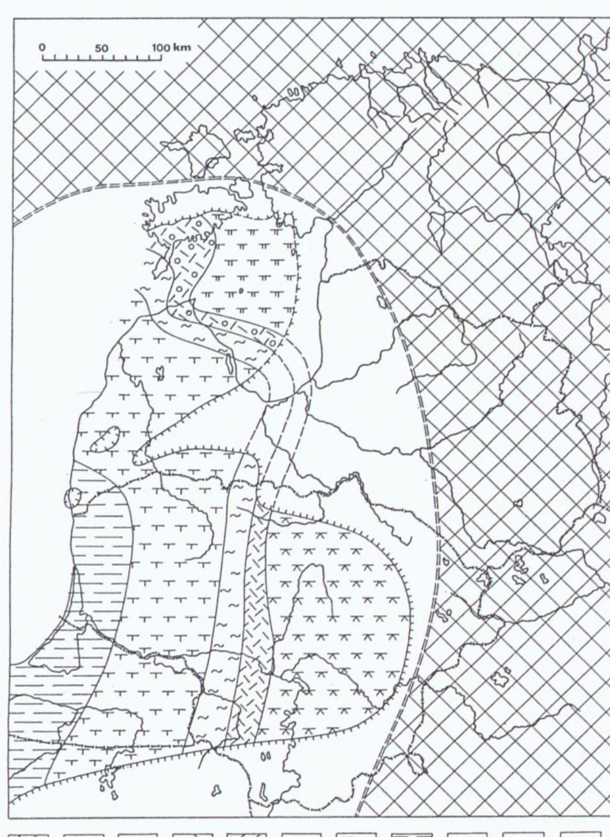


Fig. 149. Distribution of the Ludlow mid-Paadla (base of Himmiste beds) deposits and facies belts: 1 - presumable land; 2 - lagoonal dolomitic mud; 3 - high-salinity, gypsiferous dolomitic mud; 4 - pelletal-skeletal sand and silt; 5 - skeletal sand; 6 - bioclastic calcareous mud; 7 - calcareous-argillaceous mud; 8 - grey terrigenous mud with graptolites; 9 - boundary of the present extension of rocks; 10 - facies boundaries; 11 - shoreline.

reached its maximum at the beginning of the Přidoli (early Kaugatuma time). The transitional facies belt with deposition of calcareous-argillaceous muds (Šilalė beds of the Minija Formation) was very wide and extended from southwestern Latvia to northwestern Poland (Fig. 148). In onshore direction it was gradually followed by open shelf biomicritic marls, coarse-grained crinoidal gravel and sand of the shoal belt and lagoonal dolomitic muds. The latter have preserved only in the Lithuanian part of the basin. During the late Kaugatuma and Ohesaare times, the general facies pattern remained the same but all facies belts migrated gradually southwestwards.

By the beginning of the Devonian (Tilžė Age), only a remnant lagoon-like body of water was preserved in northern Kurzeme and south-western Lithuania. It was characterized by a clastic-dominated near-shore belt and argillaceous-dolomitic muds in its offshore part.

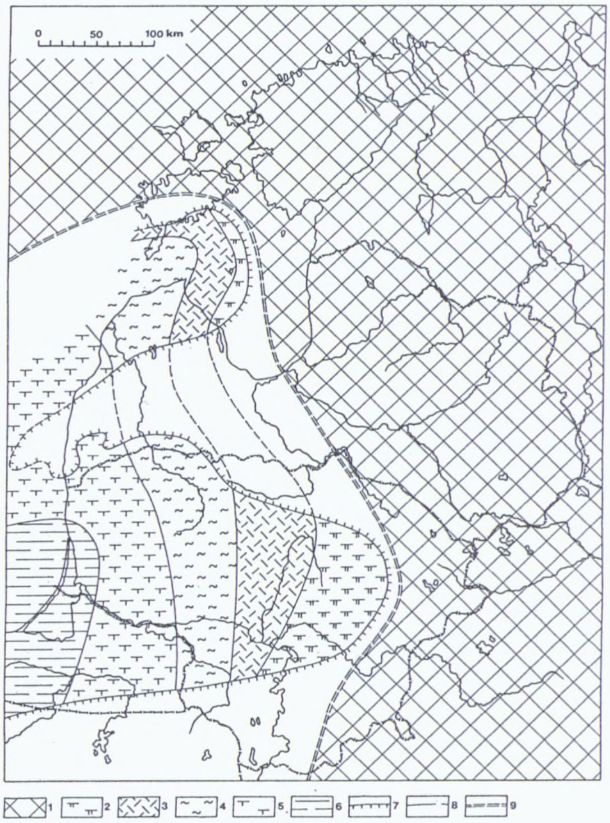


Fig. 150. Distribution of the Přidoli earliest Kaugatuma sediments and facies belts: 1 - land; 2 - lagoonal dolomitic mud; 3 - skeletal sand and gravel; 4 - bioclastic calcareous mud; 5 - green calcareous-argillaceous mud; 6 - grey terrigenous mud with graptolites; 7 - boundary of the present extension of rocks; 8 - facies boundaries; 9 - shoreline.

DEVONIAN SEDIMENTATION BASIN

Estonia belongs to the northwestern part of the Devonian Main Field of the East-European Platform which in the Devonian was situated in the equatorial region on the Laurussia Continent formed at the end of the Silurian. Its development was influenced by both tectonic movements and eustatic sea-level oscillations (Ziegler 1988). In the territory under consideration, epicontinental shallow sea sediments accumulated. In the Early Devonian and at the beginning of the Middle Devonian, this sea had a connection with the ocean in the northeast, later in the east and south-east. The sequence has a very complex cyclic structure indicative of pulsatory nature of the sedimentation process. The regression, which had started at the end of the Silurian, continued at the beginning of the Devonian and alternated with a short-term transgression in the middle of the Early Devonian during the Ķemeri Age. A significant long-term transgression started in the Rēzekne Age at the end of the Emsian and lasted until the end of the Leivu Subage in the second half of the Eifelian when a regression with a duration up to the end of the Givetian started. A new transgression started at the beginning of the Late Devonian when the northernmost marginal part of the sedimentary basin formed and extended to a restricted area in southeastern Estonia (Тихомиров 1967, Куршс 1992).

At the beginning of the Early Devonian, in the Lochkovian Age, the filling of the relict basin formed during the Silurian regression continued. The deeper axial part of this basin opening to the south-west was located in the Baltic Syneclise. Its northern margin, however, reached the present-day Estonia (Нарбутас 1984, Куршс 1992) where sandy sediments deposited. The region of the Mõniste Uplift served as a continental denudation area (Fig. 151A). The influx of terrigenous material into this basin was mainly from the Scandinavian massif in the north-west. The highly variable mineral composition of the sediments becomes more uniform in a southerly direction.

The sediments of the Lochkovian Tilžė Age occur in a restricted area in Estonia. Palaeontologically, they have been determined from the Laanemetsa and Värska core sections in southeastern Estonia (Сорокин 1981). The finds of thelodont scales in the Ruhnu core, which have redeposited from the Tilžė Stage into the basal beds of the Rēzekne Stage (Менс и др. 1992), show that earlier these sediments were distributed also in southwestern Estonia where they were afterwards subject to denudation. This supports the opinion of Kuršs (Куршс 1975) and Narbutas (Нарбутас 1984) who maintain that in the Tilžė Age the sediments formed as two tongues in the northern wing of the relict basin.

In the second half of the Lochkovian Age, sedimentation was interrupted in the northwestern part of the East-European Platform. The geocratic period, separating the Caledonian and Hercynian tectonic stages, began. In the Baltic area it covered the second half of the Lochkovian and the beginning of the Pragian. This period is characterized by marked denudation in the course of which the sediments in the marginal area of the basin were partly denudated. A new marine flooding invaded Baltic from the southwest. In the second half of the Pragian in the Ķemeri Age, sandy sediments started to accumulate here, while in the deeper part of the basin - the Polish-Lithuanian depression, sandy-silty sediments deposited. The weathering crust which had formed during the hiatus was redeposited. Therefore, Ķemeri sediments contain kaolinite in notable amounts and the sandy fraction has a relatively mature mineral composition. Additional terrigenous material was transported into the basin from the Scandinavian massifs. Possibly, the influx took place along the site of the present Gulf of Riga (Куршс 1975). In the northern part of the distribution area sandstones are much coarser. The northern margin of the Ķemeri Basin reached Estonia's territory and nearshore sandy sediments accumulated here (Fig. 151B). Supposedly, the Ķemeri sediments occur in a number of boreholes in southwestern Estonia (Ipiku, Ikla, Tõlla, Abja), although palaeontologically they have not been dated anywhere in Estonia.

The Ķemeri transgression was relatively short. At the end of the Pragian and at the beginning of the Emsian, the whole East-European Platform was governed by continental conditions and the earlier sediments were denudated. A new westerly transgression took place at the end of the Emsian during the Rēzekne Age. The sea flooded a great part of the East-European Platform, the waves reaching as far as the Moscow Syneclise (Куршс 1992). The northern part of the basin extended to Estonia's territory leaving behind sandy sediments of nearshore shallow sea. In southeastern Estonia, mostly offshore sandy-silty sediments deposited. Towards the end of the Rēzekne Age the sea became deeper and in southeastern Estonia normal-sea carbonate sediments accumulated. The influx of the terrigenous component was mostly from the north-

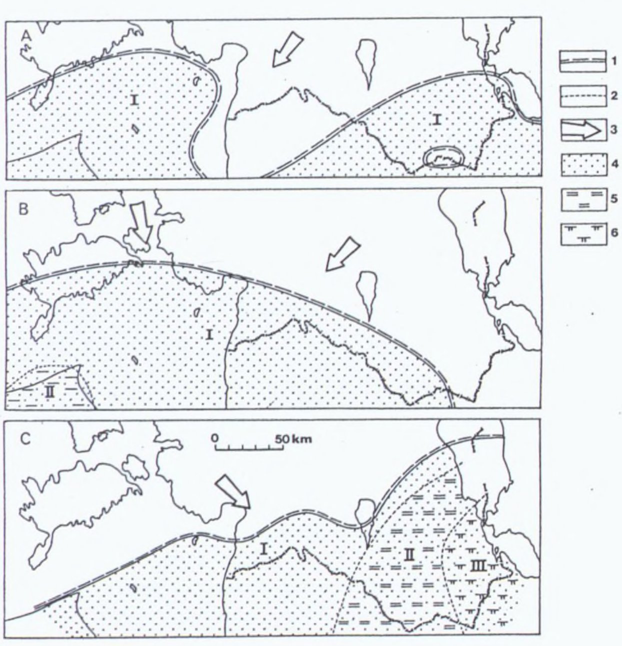


Fig. 151. Sketch-map showing the distribution of lithofacies during the Early Devonian: A - Tilžė Age, B - Ķemeri Age, C - Rēzekne Age. I - near-shore clastic sedimentation, II - shallow-marine sandy-silty sedimentation, III - shallow-marine clastic carbonate sedimentation. 1 - assumed boundary of the marine basin; 2 - boundary of deposition areas; 3 - main direction of the influx of clastic material; 4 - sandstone; 5 - siltstone; 6 - dolomitic marl.

west, from the Scandinavian massif. The grain-size of sediments and the areal location of zircon-rich belts suggest two possible ways of influx (Fig. 151C). The mineral composition shows that granitic massifs were subject to denudation.

At the beginning of the Pärnu Age, the basin retreated. The selective concentration of garnet possibly indicates that in the first half of the Pärnu Age the coastline stayed for a long time on the Strenči - Valga - Väimela - Värska line (Fig. 152A). In the middle of the Pärnu Age, the marine basin started to extend to the north, north-east and east, over a great part of the East-European Platform. In Estonia, mainly nearshore sandy sediments accumulated, but in the west evidently subwater delta sediments deposited. The influx of terrigenous material from the Scandinavian granite massifs (Fig. 152A) continued. The salinity of the Pärnu basin was presumably somewhat higher which is evidenced by the occurrence of gypsum. Gypsum is particularly characteristic of the areas located father in the east and in the southern part of the Baltic. Owing to intensive freshwater influx, the salinity was normal in the northwestern part of the basin. At the end of the Pärnu Age, beside sandy sediments also carbonate sediments with abundant sandy admixture accumulated. The occurrence of syneresis cracks and pyrite-rich surfaces in carbonate sediments indicates periodical hiata in sedimentation at that time.

The Middle Devonian marine transgression in the East-European Platform reached its maximum in the Narva Age when a shallow basin with carbonate sedimentation formed the major part of the platform. The center of sedimentation was in the Baltic Syneclise where the thickness of sediments reached 150-180 m (Куршс 1992). The northwestern part of the basin covered the current site of Estonia (Fig. 152B,C). In the Narva Age, three subages in the basin evolution are distinguished: Vadja, Leivu and Kernavė (see Fig. 83).

The basin widened markedly in the Vadja Age when almost the entire Estonian territory was included in the sedimentation area (Fig. 152B). In the northern part of the East-European Platform dolomitic muds with variable clay content were accumulated. In the northwestern part of the basin, the sequence starts with a peculiar landslide breccia or layers of breccia-like domerite. The genesis of breccia is also related to a break in sedimentation, but a more probable reason seems to be an underwater landslide caused by tectonic movements. The occurrence of syneresis cracks and pyritic surfaces suggests that water was shallow and the basin episodically turned dry. The presence of gypsum and caverns, formed due to the leaching out of halite, is indicative of brackish-water conditions at the time of deposition during the Vadja Age. Owing to intensive freshwater influx from the north, the salinity of water in Estonia's area was somewhat lower. The influx of detrital material from the north-west continued but, additionally, some material was also derived from the Kola massif in the north-east (Fig. 152B,C). As the result of the northeasterly influx across the Ordovician outcrop, the sediments of the Vadja Subage have yielded redeposited Ordovician conodonts and crinoidal stem fragments from a height of up to 10 m above the Ordovician/Devonian contact. High corundum concentration in the region of the Navesti and Narva rivers shows that these watercourses served as the influx channels, but it also evidences of the proximity of the shoreline. The character of sediments (fine alteration of dark-grey silty clays with domerites and dolomites, the occurrence of psammitic material) suggests sedimentation in the tidal belt (Hettinger 1995).

The basin turned shallower and the sea retreated southward at the end of Vadja Subage. The earlier deposited layers were denudated and weathered. This level is marked by a sandstone layer with a mature mineral composition and high roundness of minerals. It is overlain by dolomites or domerites enriched with sandy material. By the beginning of Leivu Subage, the sea had retreated back to the limits of the Baltic Syneclise. Soon a new flooding followed and the basin widened northwards reaching the maximum extent at the end of the Leivu Subage. This basin was shallow, with increased salinity and carbonate sedimentation. Its basic source area was as earlier in the Scandinavian massifs. The main influx of terrigenous material was into the region of the present-day Lake Võrtsjärv depression characterized by a NW-SE facies belt enriched with sandy-silty material. The influx from the

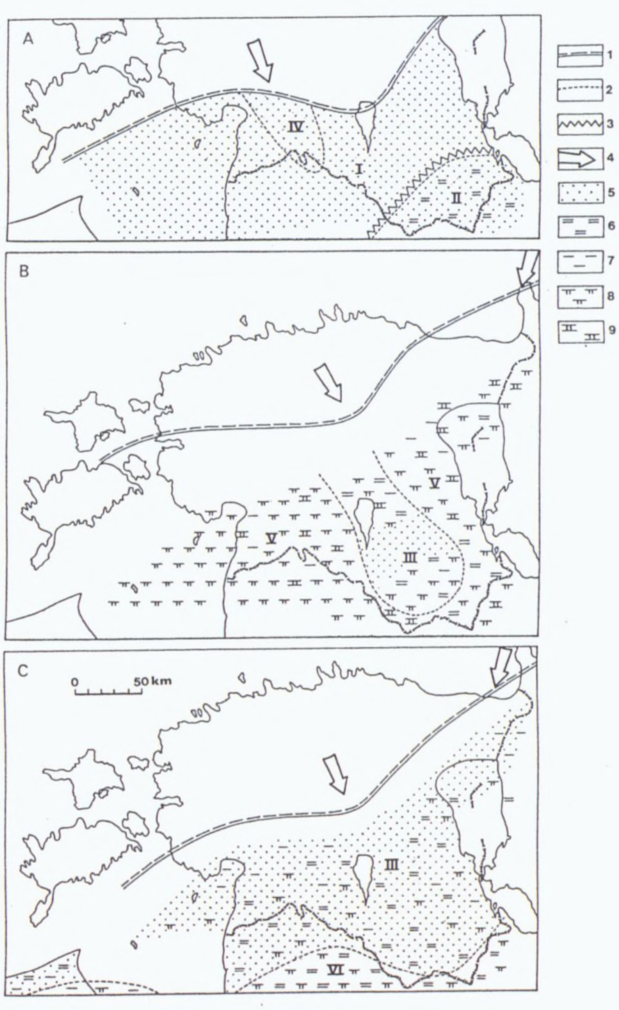


Fig. 152. Sketch-map showing the distribution of lithofacies during the Eifelian: A - Pärnu Age, B - Vadja + Leivu subages, C - Kernavė Subage. I - near-shore clastic sedimentation, II - shallow-marine sandy-silty sedimentation, III - shallow-marine clastic-carbonate sedimentation, IV - estuarian clastic sedimentation, V - coastal plain and tidal carbonate sedimentation, VI - offshore carbonate sedimentation. 1 - assumed boundary of the marine basin; 2 - boundary of deposition areas; 3 - belt of garnet concentration; 4 - main direction of the influx of clastic material; 5 - sandstone; 6 - Ssiltstone; 7 - clay; 8 - dolomitic marl; 9 - dolomite.

north-east continued as well (Fig.152B). Marked alteration of the mineralogical composition, first of all a decrease in garnet and titanite, and an increase in apatite beginning from the second member of the Leivu Substage, but also the changes in the typomorphic varieties of minerals indicate that new massifs were subjected to denudation (Fig. 78). Accumulation of red-coloured sediments at the end of the Leivu Subage refers to a humid type of weathering in the denudation area (Куршс 1992).

The Kernavė Subage was characterized by a regression trend. As a result of more intensive influx of terrigenous material and fresh water, the basin turned into a waterbody of normal salinity with rich biota where mostly red sandy, in the Baltic Syneclise also normal marine grey carbonate sediments deposited. In the territory of the present-day Estonia, which was situated in the northwestern marginal area of this basin, sandy and variegated clayey-carbonate sediments of nearshore shallow sea accumulated. The offshore facies belt with the accumulation of silty-carbonate sediments embraces only a restricted portion of the southeasternmost part of Estonia. As previously, the influx of terrigenous material was via the present Võrtsjärv Depression and northeastern Estonia (Fig. 152C). Changes in the mineral composition indicate some alteration of the initial massifs (Fig.78).

The shallowing trend in the basin development continued in the Aruküla Age when terrigenous sedimentation came to dominate. The cyclic structure of the sequence suggests frequent sea-level fluctuations against the background of general regression trend. Three main cyclic complexes permit the distinction of three successive levels - the Viljandi, Kureküla and Tarvastu beds, corresponding to definite evolutionary stages (Kleesment 1994, Fig. 85). Supposedly, the sea retreated finally from Estonia during the initial phase of the accumulation of Kureküla and Tarvastu deposits, in which short-term hiata in sedimentation occur. The preserved sediments, however, are mainly of marine genesis. The character of sediments suggests the deepening of the basin to the south-east, while the pattern of garnet concentration refers to a fairly stable position of the shoreline on the Häädemeeste - Abja - Viljandi - Aakre - Slantsy line (Fig. 153A). The influx of terrigenous material was as formerly from the north-west and north-east. The appearance of staurolite among accessory minerals shows that besides granitic massifs also metamorphic rocks were subject to denudation (Fig. 78). A general increase in the maturity of the mineral composition and the growth of the degree of roundness of grains upwards (Kleesment 1994) confirm that in the Aruküla Age the rate of redeposition was high.

The slow retreat of the marine basin, interrupted by temporary transgressions, continued in the Burtnieki Age. Against the background of the general shallowing trend, three regressive-transgressive stages can be distinguished in the sediments of the Härma, Koorküla and Abava beds (Kleesment 1995, Fig. 85). The regressive stages were characterized by short-term breaks in sedimentation. In Estonia, mainly near-shore facies and delta sediments deposited. The shoreline, marked by garnet concentration, shifted to the south-east compared with its location during the Aruküla Age (Fig. 153 A, B). Underwater delta sediments have been identified in the Joosu quarry (Куршс 1992). Source terrigenous material came as before from the Scandinavian massifs where the area of denudation had reached metamorphic rocks. Evidence is derived from the higher proportion of staurolite and kyanite among accessory minerals (Fig. 78). Redeposition processes came to play an ever increasing role as suggest the growth of the quartz, zircon and rutile content and an increase in the degree of grain roundness upward the section.

The pulsatory retreat of the marine basin and partial redeposition of older sediments continued in the Gauja Age and were accompanied by a constant influx of fresh detrital material from the Scandinavian metamorphic massifs. Concurrently with the continuing retreat of the sea and probable humidification of climate (Куршс 1992), the processes of subaerial weathering were intensive in the second half of the Gauja Age. The sediments, formed at that time, are characterized by a high degree of maturity which is revealed in the high concentration of quartz, zircon, tourmaline and rutile, and large amounts of kaolinite among clay minerals. In Estonia, nearshore marine sandy sediments, in the south-east sandy-silty sediments, deposited in the Gauja Age when a great part of the territory was under delta (Куршс 1992, Fig. 153C).

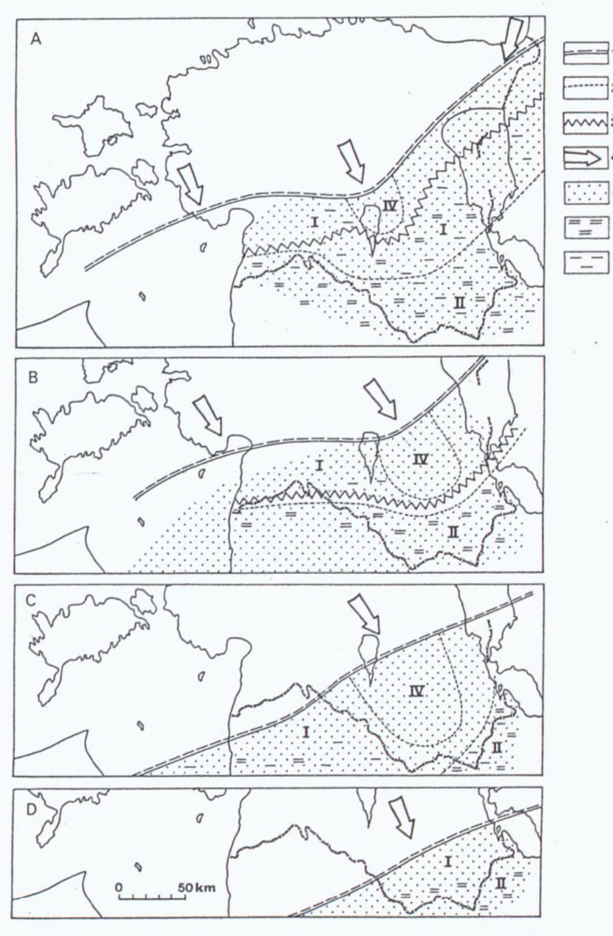


Fig. 153. Sketch-map showing the distribution of lithofacies at the end of the Eifelian (Aruküla Age) and during the Givetian: A - Aruküla Age, B - Burtnieki Age, C - Gauja Age, D - Amata Age. I - near-shore clastic sedimentation, II - shallow-marine sandy-silty sedimentation, IV - estuarine clastic deposits. 1 - assumed boundary of the marine basin; 2 - boundary of the deposition areas; 3 - belt of garnet concentration; 4 - main direction of the influx of clastic material; 5 - sandstone; 6 - siltstone; 7 - clay.

In the Amata Age, the palaeogeographical situation remained unchanged with respect to the source area and the general configuration of the marine basin. In Estonia, sandy and sandy-silty sediments of nearshore shallow sea accumulated (Fig. 153D). The scantiness of coarse terrigenous material, the lack of fresh minerals and the increasing roundness of grains suggest that the Scandinavian massifs were peneplaned and mostly older sediments were redeposited. Irregular inclination of cross-bedded series is indicative of complicated sedimentation conditions (Kleesment 1995).

At the end of the Amata Age, the link with the ocean in the south-west broke. A new, Frasne transgression started from the Moscow Syneclise in the east. The northwestern point of the sea, which covered the middle part of the East-European Platform and was characterized by Pļaviņas carbonate sedimentation, reached southeastern Estonia. In general, the salinity of the basin somewhat increased, but at the current site of Estonia it was normal due to freshwater influx (Сорокин 1978). Sammet (Саммет 1971), basing on the presence of gypsum-rich deposits east of Estonia, has treated the sediments of this area as lagoonal. As earlier, the influx of terrigenous material was from the north.

EVOLUTION OF LIFE DURING VENDIAN - DEVONIAN

Introduction

The sedimentary cover of Estonia was formed during the first half of the Phanerozoic and records, therefore, only more primitive forms of life: marine algae, cyanophytes, invertebrates, the earliest vertebrates and vascular plants. Fossils are very unevenly distributed throughout the sequence in which, besides the strata extremely rich in fossils rather extensive stratigraphical gaps and almost barren rocks occur.

The fossils are extremely rare in the Vendian and Cambrian clastic deposits, accumulated on southern high latitudes (Fig. 138). They contain neither elements of the famous Ediacara fauna nor diverse Cambrian assemblage of archaeocyathids which obviously settled only the equatorial climatic zone. Acritarchs and problematic thallus-like organical remains - *Vendotaenites* are the only fossils found from the Vendian strata of Estonia. Such poor and specific content might be explained by freshened-water conditions which probably existed during Late Vendian Kotlin time.

The very sparse Lower **Cambrian** fossil assemblage contains single representatives of probable annelids *(Platysolenites),* pogonophores *(Sabellidites),* gastropods *(Aldanella),* problematic phylum Agmata *(Volborthella),* trilobites *(Schmidtiellus)* (Photo 41:1), jellyfishes *(Medusites),* phosphatic-shelled brachiopods from the class Paterina *(Paterina, Mickwitzia* -Photo 39:1), monoplacophores *(Scenella),* ichnofossils *(Skolithos, Diplocraterion)* indicating a rather high diversity of phyla comparing with the Vendian biota. From the Upper Cambrian (Ülgase, Tsitre, lowermost Kallavere formations) representatives of the class Lingulata *(Ungula, Angulotreta, Schmidtites, etc.)* and the earliest conodonts *(Westergaardodina, Furnishina, Cordylodus, etc.)* have been identified.

In the earliest **Ordovician** (main part of the Pakerort Age), the first possible bryozoan *(Marcusodictyon)* and dendroid graptolites *(Rhabdinopora)* appeared, belonging to the earliest representatives of these groups. The first chitinozoans date from the Upper Tremadoc (Varangu Stage). In the Arenig, during the Billingen Age, articulate brachiopods appeared in the Baltic area in the Volkhov Age, cephalopods and cystoids were added. These were just brachiopods, trilobites and cephalopods that played a leading role in the latest Early Ordovician faunas in Estonia.

During the Middle Ordovician, a very rich assemblage of brachiopods, ostracodes, trilobites, bryozoans, cystoids occurred in the area under consideration. This assemblage contained a number of endemic taxa as many of clitambonaceans among brachiopods, asaphids and cheirurids among trilobites *etc.* It allows to speak about the special North-European or Baltoscandian faunal province (Jaanusson 1979). This was definitely a temperate-climate fauna as it totally lacked reef-building corals and stromatoporoids, rather wide-spread on some other cratons (North America, Siberia, Kazakhstan, North China, Australia) at that time. The first and very primitive solitary rugose corals *(Primitophyllum, Lambelasma),* as obviously less dependent on climate, appeared during the Haljala Age. The earliest known echinoid - *Bothriocidaris* (Photo 35) made its appearance during the same age which is much earlier than in any other region of the world. The first

Photo 35. Echinoid *Bothriocidaris eichwaldi* Männil, holotype, adoral view, Upper Ordovician, Pirgu Stage, Jootma at Tapa, x 9. *Photo by Ralf Männil*

stromatoporoids *(Stromatocerium)* and tabulate corals *(Lyopora, Eoflecheria)* are known since the Middle Caradoc, the Oandu Age. In several groups (bryozoans, brachiopods, trilobites, conodonts) some North American faunal elements *(e.g. Howellites, Zygospira, Rynchotrema, Bumastoides, Belodina)* immigrated. This suggests the beginning of a gradual remission of provincial differences between the Baltic region and North America which progressed during the whole Late Ordovician. Only at the beginning of the Silurian, the fauna in the Baltic area became more or less cosmopolitan as during the end-Ordovician mass-extinction only the most tolerant, wide-ranged taxa survived while the more specialized, endemic elements died out (Нестор и др. 1991).

In the **Silurian,** a clear lateral differentiation of ecological assemblages took place. Within the shallow-water faunas the role of corals and stromatoporoids rose abruptly, while in deeper-water associations graptolites, chitinozoans, trilobites and ostracodes preponderated. In most of taxonomical groups lateral ecological communities have been distinquished (see Кальо и Клааманн 1982, 1986).

During the Silurian, a remarkable evolutionary diversification took place among the coral and stromatoporoid faunas reaching its acme in the Late Llandovery - Wenlock time. The earliest known representatives of several higher taxonomical groups, such as stromatoporoid orders Actinostromatida, Syringostromatida and tabulates Theciida, Syringoporida, originated from the Baltic region.

Agnathans and fishes were also very rapidly evolving groups. The first vertebrates in the Baltic area (thelodont *Loganellia*) are known from the Middle Llandovery (Raikküla Stage), whilst in some other regions (North America, Australia) agnathans were present at least since the end of the Early Ordovician (see p. 245). Such delay in appearance might be explained by nektobenthic mode of life of the early agnathans which prevented their immigration into the Baltic Basin until the beginning of the closure of the Iapetus Ocean. During the Silurian, the representatives of gnathostomes gradually appeared in the Baltic area: first acanthodians

(Gomphonchus) during the Late Llandovery (Adavere Age), bony fishes - osteichthyans *(Andreolepis)* in the Early Ludlow (Paadla Age). They both are the oldest representatives of the corresponding groups. Chondrichthyans (elasmobranchs) are known since the Early Přidoli (Kaugatuma Age).

Since the Middle Wenlock (Jaagarahu Age), eurypterids became rather common in near-shore, lagoonal facies. However, the appearance of these arthropods in the Baltic area was obviously connected with arising environmental conditions favourable for their settlement as well as their preservation in the fossil record. In some other regions, the eurypterids are known since the Cambrian already. Nearly at the same time, the earliest representatives of another group of merostomates - xiphosurans *(Bunodes, Pseudoniscus)* – appeared for the first time in the fossil record.

On the one hand, progressive increase in the role and diversity of corals, stromatoporoids, ostracodes and especially vertebrates was the most remarkable trend in the evolution of the Silurian faunas in the Estonian area. On the other hand, it was accompanied by a decrease in the diversity of brachiopods, trilobites and some other groups.

During the **Devonian**, mainly terrigenous or carbonate terrigenous sedimentation took place in the marginal marine environment in Estonia, while in the Frasnian carbonate deposition predominated again.

The fossils of the lowermost Devonian, coming from the boring cores, are rare. Moreover, a significant part of the Lower Devonian rocks is lacking because of a stratigraphical gap between the uppermost Lochkovian and the upper Emsian. Fish faunas belonging to the Euramerica Province are important throughout the Devonian. Fishes and plants (macroremains, miospores and gyrogonites? of charophyte algae) are the most common fossils in the Upper Emsian and Middle Devonian rocks. Among invertebrates, lingulate brachiopods (*Bicarinatina*) and conchostracans (*Glyptoasmussia, Ulugkemia, etc.*) are more numerous in comparison with ostracodes, gastropods and bivalves. In the Middle Devonian, from the Narva to the Burtnieki Age, fishes reached the greatest variety. All main fish groups (agnathans, heterostracans and osteostracans) were present. In the Baltic area, the characteristic members of the fish assemblages were psammosteid heterostracans (*Tartuosteus, Pycnosteus, Ganosteus*), placoderms (*Homostius, Asterolepis*), numerous acanthodians and crossopterygians (osteolepidids, *Glyptolepis*). Dipnoans (*Dipterus*) and actinopterygians (*Orvikuina, Cheirolepis*) were also frequent. In the late Middle Devonian Gauja Age the fish fauna became gradually poorer. Among the psammosteids *Psammolepis* predominated; among crossopterygians *Laccognathus* was frequent. Later on, in Amata and, particularly, in Pļaviņas time, the psammosteid genus *Psammosteus* and the antiarch *Bothriolepis* were most significant. Among the fishes, in the latter time, ptyctodont *Ctenurella* and, actinopterygian *Moythomasia* were common. The invertebrate fauna of the Pļaviņas Age includes brachiopods, stromatoporoids, tabulate corals, *etc.*

Vascular plants show changes from the *Hyenia* flora with *Psilophytites* and *Hostinella* of the Pärnu Age and *Pseudosporochnus* of the Burtnieki Age to the *Archaeopteris* flora of the Gauja Age. The latter flora is characteristic of the Late Devonian.

The scarcity of the Devonian fossils in comparison with those of the Early Palaeozoic depends on their specific preservation conditions rather than on unsuitable conditions of life.

Ordovician chitinozoans

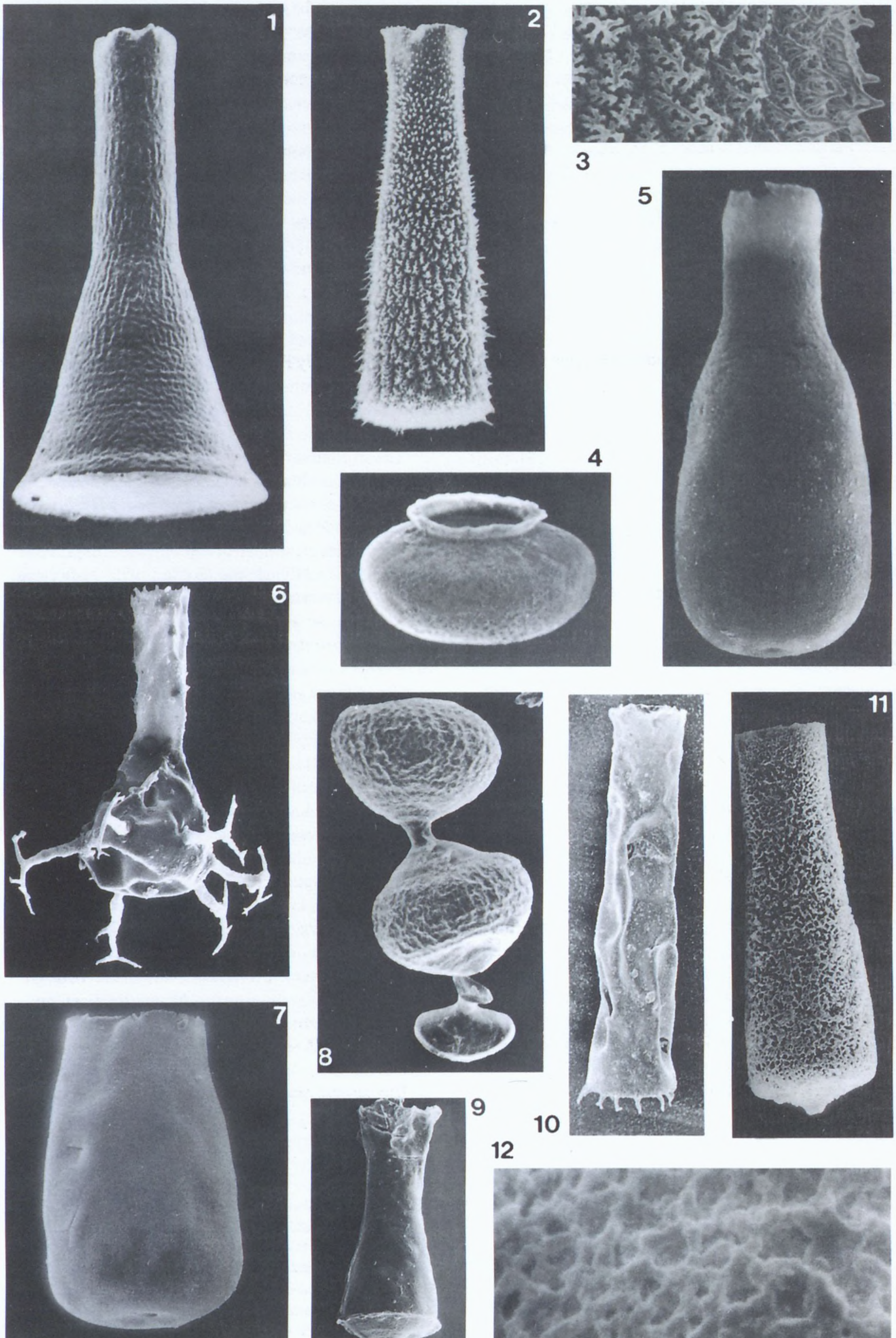
Eisenack, in his pioneer works on chitinozoans from the early 1930s, studied mainly Ordovician and Silurian glacial erratic boulders in the South-Baltic coastal area. The biostratigraphical potential of chitinozoans was proved by Ralf Männil (1971). Nõlvak and Grahn (1993) elaborated a detailed Ordovician biozonation scheme for the whole Baltoscandia, which is presented in Table 7.

The chitinozoans were exclusively marine organisms, their palaeogeographic distribution shows clearly a pelagic mode of life. They are treated as an extinct planktic group of unknown affinity ranging from the Ordovician to Devonian. Chitinozoans display a rather indistinct provincialism at the generic level. Data concerning the lateral associations of chitinozoans are uncertain yet.

The huge amount of subsurface sections, excellent outcrops and good preservation in the Ordovician and Silurian sequences of Estonia have made chitinozoans (Photo 36:1-12) useful for stratigraphical purposes (Männil 1971, Nõlvak 1972, *etc.*). They have been used for compilation of all recent stratigraphical charts (Решения... 1978,1987). Compared to other groups, chitinozoans give most precise stratification and correlation on many levels, ecpecially in the Viru and Harju series (Table 7). Beginning from the first record of chitinozoans in the uppermost Tremadoc up to the topmost Ashgill, 15 zones and 8 subzones have been defined (Nõlvak & Grahn 1993).

The earliest record of chitinozoans comes from the top-

Photo 36. Ordovician and Silurian chitinozoans:
1. *Cyathochitina campanulaeformis* (Eisenack), Ch 708/5412, Ashgill, Nabala Stage, Eikla borehole, depth 275.5 m, x 230.
2-3. *Belonechitina robusta* (Eisenack), Ch 893/4070, Caradoc, Oandu Stage, Kuusiku borehole, depth 10.2 m: 2- x 230, 3- x 1200 (detail).
4. *Calpichitina complanata* (Eisenack), Ch 723/5778, Ashgill, Nabala Stage, Kaugatuma borehole, depth 387.9 m, x 580.
5. *Lagenochitina baltica* Eisenack, Ch 907/7405, Ashgill, Pirgu Stage, Jaksai borehole, depth 968 m, x 350.
6. *Ancyrochitina ramosaspina* Nestor, Ch 427/1919, Llandovery, Raikküla Stage, Ikla borehole, depth 480.4 m, x 340.
7. *Eisenackitina dolioliformis* Umnova Ch 454/10670, Wenlock, Jaani Stage, Jaagarahu borehole, depth 41.7 m, x 500.
8. *Cingulochitina cingulata* (Eisenack), Ch 151/1952, Wenlock. Jaagarahu Stage, Ohesaare borehole, depth 294.16 m, x 350.
9. *Margachitina margaritana* (Eisenack), Ch 419/1604, Wenlock, Jaagarahu Stage, Ohesaare borehole, depth 174.4 m, x 465.
10. *Spinachitina maennili* (Nestor), Ch 2/1462, Llandovery, Raikküla Stage, Ikla borehole, depth 462.9 m, x 340.
11-12. *Conochitina cribrosa* Nestor, Ch 336/1598, Wenlock, Jaagarahu Stage, Ohesaare borehole, depth 188.2 m: 11- x 300, 12- x 2800 (detail).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most layer of the stratotype section of the Varangu Stage in northern Estonia where *Lagenochitina esthonica* Eisenack appears. The post-Tremadoc Ordovician has a relatively high taxonomic diversity of chitinozoans with more than 130 species being recorded. Only heavy secondary dolomitization in some parts of the sequence (mainly in the lowermost and uppermost Ordovician) has caused the absence of chitinozoans. They do not occur in high-energy grainstones, reef facies and marine redbeds either.

A relatively rich assemblage of chitinozoans has been found beginning from the lowermost beds of Arenig, from the glauconite-rich Leetse sandstones (Hunneberg Stage). It is diversest in the Llanvirn - Lower Caradoc strata (from the Kunda to Haljala stages) where the average number of species reaches 25 (Fig. 154). Throughout the Ordovician, the mean number per stratigraphical unit is 19 (Kaljo *et al.* 1996). Higher up in the sequence, the taxonomic diversity is below the average with some exceptions in the lower Nabala and Vormsi stages.

In general, the following main changes can be distinguished in the dynamics of the chitinozoan diversity. During early Volkhov to Kunda time, the most intensive origination of taxa took place. The Kukruse - Haljala interval was a time of rapid diversity fluctuations, characterized by many short-ranging species. The late Keila extinction and diversity minimum in the Oandu Age coincided with a considerable sedimentological change in the sequence (Hints *et al.* 1989). This relatively short episode might be called the "Oandu crisis" and is probably globally observable. At the beginning of the Nabala Age, the intensive origination started. Late Pirgu - early Porkuni time is characterized by mass extinction, only four taxa ranged into the lowermost Silurian, across the Ordovician - Silurian boundary.

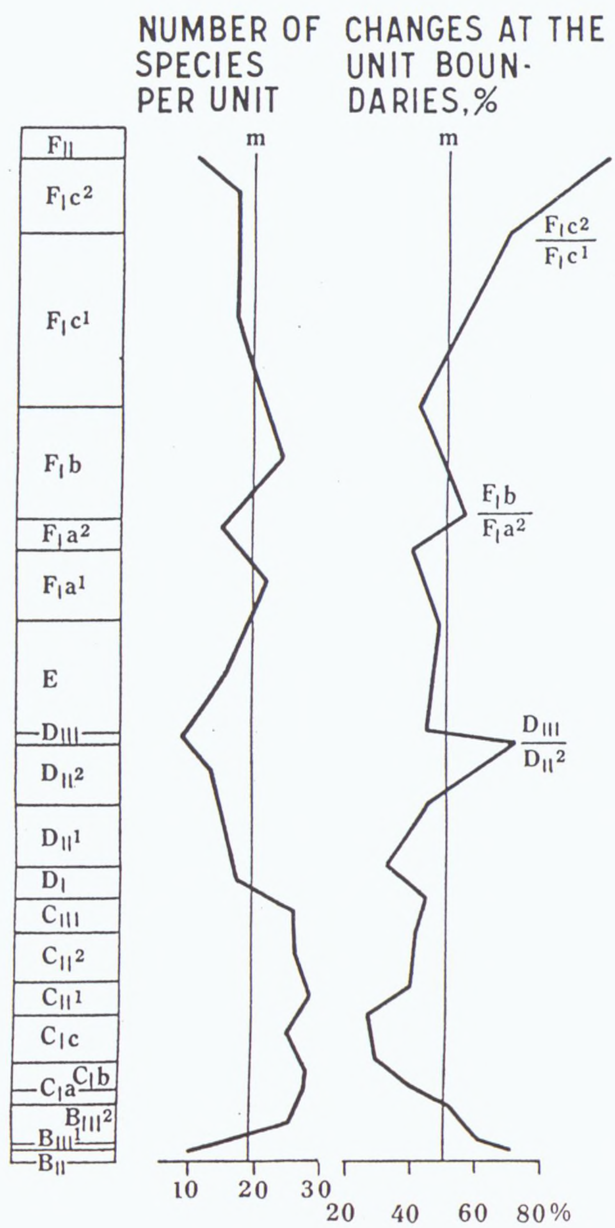

Fig. 154. The diversity of Ordovician chitinozoans in the Rapla core section. Stratigraphical column reflects natural thicknesses. Stratigraphical indices: F_{II} - Porkuni Stage; F_Ic^1, F_Ic^2 - lower and upper parts of the Pirgu Stage; F_Ib - Vormsi Stage; F_Ia^1, F_Ia^2 - lower and upper parts of the Nabala Stage; F - Rakvere Stage; D_{III} - Oandu Stage; D_{II}^1, D_{II}^2 - lower and upper parts of the Keila Stage; D_I - Jõhvi Substage, Haljala Stage; C_{III} - Idavere Substage, Haljala Stage; C_{II}^1, C_{II}^2 - lower and upper parts of the Kukruse Stage; C_Ic - Uhaku Stage; C_Ib - Lasnamägi Stage; C_Ia - Aseri Stage; B_{III}^1, B_{III}^2 - lower and upper parts of the Kunda Stage; B_{II} - Volkhov Stage; m - mean value level.

Silurian chitinozoans

The investigation of Silurian chitinozoans in Baltoscandia was initiated by Eisenack (1931-76), continued by Laufeld (1974) and Grahn (1995) in Sweden, by Ralf Männil (Мяннил 1970) and Viiu Nestor (1990, 1992, 1994, Нестор В. 1976, 1984) in Estonia.

The evolution of the Silurian chitinozoans (Photo 36:6-12) was rather slow on the higher taxonomic level (genera, families). Most of the genera entered into the Silurian from the Ordovician and continued in the Devonian, but the species assemblage and diversity experienced considerable changes throughout the Silurian (Nestor 1992). In the Silurian of Estonia, four main cycles of evolution of chitinozoans can be distinguished (Fig. 155 A) on the basis of the diversity and taxonomic innovation of chitinozoan assemblages. At the end of each cycle, extensive disappearance of taxa took place (Fig. 155 B); at the beginning of the next cycle abundant new elements appeared. All cycles are characterized by the presence of taxa with specific morphological features, typical of a definite stratigraphic interval. The Early and Middle Llandovery cycle was still characterized by the presence of some specific Ordovician genera (*Cyathochitina, Spinachitina*) and by the scarcity of mucronated vesicles. In the Late Llandovery - Early Wenlock cycle, chainlet and copulated forms (*Densichitina, Margachitina*) made their appearance. Mucronated forms (*Conochitina*) and those, in which the spines were arranged in rows on the vesicle surface (*Gotlandochitina*), gained abundant distribution. The Middle Wenlock is characterized by a high abundance and diversity of chitinozoans only in the sections of southwestern Estonia. At that time, species with a spongy and reticulated ornamentation occured. At the end of the Late Wenlock, 80% of taxa became extinct. At the beginning of the Ludlow - Přidoli cycle, in the Estonian sequence 90% of the chitinozoan assemblage got renewed. A big variety of cylindro-spherical, ovoidal and lenticular forms appeared gradually up to the end of the Kaugatuma Age, when

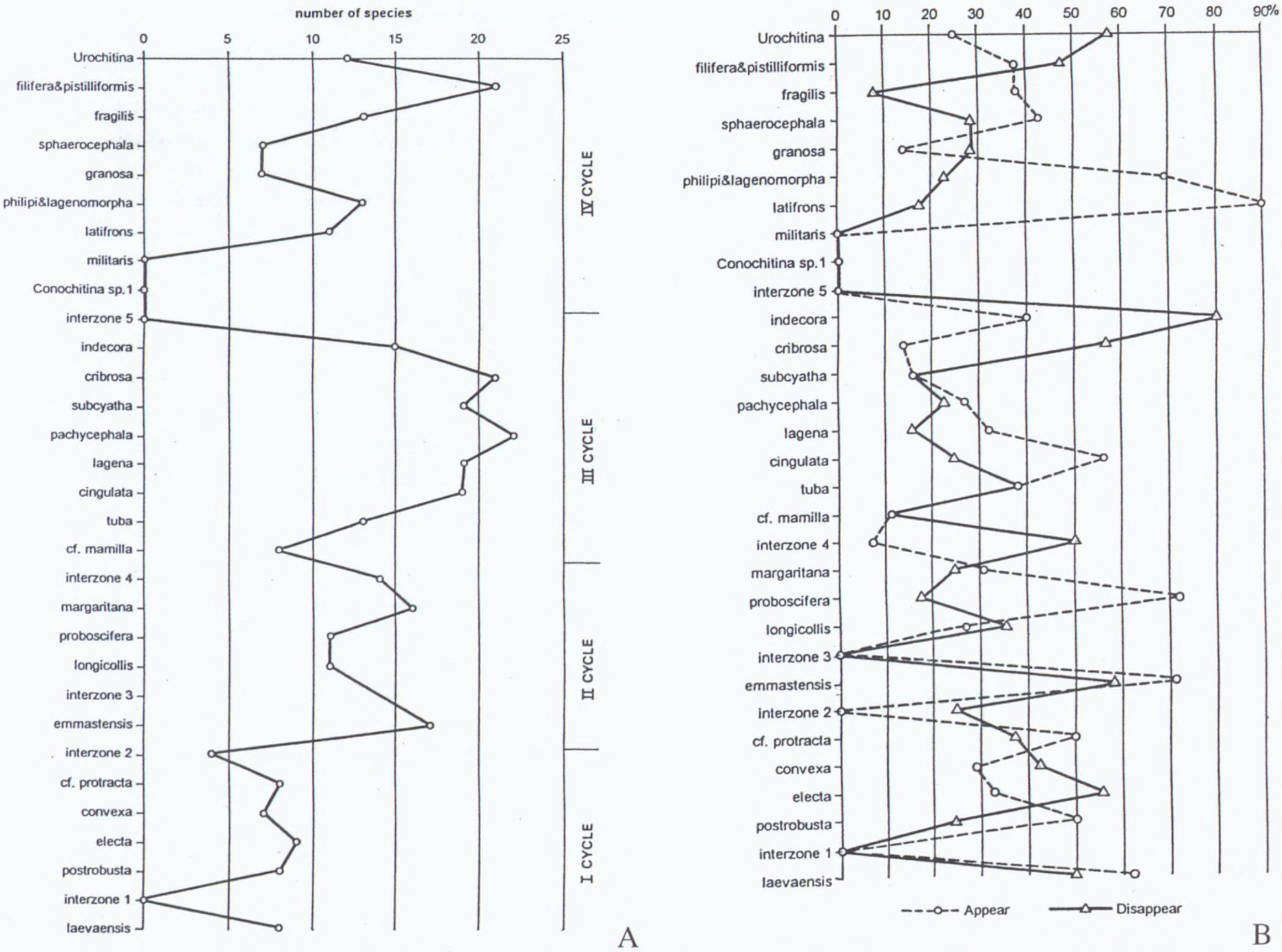

Fig. 155. Chitinozoan diversity changes and evolutionary cycles (A) and the percentage of appearing and disappearing species per chitinozoan biozone (B) in the Silurian of Estonia. The charts were compiled using data from the Ohesaare core and in the case of stratigraphic gap (the main part of the Middle Llandovery) from the Ikla core section.

the diversity of chitinozoan decreased considerably. The above-mentioned cycles were also closely related to the sea-level fluctuations in the Baltic Basin during the Silurian.

In the Silurian sequence of Estonia, 31 chitinozoan biozones have been distinguished (Nestor V. 1990, Fig. 15), five of which are called interzones as they contain scarce chitinozoans without specific forms (see Table 8). The lower limit of the biozone is usually defined by the first appearance of the zonal species or by the disappearance of a number of species occurring in the previous zone. Good relationship with the regional graptolite biozonation allows to use chitinozoan data for age determination in shelly sequences (Nestor V. 1994).

A wide spectrum of environmental parameters has affected the ecology and distribution of chitinozoans in different facies belts. The abundance and diversity of chitinozoans reach the maximum in the sections of southern and southwestern Estonia where marls and argillaceous limestones are predominating. In more carbonate sections of middle Estonia their variability and number decrease, and the occurrence becomes more sporadic.

The distribution of chitinozoans was also controlled by the temperature of water which can be generally associated with palaeolatitude. For a global biozonation of Silurian chitinozoans, the index species, irrespective of palaeolatitudes and palaeoplate configuration, were selected avoiding usage of endemic taxa. Almost all index species are represented in the Silurian succession of Estonia (Verniers *et al.* 1995, fig. 4).

Algae and vascular plants

The **Vendian-Silurian** flora in Estonia is represented by problematic organic-walled thallus-like macrofossils — *Vendotaenides* (originally found and named by Eichwald as *Laminarites*-type algae), calcareous algae, *Gloeocapsomorpha* as the main precursor organism of oil shale organic matter, acritarchs as a group of palynomorphs of controversial affinity, but possibly comprising also algal cysts, and prasinophyte palynomorphs *Tasmanites* and *Leiosphaeridia*. Oncolites and stromatolites, biosedimentary structures characteristic of intertidal and subtidal marine environments, have also revealed microbial-algal fossils.

Taxonomical studies of Estonian calcareous algal palaeoflora date back to works by Eichwald (1854, Эйхвальд 1840), Dybowski (1877a) and Stolley (1893, 1896a, 1896b, 1897, 1898). Table 28 summarizes the present knowledge of the distribution of calcareous algae in the Ordovician and Silurian and presents the species list. The distribution of acritarchs in the Vendian, Cambrian and Ordovician has been treated in several papers (Пашкявичене 1980, Volkova *et al.* 1983, Mens *et al.* 1993, Uutela & Tynni 1991).

The Lower Ordovician in Estonia lacks calcareous algal

Table 28. Distribution of calcareous algae in the Ordovician and Silurian (after Kõrts *et al.*, 1990)

SILURIAN	K_3a	*Wetheredella silurica* *Rothpletzella gotlandica, R. straeleni, R. munthei*	
	K_2	*Hedstroemia halimedoidea* *Parachaetetes gotlandicus* *Parachaetetes compactus* *Ortonella*	*Bevocastria amplefurcata* *B. crassa*
	K_1	*Wetheredella tenue* *Garwoodia aff. gregaria* *Rothpletzella gotlandica* *Parachaetetes compactus* *Solenopora*	*Bevocastria amplefurcata*
	J_2	*Rhabdoporella flexuosa* *Solenopora filiformis* *Parachaetetes globosus* *Dimorphosiphon rectangulare* *Hedstroemia bifilosa, H. halimedoidea* *Rothpletzella munthei*	*Bevocastria* *Ortonella* *Girvanella ducii* *Girvanella wetheredii*
	J_1	*Rothpletzella gotlandica* *Girvanella wetheredii, G. ducii* *Wetheredella silurica*	*Vermiporella* *Rhabdoporella* *Halysis*
	H	*Vermiporella* *Rhabdoporella*	
	G_3	*Vermiporella* *Cyclocrinites*	*Wetheredella* *Girvanella*
	G_{1-2}	*Solenopora* *Vermiporella* *Rhabdoporella*	*Girvanella* *Rothpletzella* *Wetheredella*
ORDOVICIAN	F_2	*Vermiporella* *Rhabdoporella*	
	F_1c	*Vermiporella* *Rhabdoporella* *Dimorphosiphon*	*Girvanella* *Rothpletzella* *Wetheredella*
	F_1b	*Solenopora* *Vermiporella* *Girvanella*	*Rhabdoporella* *Wetheredella*
	F_1a	*Cyclocrinites spasskii* *Vermiporella* *Rhabdoporella*	*Dasyporella* *Wetheredella*
	E	*Coelosphaeridium wesenbergensis* *Cyclocrinites balticus, C. schmidti, C. spasskii, C. mickwitzi* *Vermiporella* *Rhabdoporella* *Rothpletzella* *Dasyporella*	
	D_3	*Solenopora spongioides, S. nigra, S. dendriformis* *Vermiporella*	
	D_2	*Mastopora concava* *Cyclocrinites* *Nuia*	
	D_1	*Mastopora concava* *Cyclocrinites porosus* *Coelosphaeridium cyclocrinophilum* *Solenopora spongioides*	
	C_3	*Coelosphaeridium* *Mastopora concava* *Solenopora filiformis*	
	C_2	*Coelosphaeridium kohtlense*	
	C_1c	*Coelosphaeridium*	
	C_1b	*Coelosphaeridium excavatum* *Mastopora odini*	
	C_1a	*Coelosphaeridium excavatum*	

Stratigraphical indices of stages: C_1a - Aseri; C_1b - Lasnamägi; C_1c - Uhaku; C_2 - Kukruse; C_3 - Idavere Substage (Haljala Stage); D_1 - Jõhvi Substage (Haljala Stage); D_2 - Keila; D_3 - Oandu; E - Rakvere; F_1a - Nabala; F_1 b - Vormsi; F_1c - Pirgu; F_2 - Porkuni; G_{1-2} - Juuru; G_3 - Raikküla; H - Adavere; J_1 - Jaani; J_2 - Jaagarahu; K_1 - Rootsiküla; K_2 - Paadla; K_3a - Kuressaare.

macrofossils, but the presence of algal debris in carbonate rocks has been mentioned by Põlma (Пылма 1982). Cyclocrinitids (*Cyclocrinites* Eichwald, *Coelosphaeridium* Roemer, *Mastopora* Eichwald), unique Palaeozoic dasycladacean macroalgae evolved by the Middle Ordovician (the earliest-known are *Coelosphaeridium excavatum* Stolley in the Aseri Stage and *Coelosphaeridium kohtlense* Bekker in the Kukruse Stage), diversified and formed an abundant flora in the Baltoscandian shallow sea during the late Middle Ordovician Rakvere Age. As the biology of these organisms is not sufficiently understood, the dasyclad affinity of *Cyclocrinineae* Pia has been recently questioned by Nitecki & Spjeldnaes (1992), suggesting a separate taxonomical unit for these extinct algae.

The first dasyclads had appeared already in the Precambrian, but reached remarkable abundance in the Ordovician. During the Middle Ordovician, another group of dasycladacean algae - the vermiporellids (*Vermiporella* Stolley, *Rhabdoporella* Stolley) appeared and differentiated. These algae are abundant in the Upper Ordovician Vormsi and Pirgu stages, extending their range up to the Jaani Stage in the Silurian. Stratigraphical range of the genus *Rhabdoporella* Stolley in Estonia is based on data from the Rapla (Vormsi - Pirgu stages) and Seliste (Juuru - Jaani stages) cores where *Rhabdoporella* is abundant in certain layers. Recent unpublished finds from the Kuldiga Formation of the Porkuni Stage in the Taagepera and Ruhnu boreholes confirm that, in all likelihood, *Rhabdoporella* like *Vermiporella*, formed algal mats occupying large areas in the shallow epicontinental sea (Jux 1966) and had the widest geographical distribution in Baltoscandia and North America (Poncet & Roux 1990). The occurrence of numerous specimens of *Rhabdoporella pachyderma* Stolley in the Porkuni Stage is associated with the *Hirnantia*-fauna.

Solenoporacean algae (Rhodophyta) emerged in the Cambrian, but being rare at that time, they arose in the Middle Ordovician nearly simultaneously in Estonia (*Solenopora filiformis* in the Idavere Stage), Scotland and North America (Roux 1991) comprising two genera – *Solenopora* Dybowski and *Parachaetetes* Deninger. The latter genus is more typical of Ludlow patch reefs in Estonia.

In the Wenlock and Ludlow, the cyclocrinitid-vermiporellid flora was replaced by different algal communities consisting of solenoporaceans, codiacean *Dimorphosiphon* Hoeg and various microalgae, including *Hedstroemia* Rothpletz, *Wetheredella* Wood, *Ortonella* Garwood and *Bevocastria* Garwood (Радионова и Эйнасто 1986).

The **Devonian** plant remains (Photo 37:1-10), particularly those coming from the Tori locality on the Pärnu River are known since Eichwald's studies (Эйхвальд 1854 a.o.). Karpinsky (1906) included into his description of charophytes the gyrogonites ("trochiliscs") collected from Tartu.Thomson (1940) described both macroremains and miospores from Tori and Küllatova. The early Middle Devonian miospores were figured and listed by Vaitiekūnienė (Клеесмент и др. 1975, Валюкевичюс и др. 1986) and Kedo (Сорокин 1981). Küllatova miospores were recently identified by Loboziak (Blieck *et al.* 1996). Yurina (Юрина 1988) revised Thomson's material from Tori and Küllatova and preliminarily identified plants from the clay quarry at Joosu. The latter locality yielded a new fossil species showing sporangia (*Pseudosporochnus estonicus*), described by Kalamees (1988), who also redescribed plant macrofossils from Tori. In 1982, one more rich plant locality – Kose (Oore) downstream of Tori – was discovered.

Table 29 shows three well-known stratigraphical units with plant macroremains identified on generic and species level: (1) Tori Member, (2) upper clayey part of the Abava Substage and (3) the Lode Member. (1) The Tori Member includes *Hostinella* sp. and *Psilophytites* sp. belonging either to Pteridophyta or Progymnospermopsida (Kalamees 1988). Earlier these plants, known as *Aulacophycus sulcatus* Göpp., *Asteroxylon* and *Aneurophyton*, were considered as psilophytes. (2) *Pseudosporochnus estonicus* (Kalamees 1988) from Joosu is a pteridophyte. (3) According to Meien (Мейен 1987), the fossils *Archaeopteris fissilis* Schalh. and *Archaeopteris* sp., identified from the Lode Member by Yurina (Юрина 1988), are progymnosperms. *Hostinella* is also reported from the same level (Thomson 1940).

If to list all Devonian flora occurrences, there is hardly any stratigraphical unit without plant remains. Rather common are gyrogonites(?) of charophyte algae (*Sycidium, Trochiliscus*), particularly in the Tamme Member of the Pärnu Stage, called earlier the "Trochilisken-Sandstein" (Orviku 1930). Charophyte algae occur also in all three members of the Narva Stage and in the Viljandi Member of the Aruküla Stage. Calcareous algae are known from the Pskov Substage of the Pļaviņas Stage (Марк и Паасикиви 1960). The Burtnieki and Gauja stages have yielded silicified and ferriferous wood. Rocks of the Narva, Aruküla and Burtnieki stages contain on several levels unidentified plant remains.

Besides macroremains, plant microfossils also occur on certain stratigraphical levels. Miospores have been established in the grey-coloured rocks of the Rēzekne Stage (Lower Devonian) and of the Tori and Vadja members (Middle Devonian). They occur also much higher up, in the Lode Member of the Gauja Stage. Acritarchs have been discovered from the Vadja Member of the Narva Stage.

Table 30 shows the stratigraphical range of the miospores and the probable position of the spore zones given by Avkhimovitch *et al.* (1993) for the Devonian of Eastern Europe. The *Periplecotriletes tortus* (PT) Zone is well established in the Tori Member by the presence of the index species and *Calyptosporites velatus*. The miospores *Punctatisporites tortosus* and *Hymenozonotriletes ludzus* (= *Grandispora ludza*), characterizing the earlier, *Diaphanospora inassueta* (DI) Zone and established in the Rēzekne Stage of the southern Baltic and Belarus (Avkhimovitch *et al.* 1993), occur in the Estonian sequence in the Pärnu Stage. The Tori Member (Pärnu Stage) and the Vadja Member (Narva Stage) have quite a number of common spore species. The Vadja assemblage seems to be older than that of the *Rhabdosporites langii* (RL) Zone, characteristic of the Kernavė Member in Lithuania. The spore assemblage in the Lode Member (Gauja Stage) can be considered as that of the *Accyrospora incisa- Geminospora micromanifesta* (IM) Subzone (Blieck et al. 1996). In Belarus, the miospores of this subzone occur in the lower portion of the Lan' Stage which is an approximate equivalent of the Gauja Stage (Valiukevičius *et al.* 1995).

Stromatoporoids

In the Ordovician and Silurian strata of Estonia, 88 valied species of stromatoporoids have been described (Rosen 1867, Nicholson 1886-91, Рябинин 1951, Нестор 1960, 1964, 1966). The full list of the species, belonging to 26 genera and 16 families, and representing all orders of stromatoporoids except fine-cylindrical amphiporids, was published recently (Nestor 1990c). Stromatoporoids (Photo 38:1-3) are continuously present in all regional stages of Estonia beginning from the Lower Ashgill (Vormsi Stage) and ending with the Lower

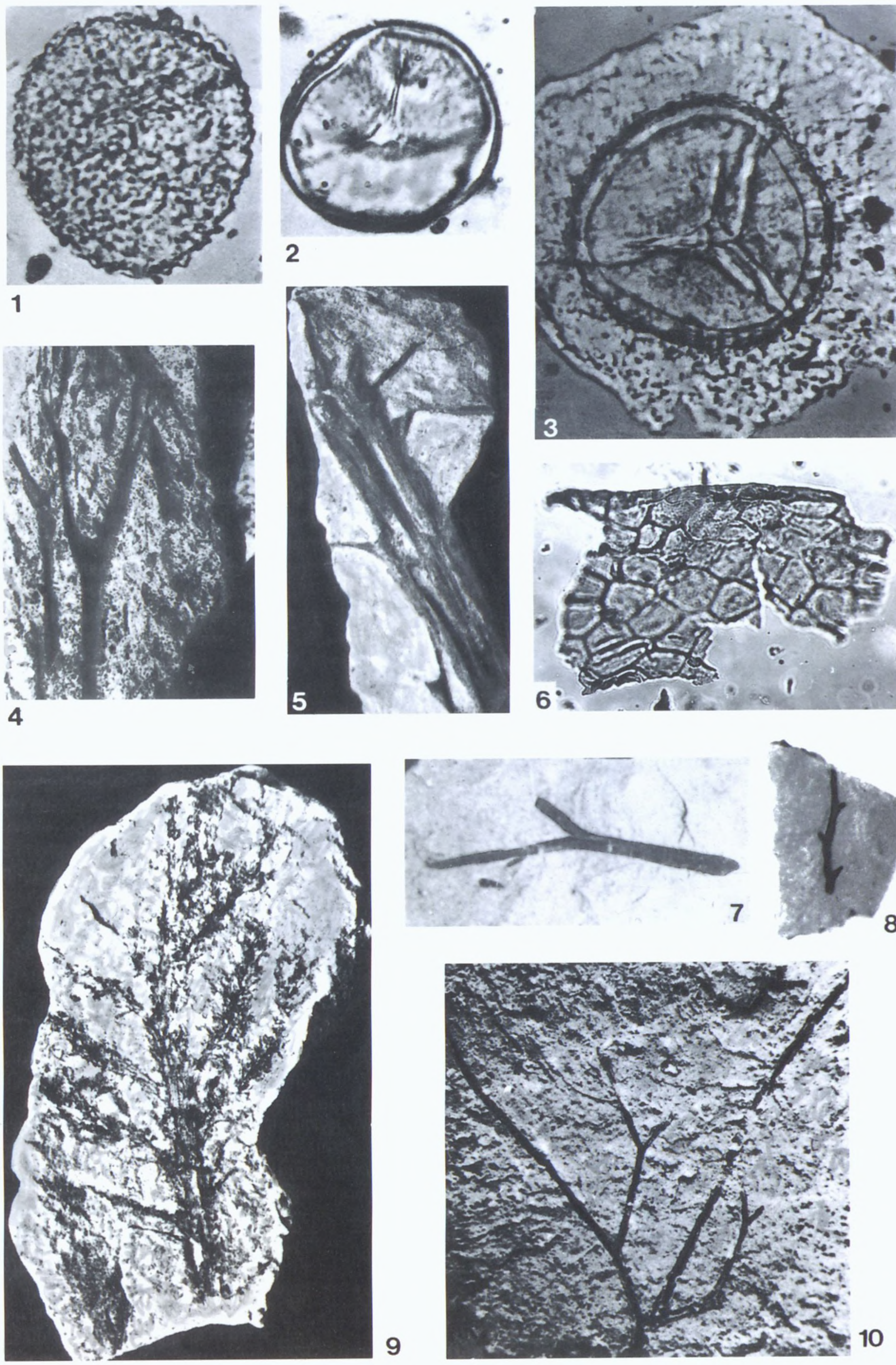
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Table 29. Plant remains in the Devonian of Estonia. Outcrop names are in bold characters

R.Stage	Member, Beds, Subst.	Location	Plant fossils	Taxon
Pļaviņas	Chudovo			
	Pskov		calcareous algae	
	Snetnaya G.			
Amata				
Gauja	Lode	**Küllatova**	stems and sprouts	*Hostinella* *Archaeopteris fissilis* *Archaeopteris* sp.
			miospores	
	Sietiņi	**Jõksi**	ferriferous wood	
		Piusa	silicified wood	
Burtnieki	Abava	**Joosu**	stems, branches with fertile & vegetative fronds	*Pseudosporochnus estonicus*
		Essi	fragmental branches	
	Koorküla			
	Härma	**Varbuse**	silicified wood	
Aruküla	Tarvastu	Värska (85 m)	fragmental branches	
	Kureküla	**Tamme**	fragmental branches	
	Viljandi	Valguta(44.8-48 m)	fragmental branches	
		Tartu, Tähtvere Häädemeeste(2.6-5.5) Otepää(182.5 m) Ennu (104 m)	charophytes	*Trochiliscus* *Sycidium*
Narva	Kernavē	**Gorodenka**	fragmental branches	
		Häädemeeste(18 m) Kuningaküla(6.3-6.4) Mära (352.2 m)	charophytes	*Sycidium*?
	Leivu	**Poruni (Borovnja)** **Narva R**.at Gorodenka Häädemeeste (49.7-52) Luutsniku(318.8-317; 366.4 m)	charophytes	*Sycidium*?
	Vadja	Tõlla (70.6-73.5 m)	charophytes	*Sycidium*
		Kaagvere(121-130.5) Mehikoorma(195-199) Luutsniku(385-389.2)	**miospores**, acritarchs **miospores**, acritarchs **miospores**	
Pärnu	Tamme	**Tammeküla**	charophytes	
	Tori	**Kose(Oore)** **Tori**	stems&dichotomously divided branches	*Hostinella* *Psilophytites*
		Tori	**miospores**	
Rēzekne	rz_2	Mehikoorma (223-224 m)	**miospores**	
	rz_1	Mehikoorma (244.1-244.6 m)	**miospores**	

Photo 37. Middle Devonian plant micro- and macrofossils:
1-3. Miospores: 1- x 850, 2- x 650, 3- x 820; 4-5. Branches of *Hostinella* and/or *Psilophytites* (either pteridophytes or progymnosperms): 4- x 0.9, 5- x 1.25; 6. Cuticula of *Hostinella* sp., x 310. All specimens from Pärnu Stage, Tori Member, Tori.
7-8. Branches of the pteridophyte *Pseudosporochnus estonicus* Kalamees, Burtnieki Stage, Abava beds, Joosu: 7- specimen Va 2193, x 1.3; 8- Va 2197, nat. size.
9. Branches of *Hostinella,* Gauja Stage, Lode Member, Küllatova, x 4.7.
10. Branches of the progymnosperm *Archaeopteris,* Gauja Stage, Lode Member, Petseri, x 1.1.

Table 30. Range of miospores in the Lower-Middle Devonian of Estonia (Клеесмент и др. **1975,** Сорокин **1981,** Валюкевичюс и др. **1986, Blieck** ***et al*****. 1996 and Obukhovskaya, pers. comm.**

	DEVONIAN					
	Lower		Middle			
Miospore zone	DI		PT	RL EX		IM
Formation	Rēzekne		Pärnu	Narva		Gauja
Subf., Member	*rz*$_1$	*rz*$_2$	*pr T*	*nr V*		*gj L*
Leiotriletes simplex Naumova	+					
Stenozonotriletes conformis Naumova	+					
Acanthotriletes perpusillus Naumova	+					
A. parvispinosus Naumova	+					
Archaeozonotriletes memorabilis V. Umnova	+					
Dibolisporites cf.*eifeliensis* (Lanninger)McGregor	+					
Diatomozonotriletes devonicus Naumova	+					
Retusotriletes simplex Naumova	+	+				
Leiotriletes microrugosus (Ibr.)Naumova	+	+				
Emphanisporites rotatus McGregor	+		+			
Retusotriletes cf.*priscus* V. Umnova		+				
Leiotriletes cf.*insuetus* V. Umnova		+				
Granulatisporites cf.*rudigranulatus* Staplin		+				
cf.*Calamospora pannucea* Richardson		+				
Periplecotriletes tortus Egorova			+			
Dibolisporites antiquus (Kedo)Arkh.			+			
Calyptosporites velatus (Eisenack)Richardson			+			
C. tener (Tchib.)Obukh. var.*concinnus* Tchib.			+			
Sinuosisporites sinuosus (V. Umnova)Arkh.			+			
Punctatisporites tortuosus (Tchib.)Arkh.			+			
Hymenozonotriletes marginodentatus Kedo			+			
H. longus Arkh.			+			
Camarozonotriletes apertus Kedo			+	+		
Retusotriletes raisae Tchib.			+	+		
R. devonicus Naumova			+	+		
R. concinnus Kedo			+	+		
R. incomptus Kedo			+	+		
Hymenozonotriletes ludzus Kedo			+	+		
H. altus Kedo				+		
H. cf.*marginodentatus* Kedo				+		
H. cf.*echiniformis* Kedo				+		
Retusotriletes planituberculatus Kedo				+		
R. cf.*brandtii* Streel				+		
R. fragosus Arkh.				+		
R. microsetosus Kedo				+		
R. lanceolatus Kedo				+		
R. luxispinus Kedo				+		
R. engurensis Kedo				+		
R. clivosiformis Kedo				+		
Grandispora protea (Naumova)Moreau-Benoit				+		
Phylotecotriletes triangulatus Tiw. et Schaarscmidt				+		
Geminospora micromanifesta (Naumova)Arkh.						+
G. lemurata Balme, emend. Playford						+
Retusotriletes rugulatus Riegel						+
Cristatisporites triangulatus (Allen)McGregor et Camfield						+
Samarisporites eximius (Allen)Loboziak et Streel						+
Ancyrospora sp. cf.*A. incisa* (naumova)M. Rask. et Obukh.						+
Dictyotriletes sp. cf.*Reticulatisporites perlotus* (Naumova)Obukh.						+
Perotrilites sp. cf.*Rugospora*? *impolita* (Naumova)Tchib.						+

Index species are in bold characters. DI - *Diaphanospora inassueta* Zone; PT - *Periplecotriletes tortus* Zone; RL - *Rhabdosporites langii* Zone; EX - *Geminospora extensa* Zone; IM - *Ancyrospora incisa-Geminospora micromanifesta* Subzone. The interval between the Leivu Member of the Narva Formation and Burtnieki Formation (incl.) is supposed to include the RL and EX Zones.

Přidoli (Kaugatuma Stage). However, some earliest representatives of stromatoporoids (*Stromatocerium canadense* and *S. sakuense*) occur in the Middle Caradoc (Oandu Stage) already. Shorter gaps in the distribution, explained with unfavourable ecological conditions or local stratigraphical hiatuses, occur at the Llandovery/Wenlock and Wenlock/Ludlow boundaries.

Stromatoporoids appeared in the Estonian sequence later than in North America, North China or Australia where the earliest indubitable stromatoporoids have been recorded from the Llanvirn - Llandeilo strata already. It has been explained with the location of the Baltica Continent in the southern temperate climate zone until the Ashgill time when it migrated finally into the equatorial belt (Webby 1980).

In the Ordovician of Estonia, stromatoporoids are rare, except its topmost part - the Porkuni Stage. A few species of the most primitive, vesicular stromatoporoids (Order Labechiida) have been recorded from the Oandu (*Stromatocerium*), Pirgu (*Cystostroma*) and Porkuni (*Pacystylostroma*) stages. *Plumatalinia,* a problematic intermediate form between labechiidae and reticulate stromatoporoids (Actinostromatidae), occurs in the Pirgu Stage (Table 31). In the latest Ordovician, representatives of the sublaminate stromatoporoids (Order Clathrodictyida) became rather common: *Clathrodictyon* appeared during the Vormsi Age and *Ecclimadictyon* in the Porkuni Age.

In the Llandovery, clathrodictyids flourished. *Clathrodictyon* Nich. *et* Murie and *Ecclimadictyon* Nestor became dominating genera. They formed more than 80% of stromatoporoid specimens. Labechiids (*Pachystylostroma*, *Forolinia*, *Rosenella* and *Labechia*) were the second abundant group. During the Llandovery, the first representatives of several families appeared in the Estonian sequence. Thus, during Raikküla time, *Intexodictyon* - the earliest known representative of the Family Atelodictyidae (Table 31), and *Plectostroma*, the first certain representative of the Order Actinostromatida, made their appearance. During the Adavere Age, *Petridiostroma* was added among clathrodictyids as the most ancient representative of the Family Gerronostromatidae. At the same time, a very peculiar form *"Stromatopora" elegans* Rosen (=*Pachystroma*) appeared, showing the closest affinities to the Family Pseudolabechiidae. Thus, during the Llandovery, the first genuine laminate stromatoporoids (Atelodictyidae and Gerronostromatidae) and different branches of reticulate stromatoporoids (Actinostromatidae, Pseudolabechiidae) were gradually added to the prevailing fauna of the sublaminate and vesicular stromatoporoids.

During the Wenlock, the enrichment and diversification of the stromatoporoid fauna continued. In Jaani time, the first known microreticulate stromatoporoid *Densastroma* appeared. It belonged to the family Densastromatidae and played an important role later in the Silurian. At the same time, *Stromatopora* appeared in the Estonian sequence, being one of the earliest representatives of the Order Stromatoporida, *i.e.* stromatoporoids with irregularly amalgamated skeletal elements. *Simplexodictyon validum* Nestor, an early representative of the tripartite-laminated stromatoporoids (Order Stromatoporellida), has been recorded from the Maasi beds of the Jaagarahu Stage. *Vikingia tenuis* (Nestor) was the main frame builder in the Jaagarahu reefs (Vilsandi beds); it may be treated as a possible ancestor of the Order Syringostromatida with clinoreticulate microstructure of vertical skeletal elements (Nestor 1994). During the Wenlock the role of clathrodictyids decreased considerably; labechiids have not been recorded from Estonia.

During the Early Ludlow (Paadla Age), the diversity of the stromatoporoid fauna reached its maximum (13 genera from 12 families) in Estonia. In different taxonomical branches new elements were added, *e.g. Lophiostroma* among labechiids, *Plexodictyon* among clathrodictyids, *Pseudolabechia* in Actinostromatida, *Syringostromella* among stromatoporids, and *Parallelostroma* - the first representative of the Order Syringostromatida in the Estonian sequence. However, representatives of all the above-mentioned genera are known from somewhat earlier strata in other regions (Nestor 1994). In the Late Ludlow (Kuressaare Age) and Early Přidoli (Kaugatuma Age), the diversity of stromatoporoid assemblages decreased considerably due to an increase in the clay content of sediments and bad exposure of the corresponding strata. Stromatoporoids have not yet been recorded from the Silurian Ohesaare Stage.

The above leads to the conclusion that favourable climatic conditions for constant colonization of the Estonian area by stromatoporoids were established during the Ashgill. The beginning of the Silurian was characterized by comparatively unilateral fauna of sublaminate clathrodictyids (*Clathrodictyon, Ecclimadictyon*) with an admixture of labechiids. During the Llandovery and Wenlock, representatives of most of the orders and families were gradually added, and in the Early Ludlow (Paadla Age) the diversity of the stromatoporoid fauna reached its maximum, falling after that rapidly due to a progressive increase in the influx of clayish clastic material from the raising Caledonides.

Stromatoporoids were highly facies-dependent organisms with comparatively narrow ecological niche. The richest and most diverse stromatoporoid association occurred in the high-energy shoal facies belt, represented in fossil record by coral-stromatoporoid boundstones, skeletal and coquinite grain- and rudstones (Nestor 1990c). They were rather numerous also in the moderate- to low-energy open-shelf facies belt where biomicritic deposits (nodular skeletal packstones) were accumulated. Parallel successions of imperfectly deliminated lateral communities have been distinguished for shoal and open-shelf environments, consisting of 23 stromatoporoid communities (Nestor 1990c). Up to now, no definite biogeographic provinces have been established for the Late Ordovician and Silurian stromatoporoids (Nestor 1990b).

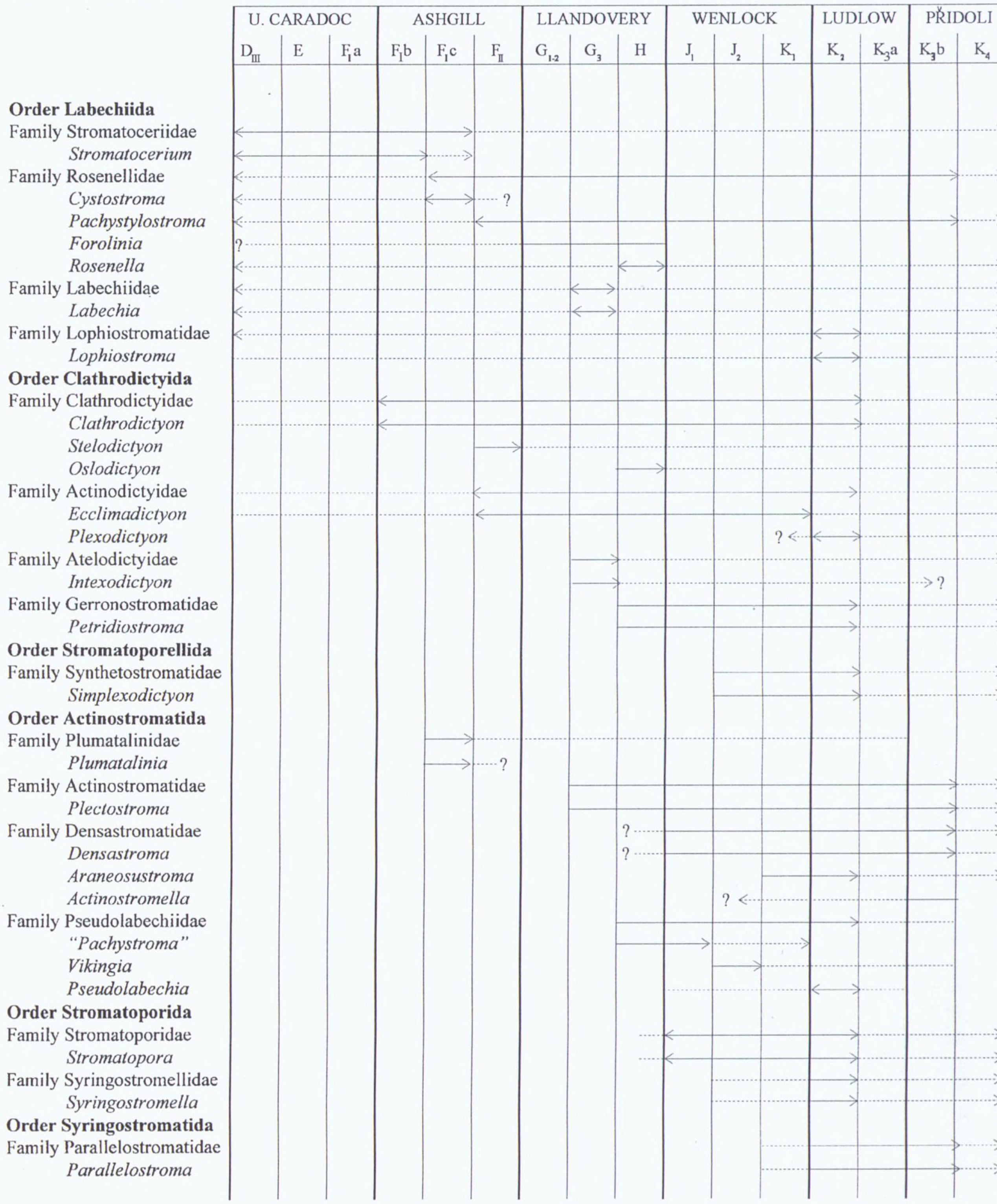
Tabulate corals

Estonian tabulate corals (Photo 38:4-5) have been studied in particular detail by Sokolov (Соколов 1951а, 1952б, 1955) and Klaamann (Клааманн 1961, 1962, 1964, 1966, 1983). The monographic studies of the above researchers formed the basis for the Estonian tabulate coral taxonomy.

The earliest fossil of possible tabulate corals was reported from the Lower Cambrian Moorowie Formation of South Australia (Fuller & Jenkins 1994). Genuine tabulates with the genus *Cryptolichenaria* appeared in the Early Ordovician of Texas, Pennsylvania and Siberia (Kaljo & Klaamann 1973).

In the Middle Ordovician, tabulates were of wider distribution. Lichenariids and tetradiids were common for Sibera and North America. The Late Ordovician was characterized

Table 31. Stratigraphical ranges of higher stromatoporoid taxa in Estonian sequences



------ total stratigraphical range, —— stratigraphical range in Estonian sequence, < > continuation of distribution. Indices of stages: D_{III} - Oandu, E - Rakvere, F_1a - Nabala, F_1b - Vormsi, F_1c - Pirgu, F_{II} - Porkuni, G_{1-2} - Juuru, G_3 - Raikküla, H - Adavere, J_1 - Jaani, J_2 - Jaagarahu, K_1 - Rootsiküla, K_2 - Paadla, K_3a - Kuressaare, K_3b - Kaugatuma, K_4 - Ohesaare.

Photo 38. Ordovician and Silurian stromatoporoids and corals:

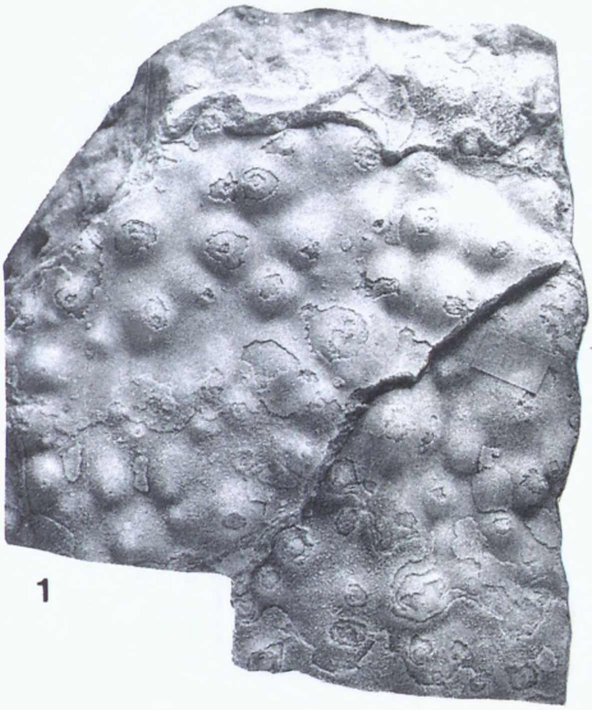
1-3. Stromatoporoid *Clathrodictyon mammillatum* (Schmidt), lectotype Co3002, top Ordovician, Porkuni Stage, Porkuni quarry: 1- upper surface, nat. size; 2- vertical section of skeleton, x 10; 3 - horizontal section, x 10.

4. Tabulate coral *Catenipora tapaensis* (Sokolov), Pirgu Stage, Tapa, nat. size.

5. Tabulate coral *Favosites* aff. *gothlandicus*, upper surface, Llandovery, Raikküla Stage, Mündi quarry, x 0.7.

6. Rugose coral *Kodonophyllum* sp., calyx, Ludlow, Paadla Stage, Unimäe, x 1.2.

7. Rugose coral *Phaulactis* sp., side view, Wenlock, Jaagarahu Stage, Sepise, nat. size.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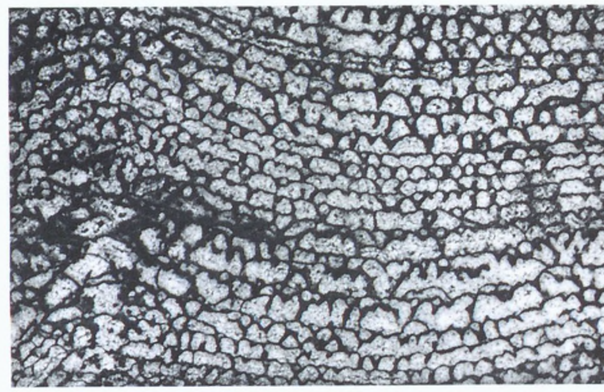
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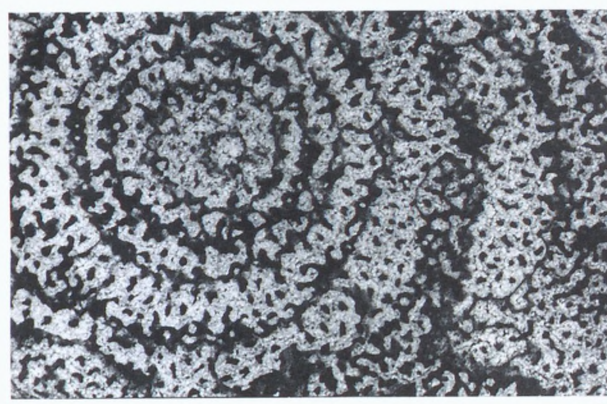
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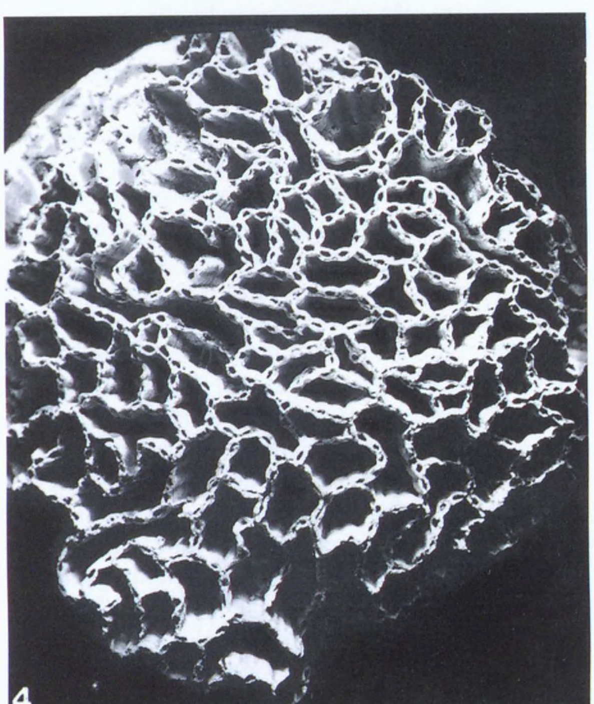
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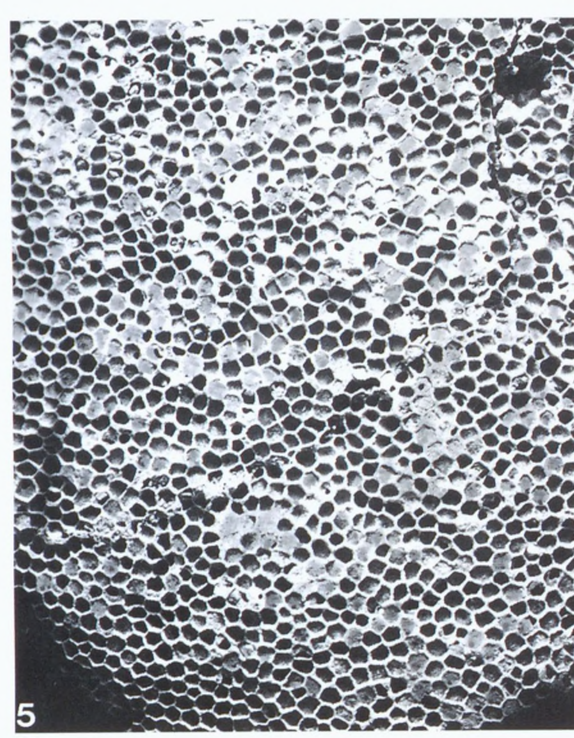
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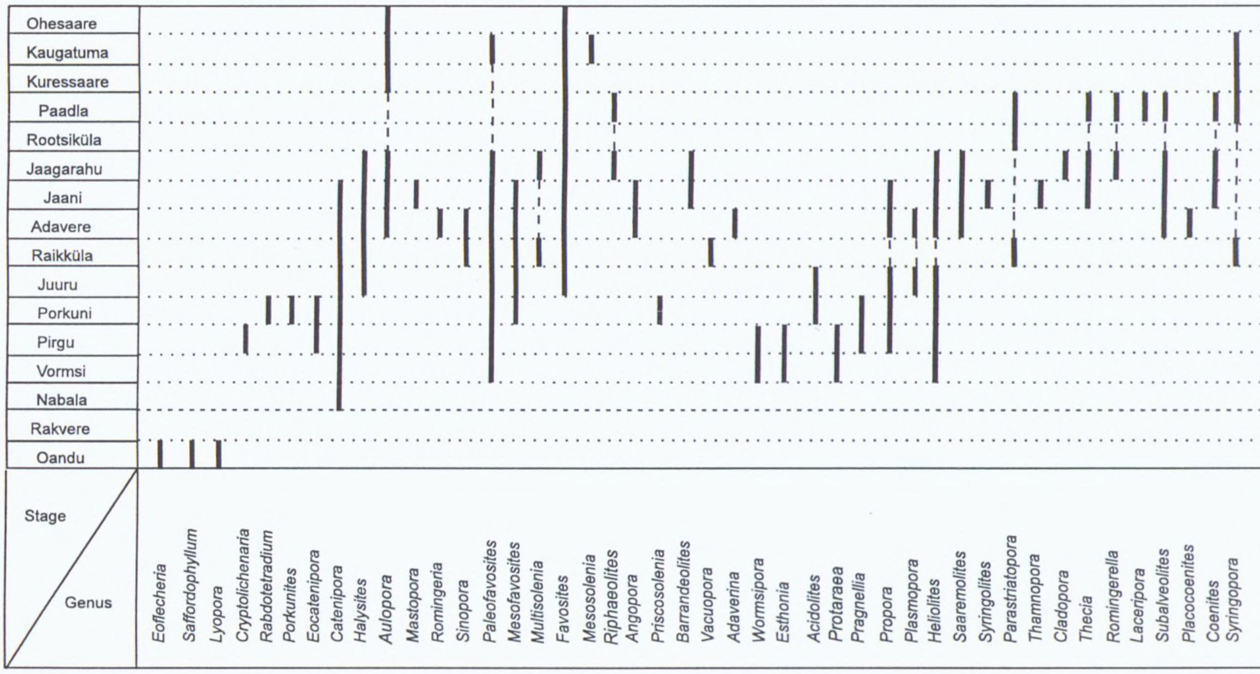
Table 32. Ranges of tabulate coral genera in the sequence of Estonia after Klaamann (Клааманн **1966, 1970б) with some modifications**



by a greater variety of tabulates; favositids, heliolitids, tetradiids, sarcinulids were widespread in different parts of the world. In the Silurian and Devonian, a great diversification of tabulate corals took place. In the Silurian the most common were favositids, halysitids, heliolitids, theciids and syringoporids, reaching the maximum diversity in the Wenlock. Halysitids disappeared at the end of the Silurian. Some new families, including Micheliniidae, Cleistoporidae, Trachyporidae, appeared already at the end of the Silurian, but became widespread in the middle of the Devonian together with pachyporids, alveolitids and coenitids. Big differences occurred also on the generic level. At the end of the Devonian, nearly all subfamilies of favositids disappeared as did all theciids, syringolitids, alveolitids and also the order *Heliolitida*. In the Carboniferous, auloporids, micheliniids, cleistoporids and palaeacids were well diversified with syringoporids dominating. The Permian fauna did not differ much from the Carboniferous one, but was poorer and less widely distributed. Tabulate corals became extinct at the end of the Late Permian.

Several palaeobiogeographic faunal regions - Americo-Siberian, Central-Asian and European (including the Baltic area), have been distinguished (Kaljo & Klaamann 1973). Due to the favourable climatic and shallow-water conditions in the Estonian part of the Baltic Basin, tabulates played a significant role in the Late Ordovician and Silurian faunal assemblages.

In Estonia, the earliest tabulates are known from the Late Caradoc. *Lyopora tulaensis* Sok., *Saffordophyllum grande* Sok., *Eoflecheria orvikui* Sok., occur in the Vasalemma reef facies of the Oandu Stage (Table 32). At the beginning of the Late Ordovician, the conditions for tabulates were unfavourable, therefore the fossils are rare. Only one species, *Catenipora obliqua* (Fischer - Benzon) has been recorded from the Nabala Stage. Diversification of the Late Ordovician tabulate fauna began in the Vormsi Age with the appearance of the first *Paleofavosites* - *P. schmidti* Sok. and *P. borealis* Tshern. The most ancient heliolitids (*Wormsipora, Esthonia, Protaraea*) are also known from the Vormsi Stage. Sarcinuliids and halysitids (*Catenipora*) became common for the first time. During the Pirgu Age diversification continued. The tetradiids (*Cryptolichenaria multiplex* Klaam.) appeared for the first time in the Baltic area. The halysitid *Eocatenipora* was widely distributed and early heliolitids were well developed. The Late Ordovician favositids were blooming in the Porkuni Age. The first records of *Porkunites, Mesofavosites* and *Priscosolenia* are from the same age.

The beginning of the Silurian was a time of rapid diversification of *Mesofavosites* and *Paleofavosites*. In the middle of Llandovery, *Multisolenia* and *Parastriapora* were added to *Favosites* and favositids became the dominant group of tabulates in the Silurian. The Juuru Stage is characterized by a few genera of tabulates. *Paleofavosites* and *Mesofavosites* were the most common genera at that time. *Catenipora* was common at the beginning of the age.The oldest representatives of the genera *Halysites* (*H. priscus* Klaam.) and *Favosites* (*F.antiquus* Sok.) appeared. In the Raikküla Stage representatives of the genus *Paleofavosites* are less numerous than in the Juuru Stage whereas *Favosites* is more abundant. Such genera as *Parastriatopora, Multisolenia, Syringopora, Vacuopora* and *Sinopora* appeared at that time. In the Adavere Stage the assemblage of tabulates is more diverse, but less endemic than the Raikküla assemblage. The first alveolitids and coenitids occur, auloporids and halysitids (*Catenipora*) are widespread. The morphology of the tabulates in the Adavere Stage is quite different from those appearing in the Wenlock.

The Jaani Age was characterized by major changes in the tabulate fauna: *Syringolites, Thecia* and *Mastopora* appeared first at the end of this age; *Mesofavosites* and *Catenipora* disappeared. *Paleofavosites, Mesofavosites, Favosites, Catenipora* and *Halysites* were poorly represented in the Jaani fauna. During the Jaagarahu Age, the diversity of tabulates rose again. That can be explained by more favourable enviromental conditions in a widespread shoal facies. The new genera *Cladopora* and *Romingerella* appeared at that time. In the Jaagarahu fauna the typical Silurian tabulate genera were almost fully represented, but at the end of the age *Multisolenia* and *Halysites* disappeared from the Estonian sequence, which preceeded their disappearance in surrounding areas. The Rootsiküla Stage is characterized by rare *Favosites* and an abundance of *Parastriatopora commutabilis,* specific to the stage.

In the Ludlow, tabulates had a low generic diversity. Of those, *Favosites* was important. A few representatives of *Thecia*, *Romingerella, Laceripora* and *Syringopora* have been found from the Paadla Stage. *Favosites* was more diverse than in the Rootsiküla Stage. Most tabulates disappeared at the end of the Paadla Age. The Kuressaare Stage has a very poor record of tabulates and only a few species of *Favosites* and *Aulopora* have been found. No changes at the generic level took place at the Ludlow - Přídoli transition. The last few species of *Paleofavosites* occur in the Kaugatuma Stage, whereas *Favosites* is quite common. *Syringopora* and *Mesosolenia* are rare. The tabulate fauna of the Ohesaare Stage does not differ significantly from that of the Kaugatuma Stage and only a few species have been recorded.

The Estonian tabulate faunas reveal quite clear differentiation into lateral communities (Клааманн 1986) as shown in Table 33. The formation of communities was influenced by the water depth, hydrodynamics and other factors. The depth of water determined the species composition of communities, while the other agents controlled the shape of coral colonies and the diversity and number of lateral communities. The more the enviromental conditions differentiated, the greater the number of lateral communities was.

Rugose corals

Knowledge of the Ordovician and Silurian rugose corals of Estonia (Photo 38:6-7) is mainly based on the studies by Eichwald (1854-60), Dybowski (1873/4), Weissermel (1894), Reiman (Рейман 1956, 1958) and Kaljo (1961, 1996, Кальо 1956, 1958, 1970б). During the recent decades, Neuman (1969, 1986), Scrutton (1988) and Weyer (1973, 1982, 1983, 1993) have published several papers describing only a few new taxa but improving considerably the taxonomy of corals identified earlier. The number of the known species-level taxa, slightly exceeds one hundred, but the share of undescribed forms might be at least 20-30%.

Rugose corals made their first appearance in the Middle Ordovician of North America. In Estonia, they are represented by *Primitophyllum primum* Kaljo and *Lambelasma dybowskii* (Kaljo) occurring in the Haljala Stage and undoubtedly having the habitus of the most primitive tetracorals. In general, the Ordovician rugose coral assemblages were dominated by simple streptelasmatid corals provided only with tabulae between septa and often having a dilated septal apparatus. The first corals with well developed dissepimentarium appeared at the very end of the Late Ordovician and gained predominance later in the Silurian. The Ordovician Period, however, ended with a serious extinction of corals (first of all species- and genus-level taxa, particularly streptelasmatids), and the earliest Silurian (Rhuddanian) was a low-diversity period

Table 33. Succession of tabulate coral communities in the East-Baltic Silurian (after Клааманн 1986, Kaljo 1990a)

Series	Regio-stage	Lagoon	Shoal: inner … outer	Open shelf	Transition	Depression
Pridoli	K_4		*Favosites ohesaarensis - F. effusus*			No tabulate corals
	K_3b		*Syringopora blanda Favosites muratsiensis*	„*F.*" *eichwaldi*		
Ludlow	K_3a		„*Paleofavosites*" *moribundus Favosites forbesi*			
	K_2		*Laceripora cribrosa-Parastriatopora coreaniformis*; *Thecia swindereniana-Favosites subgothlandicus*	*Dnestrites*; *Halysites crassus*	*Halysites laticatenatus - Favosites gothlandicus*	
Wenlock	K_1	*Parastriatopora commutabilis*	*Paleofavosites tersus - Halysites klintebergensis*; *Paleofavosites asper*		*Favosites gothlandicus*	
	J_2	*Riphaeolites lamelliformis*	*Halysites junior - Paleofavosites tersus*; *Favosites mirandus*			
	J_1		*Halysites senior*	*Angopora hisingeri*		
Llandovery	H		*Mesofavosites obliquus - Favosites favosus*			
	G_3		*Parastriatopora celebrata*			
	G_{I-II}		*Paleofavosites karinuensis*; *Mesofavosites fleximurinus - Paleofavosites paulus*; *Acidolites lateseptatus*	*Catenipora martinssoni*		

Indices of stages: G_{I-II} - Juuru, G_3 - Raikküla, H - Adavere, J_1 - Jaani, J_2 - Jaagarahu, K_1 - Rootsiküla, K_2 - Paadla, K_3a - Kuressaare, K_3b - Kaugatuma, K_4 - Ohesaare.

dominated by Ordovician carry-overs. Later, a stepwise increase in the diversity followed until the maximum was reached in the Wenlock (Kaljo 1996). Morphological differentiation was remarkable. New types of septa, stereozones, calices, many colonial forms, *etc.* appeared which formed a base for taxonomical diversity. The most characteristic were different cystiphyllids, kodonophyllids, entelophyllids, lykophyllids, arachnophyllids, *etc.* The Late Silurian shows a decline of rugose corals in general, but a few new elements appeared in the Přidoli, among them the so-called "Devonian" elements (*Acanthophyllum, Lyrielasma, etc.*, Scrutton, 1988).

The above general evolutionary pattern is well observable in Estonia. Apart from the above-mentioned primitive rugosans, *Kenophyllym* and *Streptelasma* appeared in the Keila Age, and *Borelasma* and the first tryplasmatid *Estonielasma*, in the Oandu Age. The first *Grewingkia* was identified at the end of the Middle Ordovician. It means that rugose corals were scarce in the Middle Ordovician of Estonia, but their diversity was already comparatively high.

The Late Ordovician was mostly dominated by streptelasmatids (*Kenophyllum, Streptelasma, Grewingkia, Helicelasma, Dalmanophyllum*), but there occurred also rare lambelasmatids or calostylids s. l.: *Coelostylis* (*Vormsistylis*), *Neotryplasma, Calostylis, Estonielasma.* The end of the period (Porkuni Age) was marked by the incoming of the first paliphyllids (*Paliphyllum, Strombodes*) and staurids (*Palaeophyllum*).

The Silurian rugose corals in Estonia are the most diversified in the following stratigraphical units: (1) the upper Aeronian Rumba Formation (*Dinophyllum, Entelophyllum, Prodarwinia, Phaulactis, etc.*); (2) the Middle Wenlock Jaagarahu Formation (*Acervularia, Spongophylloides, Microplasma, etc.*); (3) the Ludlow - Přidoli (*Entelophyllum* and *Tryplasma* were most common, but in the Kaugatuma Stage also *Cystiphyllum, Holmophyllum, Strombodes* and the first representatives of *Acanthophyllum* appeared).

The distribution of these corals shows a distinct facies control. Reliable records of rugose corals from the Silurian lagoonal and shelf depression facies are lacking. These corals were scarce also in the *Borealis* and *Pentamerus* banks and stromatoporoid biostromes, but rich assemblages occurred in the reefs and their surroundings (*e.g.* Hilliste reefs of the Juuru and Raikküla stages, Sepise outcrop of the Jaagarahu Stage, *etc.*). A diverse assemblage of rugose corals occurred also in the shallow part of the open shelf. However, the share of solitary corals in it was higher than in reef environments, and the role of colonial corals decreased. Up to now, only a few species have been identified from the deeper (outer) shelf (*Porpites porpita* from the Velise Formation, *etc.*), but many new taxa have not been described yet.

By now, no suggestions for the biozonations of rugose corals have been made, but Kaljo (1961,1996, Кальо 1970б) has listed the characteristic species for stratigraphic units.

Biogeographically, Estonian rugose corals belonged to the Baltoscandian (or North European) Province which had some connections with the North American - Siberian and also with the Middle Asian provinces. These connections, as well as the share of the widely distributed and endemic corals, were changing during the time under discussion. The importance of endemic corals was relatively high before the Wenlock.

Inarticulate brachiopods

Inarticulate brachiopods with calcium phosphate and calcitic shell, formerly referred to the class Inarticulata, have been recently assigned to three new subphyla, and several classes (Williams *et al.* 1996). The representatives of the subphylum Linguliformea, formerly known as phosphatic inarticulate brachiopods, are abundant in the Cambrian and Lower Ordovician of Estonia, common in the Middle and Upper Ordovician, rare in the Silurian and rather common in the Devonian. The representatives of the subphylum Craniiformea are known from the Ordovician and Silurian.

These brachiopods have been studied since the early 19th century. In their earlier works Eichwald, Pander, Mickwitz, Kutorga, Huene, Walcott, Bekker and several other authors described many new species and genera from the Cambrian and Ordovician of Estonia (see Puura 1990 for review). More recent taxonomical studies of the East Baltic and Scandinavian Cambrian and Ordovician faunas, discussing the systematic position of Estonian taxa, include Gorjansky (Горянский 1969), Biernat (1973), Holmer (1986, 1989), Popov and Nõlvak (1987), Popov and Khazanovitch (Попов и Хазанович 1989), Holmer and Popov (1990), Puura and Holmer (1993), Popov *et al.* (1994), Holmer and Popov (1994) and Popov and Holmer (1994). Rare Silurian and Devonian lingulate brachiopods have been described by Popov (Попов 1981) and Gravitis (Гравитис 1981), respectively.

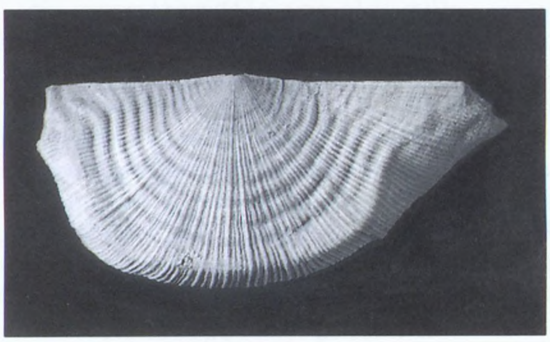
The earliest brachiopods known from Estonia, *Mickwitzia monilifera* (Linnarsson) (Photo 39:1) and *Paterina rara* Gorjansky from the Lower Cambrian, are tentatively assigned to the class Paterinata.

The representatives of the class Lingulata are most abundant in the Upper Cambrian and the lower part of the Pakerort Stage where obolid coquinas form the deposits of shelly phosphorites. In this interval, the successive assemblages have been used for defining four lingulate zones: *Ungula inornata* (Mickwitz), *U. convexa* Pander, *U. ingrica* (Eichwald) and *Obolus apollinis* Eichwald (Попов и Хазанович 1989). These zonal species, alongside with more than ten other species, apparently restricted to Baltoscandian Basin, can be used for

Photo 39. Cambrian to Silurian brachiopods:
1. Inarticulate brachiopod *Mickwitzia monilifera* (Linnarsson), L. Cambrian, Tiskre Fm., Mustametsa, x 2.5.
2. Articulate brachiopod *Vellamo verneuili* (Eichwald), Br 474, interior of dorsal valve, U. Ordovician, Vormsi Stage, Kõrgessaare, x 2.
3. Articulate brachiopod *Leptaena* sp., ventral valve, U. Ordovician, Vormsi Stage, Kärrslätt, x 1.2.
4. Articulate brachiopod *Cyrtonotella kuckersiana* Öpik, dorsal valve, M. Ordovician, Kukruse Stage, Küttejõu, x 1.2.
5. Articulate brachiopod *Ilmarinia sinuata* (Pahlen), ventral view, U. Ordovician, Pirgu Stage, Ambla, x 1.2.
6. Articulate brachiopod *Pentamerus oblongus* (Sowerby), Llandovery, Adavere Stage, Vändra, x 0.8.
7. Coquina of *Atrypoidea prunum* (Dalman), Ludlow, Kuressaare Stage, Kuressaare, x0.8.

1

2



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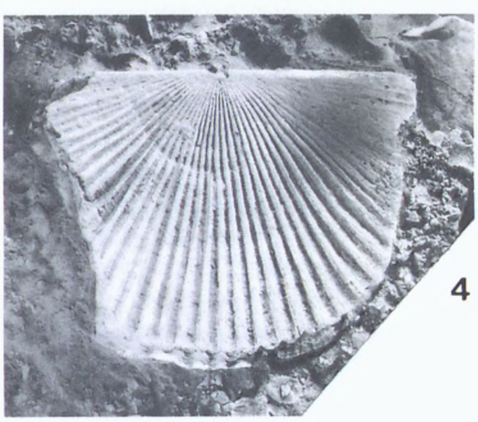

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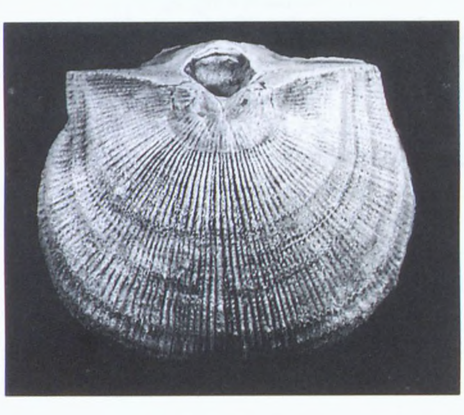

7

correlations across the basin from Sweden to Lake Ladoga (Попов и Хазанович 1989, Holmer & Popov 1990, Puura & Holmer 1993).

In the kerogenous *Dictyonema* Shale of the upper part of the Pakerort Stage, rare lingulate brachiopods *Eurytreta* sp. and *Lingulella* aff. *L. tetragona* Gorjansky occur.

The Hunneberg Stage is characterized by an assemblage of more than ten lingulate species dominated by *Thysanotos siluricus* (Eichwald) and *Leptembolon lingulaeformis* (Mickwitz). In the Billingen Stage, acrotretids *Acrotreta subconica* Kutorga and *Myotreta crassa* Gorjansky occur.

The characteristic assemblage of the Volkhov Stage includes *Acrotreta tallinnensis* Holmer and Popov, *Rowellella rugosa* Gorjansky and *Myotreta estoniana* (Biernat). *Myotreta crassa* Gorjansky, *Biernatia rossica* (Gorjansky) and *Eosiphonotreta verrucosa* (Eichwald) range from the Volkhov to the Kunda Stage, while *Eoconulus cryptomyus* (Gorjansky) ranges from the Volkhov to the Aseri Stage and *Conotreta mica* (Gorjansky) from the uppermost Kunda to the lowermost Uhaku Stage. *Siphonotreta unguiculata* (Eichwald) ranges from the Aseri to the Uhaku Stage.

In the Kukruse Stage lingulate brachiopods are represented by *Biernatia holmi* Holmer and *Schizotreta elliptica* (Kutorga). Carbonate-shelled craniate brachiopods (class Craniata) are represented by *Philhedra baltica* Koken, *Orthisocrania planissima* (Eichwald) and about ten more species assigned to the genera *Philhedra, Orthisocrania, Pseudopholidops* and *Paracraniops*.

In the Haljala Stage, the lingulate *Alichovia ramispinosa* Gorjansky and the craniates *Philhedra metatypotheisa* Huene, *Orthisocrania curvicostae* Huene occur. *O. depressa* (Eichwald) ranges from the Jõhvi Substage of the Haljala Stage to the Keila Stage and *Philhedra kegelensis* Huene from the Keila Stage to the lowermost Oandu Stage.

From the Nabala Stage, Gorjansky (Горянский 1969) has reported *Pseudolingula quadrata* (Eichwald) and *Lingulops mirus* Gorjansky. An assemblage from the Vormsi and Pirgu stages of the Viljandi core includes *Acanthambonia portranensis* Wright, *Rowellella minuta* Wright, *Spondylotreta* cf. *parva* Wright, *Eoconulus semiregularis* Biernat, *Paterula* sp. and *Schizotreta* sp. (Popov & Nõlvak 1987, Popov *et al.* 1994).

The class Craniata is represented in the upper half of the Ordovician by *Pseudopholidops stolleyana* in the Oandu Stage and by about ten species assigned to the genera *Philhedrella?* (Rakvere and Nabala stages), *Petrocrania?* (Vormsi Stage) and *Pseudocrania* (Porkuni Stage).

The only two lingulate species so far described from the Silurian of Estonia include *Opsiconidion aldridgei* (Cocks) from the Raikküla and Jaani stages and *Eschatelasma rugosum* Popov from the Jaani Stage (Попов 1981).

From the Devonian of Estonia, Gravitis (Гравитис 1981), has reported lingulate brachiopods *Bicarinatina bicarinata* (Kutorga), *B. ugalana* Gravitis from the Aruküla Stage and *B. sakalana* Gravitis from the Narva Stage. The latter two species have also been reported from some Latvian sections.

Most of the species discussed above are restricted to the East Baltic and Scandinavia, but some lingulate species have a wider distribution and are of interest for biogeographic and palaeogeographic studies. For instance, the Upper Cambrian species *Angulotreta postapicalis* is known, except for Estonia, also from Novaya Zemlya and North America (see Puura 1990). *Thysanotos siluricus* and *Leptembolon lingulaeformis*, typical of the Hunneberg Stage in Estonia, are known from about the equivalent stratigraphic level from the South Urals, Poland and Bohemia (Popov & Holmer 1994). The Upper Ordovician *Acanthambonia portranensis* and *Rowellella minuta* are known from the Upper Ordovician of Ireland (Wright 1963) and the Lower Silurian of Wales and England (Cocks 1979).

Ordovician articulate brachiopods

Reseach into articulate brachiopods, which form one of the main groups in the Ordovician shelly fauna (Photo 39:2-5), was started in northern Estonia and adjacent areas in the 19th century (Pander 1830, Eichwald 1860) and continued by Öpik (1930b, 1934, a.o.), Alikhova (Алихова 1951, 1953) Rõõmusoks (1989, Рыымусокс 1956б, 1981 a.o.), Rubel (Рубель 1961), Oraspõld (Ораспыылд 1956) and others. As a result, more than 300 species of Articulata have been described. They belong to about 130 genera, among which the representatives of the orders Orthida and Strophomenida dominated during the Early and Middle Ordovician (Table 34). The orders Pentamerida, Rhynchonellida and Spiriferida represent mainly the latest Middle and Upper Ordovician. The occurrence of numerous endemic elements in the Baltic Basin has enabled to distinguish a separate Baltic Province (Williams 1973) or a specific Baltoscandian fauna (Jaanusson 1973b). During the Ordovician, the changes in the composition of brachiopod fauna were caused in a great deal by a succesive decrease of the endemic elements and increase of the immigrants from the different faunal provinces (Рыымусокс 1967, 1970). The dynamics and comparison of different contemporaneous Ordovician brachiopod faunas, including the Baltoscandian, has been analysed by Jaanusson (1973b, 1976, 1979, 1984). Reviews on the brachiopod fauna in Estonia and data on the stratigraphical distribution of species have been presented in several publications (Мянниль 1966, Мянниль и др. 1966, Hints *et al.* 1989, Hints 1990), in particular detail by Rõõmusoks (1983, Рыымусокс 1967, 1970).

Different brachiopod associations are characteristic of northern and southern Estonia belonging to different confacies belts (Fig. 24). The richest and diversest brachiopod fauna has been described from the variably argillaceous limestones in northern Estonia. The micritic (aphanitic) limestones (Rägavere and Saunja formations) comprise relatively few brachiopods, which mainly occur in the argillaceous interbeds. An impoverished brachiopod fauna is characteristic of the red-coloured units (the whole Lower Ordovician, Jonstorp Formation) in southern Estonia.

The earliest articulate brachiopods (orthids *Prantlina, Panderia, Ranorthis*, dalmanellids *Paurorthis*, plectambonitids *Plectella* a. o.) appear in the glauconitic sandstones of the Mäeküla Member in the lower part of the Billingen Stage (see Table 34). In the carbonate rocks of the Volkhov and Kunda stages, this earliest short-living brachiopod fauna is replaced by a new fauna comprising the orthids *Productorthis, Orthambonites. Orthis*, plectambonitids *Ahtiella, Ingria* and taxa of Clitambonitidina

Table 34. Distribution of articulate brachiopods in northern Estonia

Series / Stage / Genus	Oeland				Viru								Harju			
	Hunneberg	Billingen	Volkhov	Kunda	Aseri	Lasnamägi	Uhaku	Kukruse	Haljala	Keila	Oandu	Rakvere	Nabala	Vormsi	Pirgu	Porkuni
ORTHIDA																
Apheoorthina?		x														
Paurorthina		x														
Prantlina		x														
Panderina		x	x													
Paurorthis		x	x	x				x								
Ranorthis		x	x	x												
Gonambonites		?	x	x												
Antigonambonites		?	x	x												
Productorthis			x	x												
Glossorthis			x	x		x	x	x								
Apomatella			x	?												
Anchigonites			x													
Nothorthis			x													
Neumania			x													
Raunites			x													
Paralenorthis			x	x												
Cyrtonotella			x	x				x	x	x						
Nicolella			x	x		x	x	x	x	x	x	x	x	x	x	
Platystrophia			x	x		x	x	x	x	x	x	x	x	x	x	x
Ladogiella			?	x	x											
Hemipronites			?	x												
Progonambonites			?	x												
Estlandia				x	x	x	x	x	x	x						
Clitambonites				x	x	x	x	x	x							
Lacunarites				x		x										
Krattorthis				x	x											
Iru				x												
Orthis				x												
Orthambonites				x												
Oslogonites				x												
Cremnorthis						x	x	?								
Apatorthis						x	x		x	x	x				x	
Hesperorthis						?	x	x	x	x		x		x		
Paucicrura						?	x	x					x	x		
Vellamo							x	x	x	x	x	x	x	x	x	x
Kullervo							?	x	x					x		
"Schizoramma"								x								
"Orthis"									x							
Saukrodictya									x	x	x					
Clinambon									x	x	x					
Glyptorthis										x	x			x	x	
Skenidioides										x	X		x			
"Onniella"										x	x				x	
Horderleyella?										x						
Boreadorthis											x	x	x	x	x	
Ilmarinia											x		x	x	x	x
Reushella											C					x
Oanduporella											x					
Rhactorthis											x					
Plaesiomys													x	x	x	
Sulevorthis													x	x	x	
Epitomyonia													x	x		

Series / Stage / Genus	Oeland				Viru								Harju			
	Hunneberg	Billingen	Volkhov	Kunda	Aseri	Lasnamägi	Uhaku	Kukruse	Haljala	Keila	Oandu	Rakvere	Nabala	Vormsi	Pirgu	Porkuni
Pionodema													x			
Wysogorskiella													?	x		
Dicoelosia														x	x	
Laticrura														x	x	
Mendacella														x		
Isorthis															x	
Resserella															x	
Hebertella															x	
Ptychopleurella															x	
Barbarorthis															x	x
Elsaella																x
INCERTI ORDIS																
Oxoplecia							x	x					x	x		x
Triplesia								x				x	x	x	x	
Onycoplecia												x				
Ogmoplecia													x	x	x	?
Streptis																x
PENTAMERIDA																
Angusticardinia		x	x													
Porambonites		x	x	x			x	x								
Lycophoria			x	x	x											
Equirostra					x	x	x	x	x	x	x	x	x	x	x	
Noetlingia					x											
Camerella											x					
Stenocamara ?															x	
Holorhynchus															x	
RHYNCHONELLIDA																
Rostricellula											x					
Hypsiptycha															x	
Rhynchotrema																x
ATRYPIDA																
Anazyga											x					
Cyclospira													x			
Catazyga														x	x	
Atrypidae n. gen.														x	x	
Eospirigerina															x	x
ATHYRIDA																
Hindella															x	
Cryptothyrella																x
STROPHOMENIDA																
Plectella		x														
Onegia			x													
Ingria			x	x												
Ahtiella			?	x												
Ukoa				x	x											
Plectambonites				x	x		x									
Inversella (Inversella)				x	x											

Series / Stage / Genus	Oeland				Viru								Harju			
	Hunneberg	Billingen	Volkhov	Kunda	Aseri	Lasnamägi	Uhaku	Kukruse	Haljala	Keila	Oandu	Rakvere	Nabala	Vormsi	Pirgu	Porkuni
Panderites				x	x											
Leptoplium					x	x	x	x								
Leptestia					x	x	x	x								
Christiania					x	x										
Tallinnites						x	x									
Leptelloidea						x	x	x								
Estonomena						x	x	x	x							
Bekkerina						x	x	x	x	x						
Sowerbyella (Viruella)						?	x	x								
Palaeostrophomena							x									
Septomena							x	x								
Kierulfina							x	x	x							
Actinomena							x	x								
Tetraodontella								x								
Bilobia								x	x							
Eoplectodonta									x				x	x	x	
Astamena									x							
Kurnamena									x	x						
Kiaeromena									x		x					
Sowerbyella (Sowerbyella)									x	x	x	x				
Longvillia										x						
Keilamena										x						
Dactylogonia											x					
Oandumena											x					
Kjaerina											x				x	x
Rakverina											x	x				
Holtedahlina											x	x	x		x	
Pseudostrophomena											x	x				
Microtrypa											x	x				
Sampo											?	x	x	x	x	
Geniculina												x	x	x	x	
Hedstroemina												x				
Sowerbyella (Eochonetes)													x	x	x	x
Luhaia															x	
Trigrammaria													x			
Harjumena													x	x		
Bekkeromena													x	x	x	
"Fardenia"													x	x	x	
Similoleptaena														x	x	
Pirgumena															x	x
Rugomena															x	
Anoptambonites															x	
Luhaia															x	
Schmidtomena																x

Data on Strophomenida by A. Rõõmusoks

(*Antigonambonites*, *Gonambonites*, *Progonambonites)*, mostly endemics in the Baltic Basin.

At the beginning of the Middle Ordovician, the last Early Ordovician (Oelandian) brachiopods (*Ladogiella, Lycophoria*) became extinct and the Middle Ordovician (Viruan) brachiopod fauna, which comprised several new strophomenids (*Christiania, Leptestia),* started to form. Somewhat later, several new taxa characteristic of the lower half of the Viru Series (*Sowerbyella, Leptelloidea*, *Hesperorthis, Clitambonites*, a.o. were added (Table 34). Several of these brachiopods have been found in the core sections which allows to suppose their wide distribution within the North Estonian Confacies Belt (Fig. 32). From southern Estonia, the early and middle Viru brachiopod fauna is insufficiently known. *Alwynella, Bimuria, Sampo*? (?=*Leptellina*) serve as Scandinavian faunal elements.

A remarkable renovation of the brachiopod fauna took place at the Keila - Oandu transition coinciding with the period of essential environmental changes in the basin. Most of the brachiopod species and many genera, existing earlier in the basin, disappeared at the end of Keila time (see Рыымусокс 1970). The new, post-Keila time brachiopod fauna comprised several taxa which had immigrated from North America and North Europa (*Rynchotrema, Rostricellula, Camerella, Dactylogonia*; Рыымусокс 1967). Some of those (*e.g. Reushella*) have been found only in central Estonia. In southern Estonia, the changes in the composition of the brachiopod fauna are not clear. However, since Keila - Oandu time, the occurrence and diversity of articulate brachiopods decreased notably, especially in the black shales (Keila - Oandu and Vormsi stages) and red-coloured limestones (Pirgu Stage).

In northern Estonia, the late Middle and Upper Ordovician brachiopod faunas were dominated by relatively large-shell brachiopods (*Plaesiomys, Platystrophia, Vellamo, Triplesia, Bekkeromena, "Leptaena"* and others). After gradual appearance of *Dicoelosia* (Vormsi Age) and *Eospirigerina* (Pirgu Age), and presumably also of *Holorhynchus* (latest Pirgu Age), the Baltic fauna became close to the Hiberno-Salairian fauna (Jaanusson 1979). At the same time, several descendants of endemic brachiopods occur (*Ilmarinia, Apatorthis*, *Equirostra*). The number of brachiopod genera decreased towards the end of the Ordovician. Thus, in the Vormsi and Pirgu stages, the brachiopods are represented by more than 35 genera, about half of which range over into the Porkuni Stage where they are associated with a few new taxa. *Streptis* and *Meristina* can be treated as really new taxa, whereas some other genera (*Reushella, Laticrura, Rhynchotrema*) recurred after their first entrance in Estonia during the late Viru time. Articulated brachiopods are quite common fossils in the reef complex of the Porkuni Stage (main part of the Ärina Formation) in spite of their decreased diversity. This brachiopod fauna, named tentatively the *Streptis* association (Hints 1993), disappears at the end of the Ordovician. Up to now, there are only two Ordovician species *Onniella trigona* Rubel and *Eospirigerina porkuniensis Rubel* which have been identified also from the lowermost Silurian.

During the Porkuni Age, the so-called *Hirnantia* fauna, described from different Ordovician basins (Rong & Harper 1988), was distributed in southern Estonia. Several typical representatives of that fauna (*Hirnantia, Dalmanella, Plectothyrella, Hindella*) have been established in the Kuldiga Formation (Ruhnu, Ikla, Taagepera drill cores) which commonly represents the lower and middle parts of the Porkuni Stage in the central East Baltic (Ульст и др. 1982, Ораспыльд 1975б). Up to now, the relationships between the *Streptis* and *Hirnantia* faunas are unclear. The preliminary data on the isotopic composition of the topmost Ordovician strata suggests that these faunas may be partly contemporaneous, but the *Hirnantia* fauna has existed longer.

Silurian articulate brachiopods

The Silurian articulate brachiopods (Photo 39:6-7) were numerous and diverse in carbonate facies of the cratonic seas including the Estonian area (Table 35). Up to now, about 200 species from 98 genera have been described from the Estonian Silurian by different authors in a lot of papers, including monographs by Sokolskaya (Сокольская 1954), Rubel (Рубель 1963, 1970) and Modzałevskaya (Модзалевская 1985). At least three well-known evolutionary lineages in the genera *Stricklandia*, *Dicoelosia* and *Pentamerus* have been studied and used for the dating of rocks (Rubel 1971, Рубель 1977, Мустейкис и Пуура 1983, Johnson *et al*. 1991).

The appearance of a large number of genera at the beginning of the Silurian was connected with the expansion of many cosmopolitan Silurian brachiopods all over the world, excluding spiriferids which were lacking in the Baltic area at that time. The second increase in the number of genera in the Jaani Age was probably connected with the new transgression of the sea, tied with the global sea-level rise at that time. After the Late Wenlock crisis, there followed a gradual decrease of the Late Silurian brachiopods in Estonia reflecting the stepwise retreat of the sea.

The general impoverishment trend can be proved also by the brachiopod communities. Thus, the *Linoporella, Borealis-Pentamerus, Stricklandia-Zygospiraella* and *Meifodia-Clorinda* communities are clearly recognizable in the Llandovery; the *Stegerhynchus, Whitfieldella* and *Dicoelosia-Skenidioides* communities, widespread in the Wenlock, turned into low-diversity *Didymothyris-Salopina*, *Atrypoidea* and *Homoeospira-Delthyris* communities in the latest Silurian time (Rubel 1970, Кальо и Рубель 1982).

Bivalves and rostroconchs

The Ordovician and Silurian bivalves (Photo 40:5-6) and rostroconchs have been described in Estonia since the middle of the last century (Eichwald 1840b, 1842, 1860, Schmidt 1858, 1861, 1881). In this century, Bekker (1921) and Öpik (1930a) described some new bivalves from the Kukruse Stage. Teichert (1930) recorded an Ordovician rostroconch *Ischyrinia* from the Upper Ordovician. During the last years, some more Ordovician and Silurian bivalves of Estonia have been described (Isakar 1990, 1991, Isakar & Sinicyna 1993, Исакар 1985, Исакар и Синицына 1985, Синицына и Исакар 1987, 1992, Киселев и др.1990).

Bivalves have a rather limited stratigraphical value due to their rarity and poor preservation. They usually occur as casts. However, some species are more valuable, at least in stage-level correlation, *e.g. Ahtioconcha auris*, *Ilionia prisca* and *Grammysia obliqua*. Most of bivalve and rostroconch specimens have been collected from the North and Central Estonian confacies belts.

In Estonia, the first bivalves (some undetermined small

Table 35. Distribution of the Silurian articulate brachiopod genera

	G_{1-2}	G_3	H	J_1	J_2	K_1	K_2	K_3a	K_3b	K_4
ORTHIDA	12	12	10	13	5	0	1	2	2	1
Epitomyonia	x									
Onniella	x	x								
Mendacella	x	x								
Glyptorthis	x	x	x							
Ptychopleurella	x			x						
Linoporella	x	x		x						
Hesperorthis	x	x	x	x						
Platystrophia	x	x	x	x						
Dolerorthis	x	x	x	x	x					
Skenidioides	x	x	x	x	x					
Dicoelosia	x	x	x	x	x					
Isorthis	x	x	x	x				x	x	x
Saucrodictya		x								
Dalejina		x	x	x	x			x	x	
Dalmanella			x							
Visbyella			x	x						
Ravozetina				x						
Resserella				x	x					
Salopina				x			x			
INCERTI ORDINI	3	2	1	3	0	0	0	0	0	0
Triplesia	x	x		x						
Streptis	x		x	x						
Dictyonella	x	x		x						
STROPHOMENIDA	7	7	8	9	3	1	2	2	3	1
Furcitella	x	x								
Ygerodiscus	x	x								
Aegiria	x	x	x	x						
Coolinia	x	x	x	x						
Leangella	x	x	x	x	x					
Eoplectodonta	x	x	x	x	x					
Leptaena	x	x	x	x	x	x	x	x	x	
Pentlandina			x							
Eoamphistrophia			x							
Megastrophia		x	x							
Leptostrophia			x							
Pholidostrophia				x						
Strophonella				x						
Morinorhynchus							x	x	x	x
Shaleria									x	
CHONETIDA	0	0	1	1	0	0	1	1	1	1
Strophochonetes			x	x						
Protochonetes							x	x	x	x
PENTAMERIDA	5	5	7	4	3	2	1	0	0	0
Borealis	x	x								
Parastrophinella	x	x	x							
Stricklandia	x	x	x							
Clorinda	x	x	x	x						
Anastrophia	x	x		x						
Costistricklandia			x							
Kulumbella			x							
Pentameroides			x							
Pentamerus			x		x					
Antirhynchonella				x						
Gypidula				x	x	x	x			
Rhipidium					x					
Conchidium						x				
RHYNCONELLIDA	3	3	2	8	5	2	3	2	2	1
Thebesia	x									
Rhynchotrema	x	x								
Stegerhynchus	x	x	x	x	x		x	x	x	
Fenestrirostra		x								
Plagiorhyncha			x	x						
Eocoelia				x						
Hemitoechia				x						
Rhynchotreta				x	x					
Estonirhynchia				x	x					
Sphaerirhynchia				x	x	x	x			
Microsphaeridiorhynchia				x	x	x	x	x	x	x
ATRYPIDA	8	9	4	5	4	0	2	2	2	1
Idiospira	x									
Alispira	x									
Clintonella	x	x								
Eospirigerina	x	x								
Sypharatrypa	x	x								
Zygospiraella	x	x								
Atrypopsis	x	x								
Meifodia	x	x	x							
Zygatrypa		x								
Protatrypa		x								
Gotatrypa		x	x							
Pentlandella			x							
Glassia			x	x	x					
Septatrypa				x						
Plectatrypa				x	x					
Atrypina				x	x					
Atrypa				x	x		x	x	x	
Homoeospira							x	x	x	x
SPIRIFERIDA	0	0	2	4	3	2	5	2	2	2
Cyrtia			x	x	x					
Eospiifer			x	x	x		x			
Striispirifer				x						
Howellella				x	x	x	x	x	x	x
Janius						x	x			
Quadrithyris							x			
Delthyris							x	x	x	x
ATHYRIDA	3	4	0	3	2	1	1	0	1	2
Hindella	x	x								
Koigia	x	x								
Cyclospira	x	x								
Protozeuga		x								
Meristina				x						
Whitfieldella				x	x					
Nucleospira				x						x
Didymothyris					x	x	x			
Dayia									x	
Collarothyris										x
	41	42	35	50	25	8	16	11	13	9

nuculoids) and rostroconchs (*Ischyrinia* and *Eopteria*) appeared during the Kunda Age (Table 36). During the Aseri Age, small *Similodonta* and *Cleionychia* were added. The fact that the earliest pelecypod fauna was dominated by infaunal forms (nuculoids) suggests that the ancestral mode of life of the class was infaunal (Pojeta 1971). During the Early Ordovician, rostroconchs underwent their greatest radiation (Pojeta & Runnegar 1976). At least four rostroconch species (*Ischyrinia norvegica*, *I. triangularis* and two species of *Eopteria*) existed during the Kunda Age in Estonia.

Ahtioconcha Öpik from the Kukruse Stage is the earliest probable pteriacean known so far (Pojeta 1971). Species diversity increased considerably during the Middle Ordovician, reaching the peak (13 species) in the Kukruse Age (Fig. 156). Three genera - *Tancrediopsis*, *Dystactella*? and *Ahtioconcha*, occur only in the Kukruse Stage. In the Middle Ordovician, the common and numerous pelecypod genera (unfortunately usually badly preserved) were *Modiolopsis*, *Cleionychia*, *Plethocardia*?, *Vanuxemia*?, *Ambonychiopsis* and *Aristerella*. Ulrich (1894) described *Aristerella* as having the left valve smaller than the right. Most of the Estonian representatives of *Aristerella* have markedly right convex inequivalved shells but "*Aristerella*" from the Jõhvi Stage has an even plano-convex inequivalved shell and obviously belongs to a separate genus. In the Porkuni Stage there occurs a specific pelecypod association - *Similodonta*, *Ctenodonta* ?, some undetermined small nuculoids, *Modiolopsis*, *Mytilarca*, *Ambonychia*, *Ambonychiopsis*, *Cleionychia* ? and *Pterianea*. The small rostroconch *Hippocardia* with fine hood and long tubular rostrum occurs also in the Porkuni Stage.

Bivalves have not been found from the Landovery. This may be due to unsuitable conditions for preservation of the shells. The bivalves association of the Jaani Stage is dominated by small nuculoids - *Nuculoidea*, *Deceptrix* ?, *Praectenodonta* and *Orthonota*, *Praecardium*, *Grammysia* (the latter comes from the Nässuma borehole, see Fig. 3 - *223*). A massive, thick-shelled *Megalomus* formed banks in the shoal deposits of the Jaagarahu Age. *Mytilarca* and *Modiolopsis*

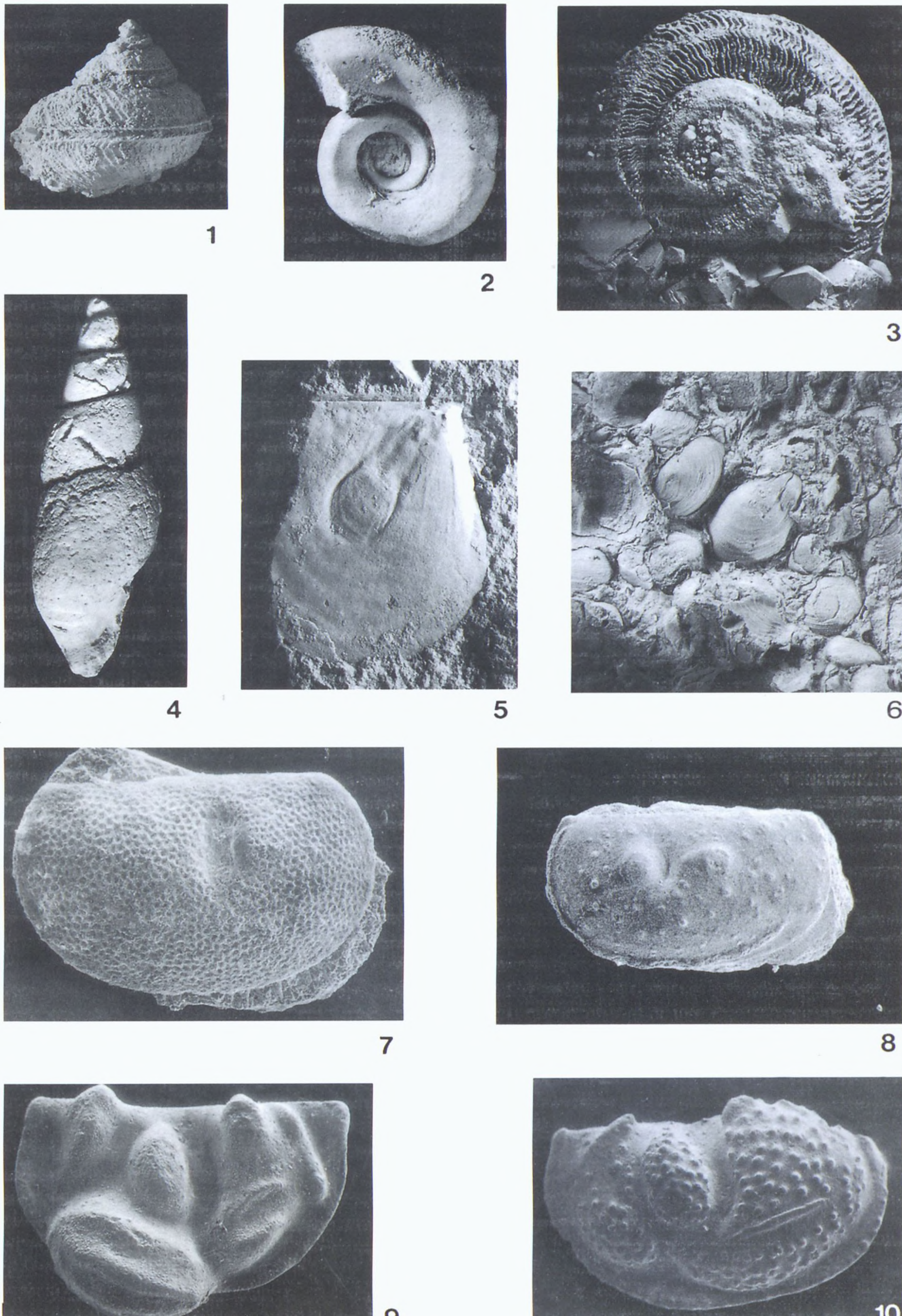
1
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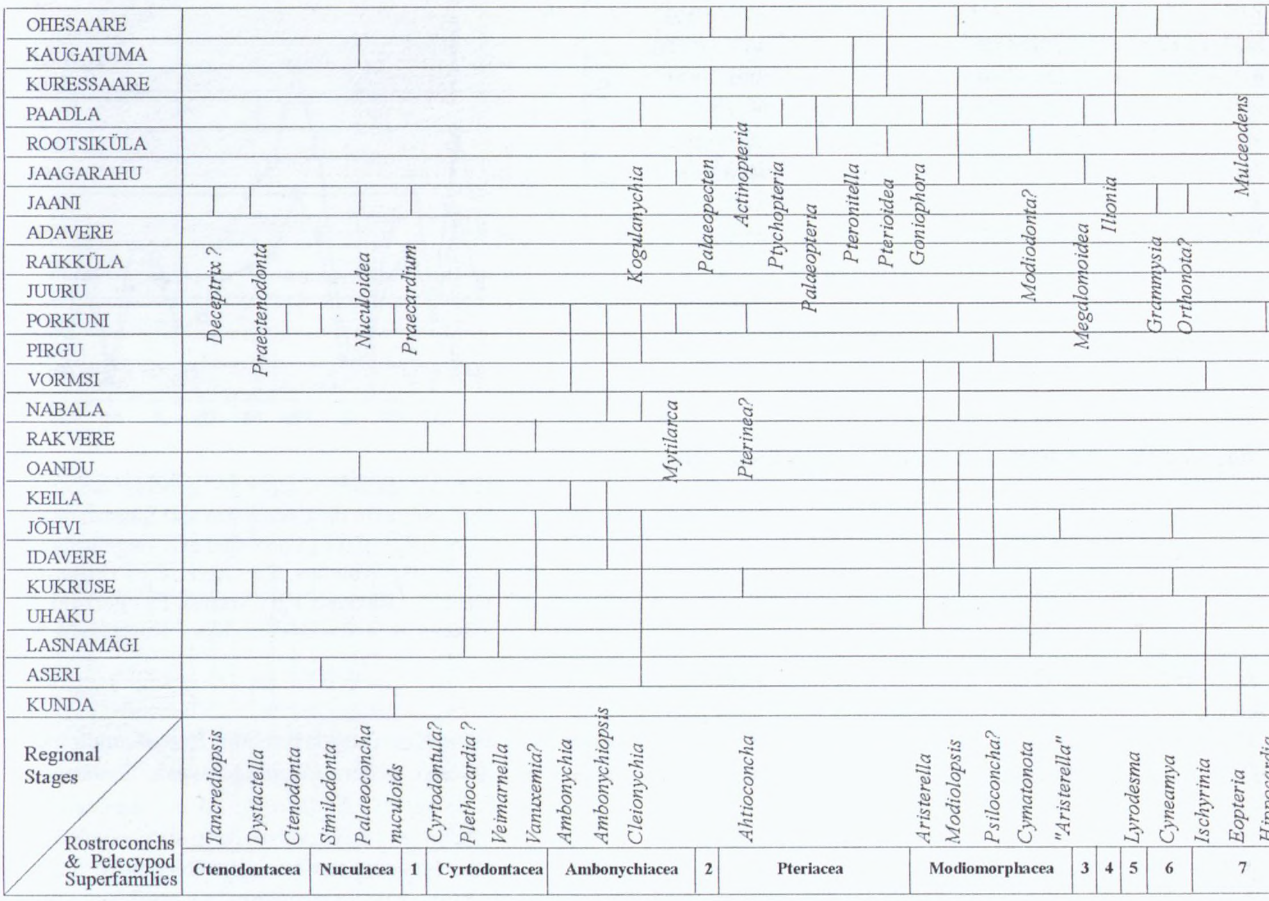
Table 36. Ranges of Ordovician and Silurian pelecypod and rostroconch genera in Estonia



Pelecypod superfamilies and rostroconchs: 1 - Praecardiacea; 2 - Pectinacea; 3 - Megalodontacea; 4 - Lucinacea; 5 - Trigoniacea; 6 - Pholadomyacea; 7 - Rostroconchia.

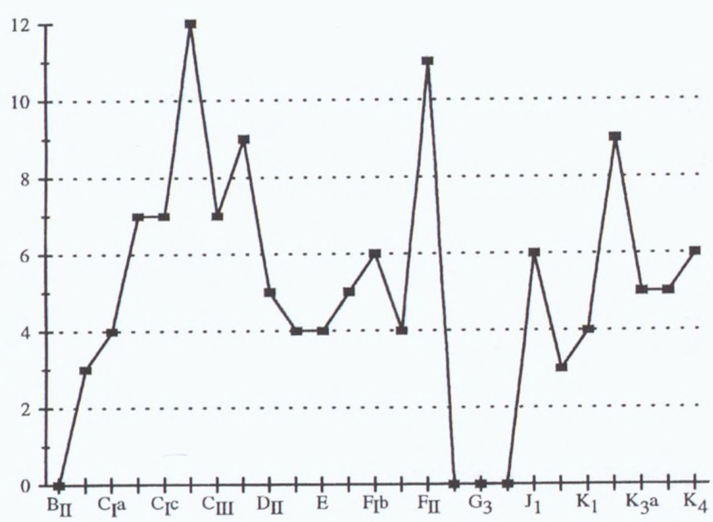


Fig. 156. Frequency dynamics of pelecypod genera in the Ordovician and Silurian of Estonia. Vertical axis - number of genera, horizontal axis - regional stages. Indices of stages: B_{II} - Volkhov, $C_{I}a$ - Aseri, $C_{I}c$ - Uhaku, C_{III} - Idavere, D_{II} - Keila, E - Rakvere, $F_{I}b$ - Vormsi, F_{II} - Porkuni, G_3 - Raikküla, J_1 - Jaani, K_1 - Rootsiküla, K_3a - Kuressaare, K_4 - Ohesaare

Photo 40. Ordovician and Silurian gastropods, bivalves and ostracodes:
1. Gastropod *Crenilunula limata*, Wenlock, Jaagarahu Stage, Jaagarahu quarry, x 1.5.
2. Gastropod *Megalomphala taenia* (Lindström), TÜG 2/27, Ludlow, Paadla Stage, nat. size.
3. Gastropod *Poleumita discors* (Sowerby), Wenlock, Jaagarahu Stage, Jaagarahu quarry, x 2.
4. Gastropod *Subulites amphora* (Eichwald), M. Ordovician, Jõhvi Stage, Sompa, x1.2.
5. Bivalve *Kogulanychia bekkeri* Isakar, La 1603, internal mould of right valve, Ludlow, Paadla Stage, Kogula quarry.
6. Bivalve coquina with Pterioidea, TÜG 40/59, Wenlock, Rootsiküla Stage, Anikaitse Cliff.
7. Ostracode *Retiprimites reticularis* Meidla, Os 3181, tecnomorphic right valve, M. Ordovician, Rakvere Stage, Puhmu borehole, depth 132.5 m, x 62.
8. Ostracode *Cystomatochilina clivosa* Meidla, Os 2995, left valve, lateral view, M. Ordovician, Rakvere Stage, Vinni borehole, depth 41.4m, x 41.
9. Ostracode *Plicibeirichia numerosa* Sarv, Os 5315, heteromorphic left valve, Ludlow, Kuressaare Stage, Kuressaare, x 30.
10. Ostracode *Nodibeirichia protuberans* (Boll), Os 5379, tecnomorphic left valve, Přidoli, Ohesaare Stage, Ohesaare Cliff, x 23.

occur also in the same stage. From the Rootsiküla Stage, *Palaeopteria*, *Modilopsis*, *Modiodonta* have been recorded. One type of pterioids inhabited the near-shore shallow-water environment with reduced salinity (see Эйнасто 1968), forming about a 0.8-m-thick deposit. A lot of bivalves: *Ilionia*, *Kogulanychia*, *Pteronitella*, *Palaeopecten*, *Modiolopsis*, *Ptychopteria*, *Megalomoidea*, *Goniophora* and *Palaeopteria* inhabited in the nearshore environments of the Paadla Age. The Kuressaare and Kaugatuma stages are characterized by *Pteronitella*, *Ilionia* and various *Pterioidea*. In the latter, a rostroconch *Mulceodens* and a small *Nuculoidea* have also been found. In the Ohesaare Stage, *Grammysia*, *Ilionia*, various *Pterioidea*, *Palaeopecten*, *Actinopteria* and *Modiolopsis*? are represented.

The above shows that in the Ordovician representatives of the superfamilies Cyrtodontacea, Modiomorphacea and Ambonychiacea dominated, while in the Silurian Pteriacea were most common.

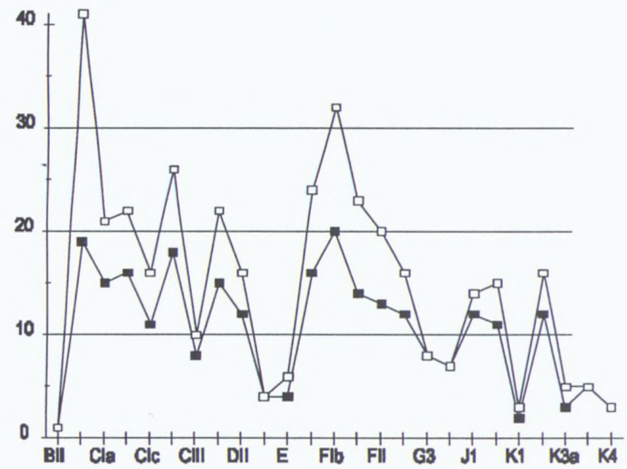


Fig. 157. Frequency dynamics of gastropod genera (black boxes) and species (white boxes) in the Ordovician and Silurian of Estonia. Vertical axis - number of taxa, horizontal axis - regional stages. Indices of the stages: B_{II} - Volkhov, C_1a - Aseri, C_1c - Uhaku, C_{III} - Idavere, D_{II} - Keila, E - Rakvere, F_1b - Vormsi, F_{II} - Porkuni, G_3 - Raikküla, J_1 - Jaani, K_1 - Rootsiküla, K_3a - Kuressaare, K_4 - Ohesaare.

Gastropods

The first data on the Ordovician and Silurian gastropods (Photo 40:1-4) from Estonia were published in the last century by Eichwald (1840b, 1842, 1860), Schmidt (1858, 1861, 1881) and Koken (1896, 1897, 1898). The latter also published a monograph on the Ordovician gastropods (Koken 1925). Teichert (1928) recorded a new Silurian gastropod species *Cyclonema hiiumaa* from Hiiumaa Island and Öpik (1930a) described some new gastropods from the Kukruse Stage. During the last years, the Ordovician and Silurian gastropods have been described by Isakar (1990, 1991, 1995) and Kiselev *et al.* (Киселев и др. 1990). The earliest gastropod in the Estonian sequence – *Aldanella kunda* (Öpik) – has been known from the Lontova Formation already, *i.e.* from the pre-trilobite Early Cambrian rocks.

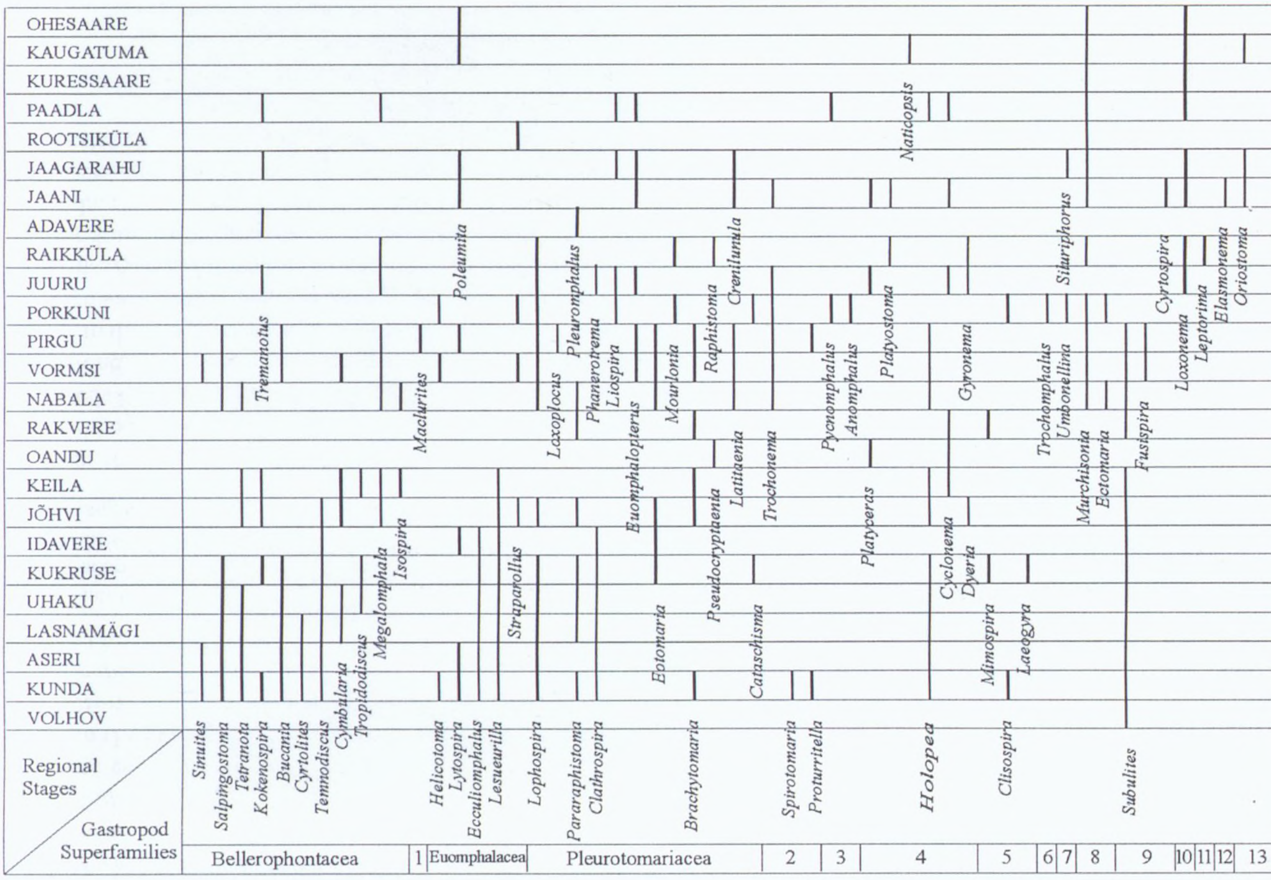
More than 200 species of gastropods from 63 genera have been identified from the Ordovician and Silurian of Estonia. Since all this material has been collected from northern and central Estonia, we have not been able to record the changes in the lateral distribution of gastropod associations.

The earliest Ordovician gastropod (*Subulites huenei* Koken) was presumably found from the Lower Ordovician glauconite limestone of the Volkhov Stage (Koken 1925). The diversity of the gastropod association in the next, Kunda Stage is much higher, comprising about 40 species which belong to 20 genera (Fig. 157). In the carbonate rocks of the Kunda Stage, there appeared bellerophontaceans – *Salpingostoma*, *Sinuites*, *Tetranota*, *Bucania*, *Cyrtolites*, *Temnodiscus*; pleurotomariaceans – *Pararaphistoma*, *Clathrospira*, *Lophospira*, *Brachytomaria*; euomphalins – *Lesueurilla*, *Lytospira*, *Ecculiomphalus*, *Helicotoma*; trochonemataceans – *Proturritella*, *Spirotomaria*, platyceratacean – *Holopea*, clisospiracean – *Clisospira* and subulitacean – *Subulites* (Table 37).

At the Early/Middle Ordovician transition between the Kunda and Aseri ages, the gastropod fauna underwent an essential change. The taxonomic diversity reduced remarkably; this is characteristic of the carbonate sequences throughout the whole Baltoscandian region (Jaanusson 1976). During the Lasnamägi, Uhaku and Kukruse ages, the new genera *Cymbularia*, *Tropidodiscus* and *Eotomaria* appeared, respectively. There were rather abrupt changes in species and generic levels at the Kukruse/Idavere and Keila/Oandu transitions. Many Idavere genera (*Salpingostoma*, *Temnodiscus*, *Cymbularia*, *Sinuites*, *Ecculiomphalus*, *Lesueurilla*, *Clathrospira*, *Subulites* and *Holopea*) have also been recorded from the erratic boulders in Germany (Neben & Krueger 1973). At the Keila /Oandu transition, a significant renovation of the Middle Ordovician fauna took place. Practically all gastropod species and numerous genera (*Kokenospira*, *Tropidodiscus*, *Temnodiscus*, *Lesueurilla*) disappeared. The diversity of the gastropods of the Oandu Age was relatively low and quite distinct from the assemblages of the preceding ages: platyceratid gastropods (*Cyclonema* and *Platyceras*) and rare pleurotomariacean *Pseudocryptaenia* appeared at that level. The Rakvere Stage consists mainly of calcilutites which contain a gastropod fauna quite different from that of the preceding stages. The genus *Murchisonia* and new species in *Subulites*, *Pararaphistoma*, *Mimospira* made their appearance at that time.

Most of the gastropod genera in the Nabala Stage (16) are long-ranging. *Tetranota conspiquaa* is a species occurring only in the Nabala Stage. During the Vormsi Age, the number of the gastropod genera reached 20 which is the maximum for the Late Ordovician. The argillaceous limestones of the Kõrgessaare Formation contain a rich and diverse assemblage of gastropods. The lower boundary of the stage is marked by a sharp change in lithofacies, but is comparatively weakly expressed in the distribution of gastropods. The species occurring only in the Vormsi Stage include *Straparollus vortex* and *Cymbularia aequalis*. *Sinuites*, *Bucania*, *Cymbularia*, *Helicotoma* and *Straparollus* reappear in the Vormsi Stage, and range into the Pirgu and Porkuni stages. In the Pirgu Age, the gastropod diversity decreased to 14 genera. The only Ordovician index gastropod - *Maclurites neritoides* - has been recorded from the Adila Formation (Решения... 1987). In the Porkuni Stage, a specific gastropod association is distributed. It consists of the euomphalaceans *Helicotoma*, *Straparollus*, pleurotomariaceans *Lophospira*, *Mourlonia*,

Table 37. Ranges of Ordovician and Silurian gastropod genera in Estonia



Gastropod superfamilies: 1 - Macluritacea; 2 - Trochonematacea; 3 - Anomphalacea; 4 - Platyceratacea; 5 - Clisospiracea; 6 - Pseudophoracea; 7 - Craspedostomacea; 8 - Murchisoniacea; 9 - Subulitacea; 10 - Loxonematacea; 11 - Genera Inquirenda; 12 - Microdomatacea; 13 - Oriostomatacea.

Cataschisma, trochonematacean *Trochonema*, anomphalaceans *Pycnomphalus*, *Anomphalus*, pseudophoraceans *Umbonellina*, *Trochomphalus*, murchisoniaceans *Ectomaria*, *Murchisonia*. Three genera – *Anomphalus*, *Trochomphalus* and *Umbonellina* – are restricted to this stage.

At the Ordovician/Silurian boundary, the gastropod fauna changed considerably – *Helicotoma*, *Cataschisma*, *Anomphalus*, *Clisospira*, *Trochomphalus*, *Umbonellina*, *Ectomaria* disappeared finally. The lowermost Silurian strata are characterized by a comparatively poor gastropod association, dominated by new species of pleurotomariaceans together with the first loxonemataceans (Isakar 1990). In the Wenlock gastropods are more abundant, with 12 genera being recorded from the Jaani and 8 genera from the Jaagarahu Stage. Practically all these genera (except *Murchisonia*) disappear at the Jaagarahu/Rootsiküla transition. From the upper part of the Rootsiküla Stage, only *Straparollus* (former "*Platyschisma helicites*") and *Murchisonia* have been recorded. In the Paadla Age, there were 9 genera of gastropods, but during the Kuressaare, Kaugatuma and Ohesaare ages, the gastropod taxa reduced in number – only 3-4 genera have been identified. The Silurian index gastropods – *Straparollus* ("*Platyschisma helicites*") sp. and *Murchisonia compressa,* have been recorded from the Rootsiküla Stage (Vesiku beds) and Paadla Stage, respectively (Решения… 1987).

The diversity and abundance of gastropods increased essentially during the transgressive phases of basin development coinciding with the Kunda, Kukruse, Jõhvi, Vormsi, Jaani and Paadla ages (Fig. 157). It shows that gastropods tended to prefer open-shelf conditions.

Ostracodes

The first data on the Lower Palaeozoic ostracodes from Estonia (Photo 40:7-10) were published in the middle of the last century by Eichwald (1854, 1860), Schrenk (1854) and Schmidt (1858). The latter published a monograph on Silurian leperditiids (Schmidt 1873). In the first half of this century, the ostracodes from the certain stratigraphical levels (Volkhov, Uhaku, Kukruse, Aruküla stages) were described (Bonnema 1909, Öpik 1935a, 1935b, 1937a). The regular study of Estonian ostracodes was started in the 1950s by Netskaya (Нецкая 1953, 1958, 1966, 1973), Stumbur (Стумбур 1956) and Sarv (Сарв 1959, 1968, 1977, 1980). Currently, these studies are carried on by Meidla (1996).

The earliest ostracodes, known from Estonia, are bradoriids *Bradoria? estonica* Melnikova and *Konicekion kundaensis* Melnikova from the Lower Cambrian Tiskre Formation. The mass appearance of ostracodes took place in the Early Ordovician Volkhov Age with incoming of the oldest ctenonotellids, eurychilinids and leperditellids (Sarv 1972). The first tvaerenellids, bolbinids and tetradellids appeared at the end of the Early Ordovician (Kunda Age). The Middle Ordovician (Aseri to Rakvere ages) was an acme for the representatives of Beyrichicopa - especially ctenonotellids, also tvaerenellids and tetradellids. The species diversity increased considerably from the beginning of this period reaching the peak during the Kukruse Age. Noticeable renovation of the fauna in the Idavere Age was followed by the period of stabilization lasting until the Keila Age. At the beginning of the Oandu Age most of the species were replaced and the number of metacopids began to increase, reaching 1/3 by the end of the Ordovician. During this period tetradellids, oepikellids, tvaerenellids and bollids were dominating among beyrichicopids, while the importance of ctenonotellids decreased considerably. The highest species diversity for the Ordovician Period was reached at early Pirgu time (ca 120 taxa according to Meidla 1996).

By the beginning of the Silurian, a rich and diverse Ordovician palaeocope fauna disappeared and only podocopes continued their evolution. In the Early Llandovery the first, strictly Silurian craspedobolbinids appeared, the oldest beyrichiids came in during the Middle Llandovery. The Wenlock Epoch was marked by the appearance of early cavellinids and Silurian primitiopsids. The species diversity of the Silurian ostracodes reached the maximum in Ludlow time, partly also at the beginning of Přidoli, gradually decreasing by the end of the period.

The Devonian ostracodes of Estonia are poorly known. Some specimens of leperditiids (probably *Leperditia geographica* Hecker) were found from the Middle Devonian Narva Stage and 7 species of different families were described by Öpik (1935a) from the Aruküla Stage. These species are the most ancient Devonian ostracodes recorded from the East-European Platform.

Altogether, some 400 Ordovician and 300 Silurian ostracode taxa have been identified in Estonia. In the stratigraphical distribution of the Ordovician ostracodes the Lower Ordovician (Volkhov and Kunda stages), lower Mid-Ordovician (Aseri - Kukruse stages), middle Mid-Ordovician (Idavere - Keila stages) and the topmost Mid-Ordovician - Upper Ordovician (Oandu - Porkuni stages) complexes can be distinguished. The vertical range of the Silurian ostracodes has been divided into the Llandovery (Juuru and Raikküla stages), topmost Llandovery - Wenlock (Adavere - Rootsiküla stages) and Ludlow - Přidoli complexes. All stages have their own characteristic species or species complexes, underlying the subdivision of sections and stratigraphical correlations (Сарв 1959, 1968, Кальо 1970в, Meidla 1996). On the basis of index species, ostracode zones have also been established (Table 7, see also Meidla & Sarv 1990).

During the Early Ordovician, the species assemblage of ostracodes was rather uniform throughout Estonia. In the Middle and Late Ordovician, the ostracode faunas were different in northern and southern Estonia. In the lateral distribution of the Silurian ostracodes some characteristic features have been mentioned (Нестор и Эйнасто 1977). Lagoonal sediments have yielded only representatives of large leperditiids characteristic of the dolomitic rocks of the Raikküla and Rootsiküla stages. Ostracodes of the shoal facies belt are less studied. Most of the Silurian ostracodes of Estonia have been found from the sediments of the open shelf facies belt. They differ noticeably from the ostracodes of the transitional facies belt, the differences being more distinct in the Ludlow and Přidoli.

The Ordovician and Silurian ostracodes have been successfully used for correlation purposes within the limits of the Baltic region. More long-distance correlations based on the common genera and species are possible, first of all, with Podolia, Norway, Great Britain and Canada (Абушик и Сарв 1983, Vannier *et al.* 1989). For a long time, ostracodes have been applied to dating of erratic boulders in northern Germany.

Ordovician trilobites

The Ordovician trilobites of Estonia (Photo 41:2-4) were described in a series of profound monographs by Schmidt (from 1881 to 1907) and Holm (1886). Important additions were later provided by Öpik (1937b) and Männil (Мянниль 1957).

During the Tremadoc and the Hunneberg Age, in Estonia the environment was unfavourable for trilobites. As a result of an environmental change during the following Billingen Age, a varied trilobite fauna appeared in northern Estonia (H. Pärnaste and particularly V. Jaanusson, personal communications). It consisted mainly of genera known from contemporaneous or even earlier strata of the Central Baltoscandian Confacies Belt in Sweden. In northern Estonia, the trilobite fauna remained diverse throughout the rest of the Ordovician (altogether some 100 genera or subgenera, see Table 38). The *"Chasmops" praecurrens, "C." maximus* and *"C." eichwaldi* groups are virtually new genera with several new species not yet described.

Faunal differences between the North Estonian and Central Baltoscandian confacies belts are clearly reflected also in the distribution of trilobites. The central belt extends to southern Estonia (Мянниль 1966, Мянниль и др. 1968, Jaanusson 1976), but as chances of finding macrofossils in drill cores are limited enough, there is at present no information available from the Lower Ordovician of southern Estonia. In the Middle and Upper Ordovician, 19 genera have been established in both northern and southern Estonia, whereas 5 genera of the central belt ("S" in Table 38) have never been recorded from northern Estonia. The differences between the North Estonian and central belts are still more conspicuos when the faunas from individual stages are compared.

Biogeographically, the benthic fauna of the Estonian Ordovician can be regarded as belonging to a separate

Photo 41. Cambrian to Silurian trilobites:

1. *Schmidtiellus mickwitzi* (Schmidt), the earliest trilobite from Estonia, L. Cambrian, Lükati Fm., Saviranna, x 1.6.
2. *Neoasaphus* sp., a complete specimen, M. Ordovician, Aseri Stage, nat. size.
3. *Cyrtometopus affinis* Angelin, Tr 2350, cranidium, L. Ordovician, Kunda Stage, Lõnna, x 3.
4. *Chasmops inge* Rõõmusoks, cranidium, M. Ordovician, Oandu Stage, Oandu, x 2.
5. *Calymene frontosa* Lindström, Tr 1963, enrolled specimen, Llandovery, Adavere Stage, Konovere River at Lätiküla, x 3.

6-7. *Encrinurus (E.) punctatus* (Wahlenberg), Wenlock, Jaani Stage, Paramaja Cliff: 6-Tr 2782, pygidium, x 2.5, 7- Tr 2775, partly enrolled specimen, x 2.6.

1

2

3

4

5

6

7

Table 38. Distribution of the Ordovician trilobite genera, in ordered of the appearance

	Ontika				Viru									Harju			
	B_Ia	B_Ib	B_{II}	B_{III}	C_Ia	C_Ib	C_Ic	C_{II}	C_{III}	D_I	D_{II}	D_{III}	E	F_Ia	F_Ib	F_Ic	F_{II}
ASAPHIDAE																	
Megalaspides (Megalaspides)		N															
Proasaphus		N															
Megistaspis (Megistaspis)			N														
Megistaspis (Paramegistaspis)		N	N														
Asaphus			N	N													
Onchometopus			N														
Megistaspidella			N	N													
Rhinoferus (Rhinoferus)			N														
Rhinoferus (Lawiaspis)				N													
Metaptychopyge			N														
Niobe (Niobella)			N														
Paraptychopyge			N														
Pseudoasaphus				N	N	X	N	N									
Ptychopyge				N													
Neoasaphus				N	X	N	X	X	X	X	X						
Homalopyge				N													
Heraspis				N													
Pseudobasilicus					N	N	N	N		N	N						
Xenasaphus						N	N										
Ogmasaphus							X	S									
Isotelus													N	N			
Brachyaspis																N	N
PLIOMERIDAE																	
Evropeites		N															
Pliomera				N													
NILEIDAE																	
Nileus		N	N			S	S										
BATYURIDAE																	
Agerina		N															
CHEIRURIDAE																	
Krattaspis		N															
"Ceraurinella"			N	N							N						
Cyrtometopus			N	N													
Acanthoparypha			N	N													
Pseudosphaerexochus					N		N				N			N	X	N	?
Nieszkowskia						N	N	N	N	N	N						
Paraceraurus					N	N	N	N									
Reraspis							N	N									
Sphaerocoryphe							N		N				N	N		N	
Hemisphaerocoryphe									N	N	N						
Cyrtometopella										N							
Ainoa											N						
Remipyga													N				
Sphaerexochus														N		N	
HARPIDAE																	
Scotoharpes					N												
Paraharpes																N	
ENCRINURIDAE																	
Cybele		N	N		N												
"Encrinuroides"		N															
Atractopyge							N	N	N	N	N	N	N		S	N	
Cybellela							N	N	N	N	N						
Erratencrinurus												N	N	N			
"Encrinurus"															N	N	N
PTERYGOMETOPIDAE																	
Pterygometopus			N	N													
Ingriops				N													
Estoniops					N		X	N	X	S	S	X					
Upplandiops							S										
Achatella (Vironiaspis)							N	N		N	N						

N - present in the North Estonian Confacies Belt; S - present in the Central Baltoscandian Confacies Belt (Livonian Tongue, S Estonia); X - present in both confacies belts. Indices of stages and substages (*): B_Ia - Hunneberg, B_Ib - Billingen, B_{II} - Volkhov, B_{III} - Kunda, C_Ia - Aseri, C_Ib - Lasnamägi, C_Ic - Uhaku, C_{II} - Kukruse, C_{III} - Idavere*, D_I - Jõhvi*, D_{II} - Keila, D_{III} - Oandu, E - Rakvere, F_Ia - Nabala, F_Ib - Vormsi, F_Ic - Pirgu, F_{II} - Porkuni.

Table 38 (Continued)

Achatella (Achatella)													N				
Chasmops praecurrens-group					N	N	N										
Chasmops							N	X	X	N	N						
Scopelochasmops									X								
Chasmops maximus-group										N	N						
Bolbochasmops										N	N						
Rollmops										N							
Oculichasmops										X	N						
Toxochasmops											S	X	N		N		
Calyptaulax														S			
Chasmops eichwaldi-group														N	N	N	
RAPHIOPHORIDAE																	
Globampyx			N														
Ampyx		N		N													
Lonchodomas				N			X	X	S	S	S	S					
ODONTOPLEURIDAE																	
Boedaspis				N													
Apianurus								N			N					N	
Acidaspis													N				N
Diacanthaspis										N		N		N			
ILLAENIDAE																	
Ottenbyaspis		N															
Illaenus (Illaenus)				N	N	X	X	N	N	N	N						
Stenopareia								N	N		N	N	N		N	N	
Panderia							S	X	S	S	S	S					
Theamataspis								N									
Bumastoides												N	N				
Illaenus (Parillaenus)														N	N	N	N
LICHIDAE																	
Metopolichas				N		N	N	N									
Hoplolichas					?	N	N	N			N						
Hoplolichoides							N	N									

Otarozoum								N	N	N		N	N				
Conolichas									N		N	N		N	N	N	
Autoloxolichas										N				N	N		
Platylichas (Platylichas)										N				N	S		N
Leiolichas											N						
Hemiarges												N	N	N			N
Amphilichas													N	N	N	N	
Platylichas (Rontrippia)?																	N
"Trochurus"																	N
REMOPLEURIDIDAE																	
Remopleurides					N	S	X	X	S	S	S	S		N			
SCUTELLUIDAE																	
Bronteopsis							?	?									
Eobronteus																N	
PROETIDAE																	
Stenoblepharum								N									
Decoroproetus													N				
Ascetopeltis															N		N
AULACOPLEURIDAE																	
Panarchaeogonus								N									
"Harpidella"								N									
DIMEROPYGIDAE																	
Dimeropyge								N									
CALYMENIDAE																	
Pharostoma								N					N	N	N		
"Flexicalymene"															X		N
PHILIPSINELLIDAE																	
Phillipsinella									S						S		
TRINUCLEIDAE																	
Tretaspis														X	S	S	
HOMALONOTIDAE																	
Brongniartella																	S
DALMANITIDAE																	
Dalmanitina (Mucronaspis)																	S

Baltoscandian Province (for a review see Jaanusson 1979). With regard to trilobites, this province is characterized by a great taxonomic diversity of large asaphid trilobites, especially *Megistaspis, Ptychopyge* and the related genera in the Ontika Series, and *Neoasaphus* and the related forms in the Viru Series ("Asaphid Fauna", Whittington 1966, "Asaphid Province Fauna", Whittington & Huges 1972). Further characteristics include the taxonomic diversity of the Pterygometopinae in the upper Ontika and Viru rocks (Jaanusson & Ramsköld 1993), Chasmopinae in the Viru and Harju series and various lichids.

The temporary immigration of benthic genera, whose affinities are mainly with the North American midcontinent region, is in the trilobite fauna of northern Estonia reflected by the appearence of *Bumastoides* and, possibly, *Achatella (Achatella).*

There is an apparent general tendency towards an increasing difference in the trilobite faunas of the North Estonian and Central confacies belts from the mid-Viru upwards with a culmination in the Harju Series. The normal Upper Ordovician fauna of the central belt, mostly dominated by trinucleids, has a few genera in common with the North Estonian Belt. The uppermost Ordovician Hirnantia fauna with its characteristic trilobites *Dalmanitina (Mucronaspis)* and *Brongniartella*, is known also from southern Estonia.

Silurian trilobites

The Silurian trilobites of Estonia were described by Nieszkowski (1857,1859), Holm (1886) and Schmidt (1881-1907) in the last century already. In this century particular families have been studied, such as encrinurids (Rosenstein 1941, Мяэннил 1958г, Мяэннил Р. П. 1977а, Edgecombe & Ramsköld 1996), calymenids (Мяэннил Р. П. 1977б, 1983), phacopids, *etc.* As a result, about 35 trilobite genera (Table 39) and 100 species have been recognized up to now. Most common representatives of the Silurian trilobite fauna in Estonia are calymenids, encrinurids and proetids which form more than half of the described species. Encrinurids prevailed in the Llandovery, calymenids and proetids in the Wenlock, and particularly in the Late Silurian.

During the Late Ordovician, the typical Ordovician families gradually disappeared and only a few genera crossed the Ordovician/Silurian boundary. The basal Silurian is recognized by the appearance of the family Phacopidae and several new genera, including *Calymene* and the typical Llandovery genera *Acernaspis*, *Opsypharus, Elsarella, etc.* The lowermost Llandovery contains an impoverished fauna. A relatively diverse and rich assemblage is known from the Tamsalu Formation of the Juuru Stage containing *Acernaspis, Calymene* and encrinurine trilobites of the *"Encrinurus" variolaris* plexus (Strusz 1980, Edgecombe & Ramsköld 1996) in the shoal facies, and locally abundant *Opsypharus* and *Stenopareia* in bioherms. From the overlying shallow-water deposits of the Raikküla Stage, only sparse trilobites have been found. In southern Estonia, the deeper-water facies of the Juuru and Raikküla stages are characterized by rare trilobites, predominantly by different species of *Acernaspis* and encrinurines. The maximum rise in species diversity has been recorded from the uppermost Aeronian - Rumba Formation of the Adavere Stage, characterized by the appearance of a number of new species and some short-ranging genera (*Radiurus, Distyrax,* a.o.) (Männil 1992). In general, the fauna of the Rhuddanian and Aeronian ages was relatively homogenous in terms of the dominating genera. The following Telychian transgression caused a notable evolutionary change. In Estonia, it coincides with the boundary between the Rumba and Velise formations and is characterized by the disappearance of the genus *Stenopareia* and encrinurine genera of *variolaris* plexus which are replaced by *Encrinurus* and *Wallacia.* At this boundary an almost complete turnover of species assemblage took also place.

Across the Llandovery/Wenlock boundary, the trilobite fauna changed remarkably. The genus *Acernaspis* disappeared finally, while *Proetus* (*s.str.*) and *Dalmanites* came in, the latter being abundant in southern Estonia. Calymenids became prevalent occurring in all trilobite-bearing facies. The role of proetids increased. During the Jaagarahu Age, the genera *Cyphoproetus, Pseudotupolichas* and some new species appeared. Higher up in the sequence, the diversity decreases considerably due to long regression, and no trilobites have been found from the uppermost Jaagarahu and Rootsiküla stages.

The Upper Silurian trilobite faunas of Estonia are of shallow-water origin and unilateral, greatly dominated by *Calymene* and *Pulcherproetus.* No trilobites are known from the lower part of the Paadla Stage. The species, recorded from the Uduvere beds, belong to *Struszia, Pulcherproetus, Calymene* and rare lichids. Trilobites of the Kuressaare Stage are of very low diversity. Only *Calymene* and *Pulcherproetus* are known from the Kudjape beds.

In the Přidoli, the diversity and frequency of trilobites increased. The open shelf sediments of the Kaugatuma Stage yield numerous specimens of *Pulcherproetus* and three species of *Calymene*, accompanied by rare representatives of *Eophacops* and *Acaste.* The same genera occur also in the Ohesaare Stage, but the fauna differs on the species level.

In general, the Silurian trilobites had a wide environmental range, but because of their strong dependence on lithofacies, their generic and specific compositions changed considerably along the palaeoslope of the Baltic Silurian Basin (Мяэннил 1986). Accordingly, several trilobite communities, related to the different facies belts, have been established (Männil 1982, Мяэннил 1982). In the Llandovery, the shallow-water shoal facies belt was dominated by a styginid-illaenid fauna which was replaced by a calymenid-encrinurid fauna in the open shelf environment and an encrinurid-phacopid fauna in the transitional and depression facies belts. During the Wenlock, calymenids and lichids (in bioherms) were prevailing in the shoal belt, being replaced by the encrinurid-proetid-calymenid fauna in the open shelf environment and by a rich dalmanitid-calymenid fauna in the transitional and depression facies belts. The Late Silurian trilobite faunas were similar to each other by dominating genera because of their more or less uniform shallow-water origin all over the area in consideration.

Echinoderms

In spite of their great importance in the shallow water faunal associations, echinoderms in Estonia have been studied insufficiently and unevenly. On some levels, the dissociated skeleton elements of different echinoderms, especially of pelmatozoans, form an essential part of the bioclastic material (skeletal sand) in the composition of carbonate rocks (5 -

Table 39. Ranges of Silurian trilobite genera in Estonian sequence

Trilobite genera	G_{1-2}	G_3	H	J_1	J_2	K_1	K_2	K_3a	K_3b	K_4
Opsypharus (St)	+									
Thebanaspis? (Pr)	+									
Astroproetus (Pr)	+									
Lichas (L)	+									
Hadromeros (Ch)	+	?								
Deiphon (Ch)	+	+								
Hedstroemia? (Pr)	+		?							
Elsarella (En)	+		+							
Stenopareia (Il)	+	+	+							
Acernaspis (Ph)	+	+	+							
Exallaspis (Od)	+	?		+	+					
Anacaenaspis (Od)	?		?	+						
Calymene (Cl)	+	+	+	+	+		+	+	+	+
Diacalymene (Cl)	?	+								
Bumastus (St)	?		+	+						
Otarion (Al)		+	+							
Proetus (Pr)		?		+	+					
Meroperix? (St)			+							
Radiurus (Ch)			+							
Distyrax (En)			+							
Wallacia (En)			+							
Kettneraspis? (Od)				+						
Encrinurus (En)			+	+	+					
Maurotarion (A)				+						
Warburgella (Pr)				+	+					
Dalmanites (Dl)				+	+					
Richterarges (L)				+			+			
Cyphoproetus (Pr)					+					
Pseudotupolichas (L)					+					
Lichidae gen. indet. (L)					+					
Struszia (En)							+			
Pulcherproetus (Pr)							+	+	+	+
Eophacops (Ph)									+	+
Acaste (Ac)									+	+

Abbreviations of families in brackets: St - Stryginidae, Pr - Proetidae, L - Lichidae, Ch - Cheiruridae, Il - Illaenidae, En - Encrinuridae, Ph - Phacopidae, Od - Odontopleuridae, Cl - Calymenidae, A - Aulacopleuridae, Dl - Dalmanitidae, Ac - Acastidae.
Indices of stages: G_{1-2} - Juuru, G_3 - Raikküla, H - Adavere, J_1 - Jaani, J_2 - Jaagarahu, K_1 - Rootsiküla, K_2 - Paadla, K_3a - Kuressaare, K_3b - Kaugatuma, K_4 - Ohesaare.

35%, in some cases up to about 90%, see Пылма 1982). Frequent occurrence of some cystoids is characteristic of distinct beds of the Middle Ordovician, the *"Echinosphaerites* Limestone" in the lower part and the "*Hemicosmites* Limestone" in the upper part of the Viru Series (see Chapter IV).

Eichwald (1860) was the first to describe Estonian echinoderms. Afterwards they were studied by Schmidt, Jaekel, Hecker, Ralf Männil and others (references see Геккер 1964). According to Ralf Männil (unpublished data), in Estonia some 50 genera and 150 species of Ordovician echinoderms have been established by more or less complete skeletons. They were dominated by cystoids and crinoids (more than 100 species). To the last group belong also most of the taxa established by columnals.

The diversest and richest associations of echinoderms (cystoids, crinoids, eocrinoids, edroiasteroids, carpoids, paracrinoids, ophiocystoids, a. o.) occurred in northern Estonia. In southern Estonia, the environmental conditions were presumably quite unsuitable for the distribution of echinoderms during the most of the Ordovician.

In Estonia the first echinoderms (crinoids) appeared in the Billingen Stage (Table 40), in the glauconitic sandstones of the Mäeküla Member. Upwards, in the Volkhov and Kunda stages, several cystoids (*Cheirocrinus, Echinoencrinites*) and eocrinoids *(Bockia, Rhipidocystis)* appear. The upper part of the Kunda Stage is characterized by the first appearance of some crinoids known by columnals (*Babanicrinus, Schizocrinus, Baltocrinus)* which are widespread in the Middle Ordovician rocks (Table 40).

In the lower and middle parts of the Viru Series, the echinoderms are represented by the largest number of taxa in the Ordovician. At the base or in the lower part of the series, there appear new cystoids (*Echinosphaerites, Scoliocystis*), crinoids (*Hoplocrinus)* and some eocrinoids (*Rhipidocystis, Bockia*) become frequent. In offshore facies in central and southern Estonia, some taxa (*e.g. Echinosphaerites*) have a wider stratigraphical distribution than in onshore facies in northern Estonia. Rich and diverse association of echinoderms characterizes the Oandu Stage in northern Estonia (Table 40). At that, the carbonate mounds of the Vasalemma Formation

Table 40. Distribution of selected Ordovician echinoderms in northern Estonia

Series	Oeland				Viru								Harju			
Genus \ Stage	Hunneberg	Billingen	Volkhov	Kunda	Aseri	Lasnamägi	Uhaku	Kukruse	Haljala	Keila	Oandu	Rakvere	Nabala	Vormsi	Pirgu	Porkuni
EOCRINOIDEA																
Rhipidocystis			x	x				x	x							
Bockia			?	?	x		x	x	x							
Neorhipidocystis								x	x							
PARACRINOIDEA																
Achradocystis										x	x					
EDRIOASTEROIDEA																
Cyathocystis							x				x					
CYSTOIDEA																
Cheirocrinus			x	x		x	x	x								
Echinoencrinites			x	x												
Hemicosmites				x			x		x		x			x	x	x
Echinosphaerites					x		x	x	x							
Glyptosphaerites					x		x									
Heliocrinites							x							x	x	
Protocrinites								x	x							
Revalocystis											x	x				
Esthonocystis											x					
Asperellacystis											x					
Oanducystis											x					
Tricosmites																x
CRINOIDEA																
Asterocrinus		x	x													
Monilecrinus		x	x	x												
Tetragonocrinus			x													
Irucrinus				x												
Babanicrinus				x	x	x	x	x	x	x						
Baltocrinus				x	x	x	x	x	x	x						
Schizocrinus?				x	x	x	x	x	x	x	x					
Revalocrinus					x	x										
Hoplocrinus					x		x	x	x	x	x					
Baerocrinus							x									
Ristnacrinus								x	x	x	x	x	x	x	x	
Rhaphorocrinus								x	x	x						
Carabocrinus											x					
Metabolocrinus											x					
Xenocrinus											x	x	x	x	x	
Rugulosocrinus											x					
Fossulacrinus											x	x	x			
Unufossulacrinus												x	x	x		
Pentagonocrinus														x	x	
Niibicrinus															x	
Porkunicrinus																x
ECHINOIDEA?																
Bothriocidaris								x	x		x			x	x	x

comprise edrioasteroid *Cyathocystis* which seems to form a frame in some mounds. The inter mound bioclastic limestones consist mainly of pelmatozoan columnals and skeleton elements of cystoids. Among the latter ones, the most well-known is *Hemicosmites*.

In the upper part of the Viru Series (Rakvere Stage) and in the Harju Series, the distribution of echinoderms is uneven. They are rare in the micritic (aphanitic) limestones, but their columnals are abundant (Table 40) in the talus facies of reefs or in the argillaceous limestones.

The data on the distribution of echinoderms in the Silurian sequence of Estonia are very scanty. The pelmatozoan columnals occur (in places abundantly) in the argillaceous limestones, but there are few data on their taxonomic composition. In the Adavere Stage, *Myelodactylus* and *Glyptocrinus* can be mentioned as taxa distributed widely in the Llandovery rocks in the East-European Platform and some other regions. Inadunate crinoid *Pisocrinus* is characteristic to the Jaani and Jaagarahu stages, to the deep-water deposits lying between graptolite and shelly facies (Рожков и др. 1989). The Paadla and Kuressaare stages are characterized by the occurence of *Crotalocrinites* which presumably appears first in the uppermost Llandovery, in the Adavere Stage. In the lowermost Přidoli, in the Kaugatuma Stage, *Crotalocrinus* is a rock-forming fossil in the cyclically recurrent deposits of crinoidal limestones.

The data available from the uppermost Silurian suggests that *Leptocrinites, Eucalyptocrinites* and *Anthinocrinus* are of biostratigraphic importance for the Kaugatuma Stage and *Hexacrinites* and *Cicerocrinus* for the Ohessaare Stage.

Graptolites

In the Early Palaeozoic biota of Estonia, graptolites played a leading role only in a few time intervals and in limited areas. They occurred in the Tremadoc of northern Estonia and in the Aeronian and Early Wenlock of southwestern Estonia. Except for the Tremadoc *Dictyonema* Shale, these graptolite-bearing rocks (marlstones, argillites) give evidence of the distribution of deeper, outer shelf margin and basin facies in Estonia.

In carbonate rocks of the post-Tremadoc Ordovician and Silurian, graptolites occur sporadically in Estonia. Beside rare finds of graptolite remains on slab surfaces, many new occurrences have been established by processing rock samples to obtain organic-walled microfossils (Мяннилъ 1976, Kaljo & Männil 1990).

Taxonomically, these graptolite associations from argillites and carbonate rocks are completely different: nearly all Tremadoc graptolites belong to the order Dendroidea. From the Varangu Stage *Didymograptus* (?) *primigenius* has been identified (Кальо и Кивимяги 1976); in most cases it is probably a dendroid *Kiaerograptus* (bithecas were not seen because of poor preservation). In the Tremadoc sequence of Estonia, the following succession of biozones has been established (in ascending order): the *Rhabdinopora flabelliformis* (with *R. f. socialis* and *R. f. norvegica, etc.*) and *R. anglica - R. multithecata* zones in the lower and the *Adelograptus - Kiaerograptus* Zone in the upper Tremadoc. The lowermost Tremadoc zone (*R. parabola*) has not been established in Estonia.

Higher in the Ordovician and Lower Silurian benthic dendroids (*Dictyonema, Thallograptus, Inocaulis, Estonicaulis, etc.*, Обут 1953, Обут и Рыук 1958) occur sporadically.

A specific assemblage of dendroid graptolites (*Rhadinograptus jurgensonae, Leveillites, Mastigograptus, Crinocaulis, etc.*, Обут 1960) was described from the Raikküla Stage in central Estonia (Kursi, Laeva, Survaküla, Kanaküla, Nurme, *etc.* core sections). However, a graptolite origin of some of these fossils is doubtful.

In the post-Tremadoc Ordovician and Lower Silurian, the sporadic occurrences and, naturally, the graptolite-bearing facies were dominated by planktic representatives of the order Graptoloidea. The last order was a rapidly evolving group providing good key species for correlations and biozonation.

From the Ordovician carbonate rocks of Estonia, Öpik (1927) and Jaanusson (1960b) described several species of graptoloids. As a summary, Männil (Kaljo & Männil, 1990) listed the following most significant graptoloids: *Didymograptus pakrianus* from the Kunda Stage, *D. acutus* from the Aseri Stage, *Climacograptus distichus* from the Lasnamägi Stage, *Gymnograptus linnarssonii* from the Uhaku Stage, *Amplexograptus bekkeri* and *Nemagraptus gracilis* from the Kukruse Stage and *A.* cf. *fallax* (= *A. baltoscandicus,* Jaanusson 1995) from the Haljala and Keila stages, *Pseudoclimacograptus* cf. *scharenbergi* from the Haljala Stage, *Climacograptus spiniferus* from the Rakvere Stage and from the lower part of the Nabala Stage, *Archiretiolites regimontanus* and *Rectograptus gracilis* from the Saunja Formation (upper part of the Nabala Stage). The last species seems to occur until the lower part of the Pirgu Stage. From the uppermost Pirgu Stage (Kabala Formation), *Climacograptus supernus* has been recorded.

The same sporadic style of occurrences continued in the Silurian carbonate rocks: *e.g. Pseudoclimacograptus hughesi* in the Juuru Stage, *Paraclimacograptus estonus* and *Coronograptus* aff. *gregarius* in the Raikküla Stage, rare specimens of *Climacograptus* in the lowermost part of the Adavere Stage. However, a more complete picture on the graptolite diversity changes can be obtained from the subsurface sections south of Pärnu and in the Gulf of Riga. Based mostly on the data from the Ikla (Кальо и Вингисаар 1969), Holdre (Ульст 1970), Ohesaare (Кальо 1970a) and Ruhnu (unpubl.) cores, the following succession of graptolite zones has been established (see Table 8): *Dimorphograptus confertus, Coronograptus cyphus, C. gregarius* and *Demirastrites triangulatus*, a gap in the succession in the territory of Estonia, *Spirograptus turriculatus, Monograptus crispus, Monoclimacis griestoniensis, Oktavites spiralis* (partly), *Cyrtograptus murchisoni, Monograptus riccartonensis.* Higher in the Estonian sections, the graptolites become sporadic again; only *Pristiograptus sardous, Monograptus flexilis, M.* ex. gr. *flemingi* and *Gothograptus nassa* have been recorded from the upper part of the Wenlock sequence at Ohesaare (Кальо 1970a).

All these graptolite data have proved highly valuable for stratigraphical correlations.

Conodonts

The history of study on conodonts dates back 140 years (Pander 1856). The systematic study of the Ordovician and Silurian conodonts of Estonia (Photo 42:1-14) started in 1966.

In a number of papers and two doctoral theses (Вийра 1974, Männik 1992b) the ranges of the Lower Palaeozoic conodonts were established and the faunas of all main stratigraphic units were described. Conodonts from the Upper Silurian (Viira 1982, Вийра 1983), Upper Llandovery (Männik 1992a, Jeppsson & Männik 1993) and the Cambrian-Ordovician boundary beds (Kaljo *et al.* 1986, 1988, Viira *et al.* 1987) have been studied in more detail. Multielement taxonomy in conodont studies was accepted in the seventies. New Silurian conodont apparatuses were described and illustrated by Viira (Вийра 1982, Viira 1994) and Männik (1992a, b).

The Estonian Lower Palaeozoic permits to trace the evolution of the conodont faunas from the beginning of their appearance in the Late Cambrian through their highest diversity in the Early Ordovician up to the slow decline in the Silurian. All the seven known orders of Conodonta and representatives of 27 families (from 43) have been found in Estonia (Table 41).

The oldest conodonts are proconodontids belonging to the family Cordylodontidae. They are most numerous in the Cambrian-Ordovician boundary beds (Kaljo *et al.* 1986, 1988). The Early Tremadoc zonation is based on the evolutionary lineage of *Cordylodus*. All zones of this age are defined in the Kallavere and Türisalu formations. The appearance of the first striated coniform protopanderodontids (*Paltodus, Paroistodus, Drepanoistodus, Drepanodus, Scolopodus, etc.*) marks the beginning of a very diverse conodont fauna in the Varangu and Hunneberg stages. The occurrence of the first prioniodontids with combshaped or platform P elements (*Oistodus, Acodus, Oelandodus, Prioniodus, Oepikodus, Periodon, Baltoniodus, etc.*) followed in the Billingen and Volkhov stages (Вийра 1966, 1970, Mägi et al. 1989). Species and subspecies of *Baltoniodus* are the most frequent components of the fauna in the interval of the Kunda to Kukruse stages. The evolutionary lineage of platformed prioniodontid genus *Eoplacognathus* is used as a basis for the zones and subzones in the same stratigraphic interval. During the Aseri Age, the laterally furrowed coniform panderodontid genus *Panderodus* appeared passing the rest of Ordovician and Silurian. The genus *Amorphognathus*, the best known member of the family Balognathidae, is widespread, especially in the Kukruse Stage (*A. tvaerensis*), and has a long range upwards. The *A. superbus* Zone corresponds to the interval from the Keila Stage up to the Nabala Stage, and the *A. ordovicicus* Zone to the Vormsi, Pirgu and Porkuni stages. Some representatives of the prioniodontids are also known from the Upper Ordovician, *e.g.* the earliest *Icriodella* species, *Phragmodus* and *Plectodina* have been found in the Vasalemma Formation. A belodellid *Hamarodus* with rastrate elements is characteristic of the Nabala Stage, particularly of the Mõntu and Saunja formations. The occurrence of the youngest protopanderodontids – *Strachanognathus* – is limited to the Pirgu Stage. By the end of the Ordovician, the diversity of conodont faunas considerably decreased - only 10 families of 21 survived during the end-Ordovician extinction.

The Ordovician and Silurian boundary is marked with the first appearance of ozarkodinids and prioniodinids (exceptional is the genus *Erraticodon* occuring in the Middle Ordovician already). Conodonts are rare in the lowermost Silurian. Only a single prioniodontid *(Distomodus)*, a pectiniform ozarkodinid (*Ozarkodina*), a prioniodinid (*Oulodus*) and some coniforms (*Walliserodus, Dapsilodus, Decoriconus*) characterize the Juuru Stage (Männik 1992a,b, Männik & Viira 1993). The diversity of the conodont faunas considerably increased during the Raikküla Age when the earliest representatives of the platform ozarkodinids (*Kockelella*) and pterospathodontids (*Pranognathus*) appeared. An extremely rich fauna, dominated by the pterospathodontids (*Pterospathodus, Apsidognathus, Astropentagnathus*), ozarkodinids (*Aulacognathus, Ozarkodina*) and panderodontids *(Panderodus, etc.)* characterizes the Velise Formation and the lowermost part of the Jaani Formation (Männik 1992a,b, 1994, Männik & Viira 1993, Jeppsson & Männik 1993). In the Jaagarahu Stage and upwards, the conodont fauna is represented mainly by ozarkodinids *Ozarkodina, Kockelella, Ctenognathodus*; the prioniodinid *Oulodus. Ctenognathodus murchisoni* is characteristic of the Rootsiküla Formation, particularly of the Vesiku beds on Saaremaa (Вийра 1982). Mainly three species – *Ozarkodina confluens, O. excavata, and Oulodus siluricus* – are abundant in the fauna of the Paadla Stage. *Ozarkodina roopaensis* is characteristic of the Sauvere and Himmiste beds on Saaremaa. Biostratigraphically important are *Ozarkodina snajdri* and *O. crispa* in the Paadla and Kuressaare stages. The fauna of the three succeeding units –

Photo 42. Ordovician and Silurian conodonts:
1. *Cordylodus andresi* Viira et Sergeyeva, Cn 1082, U. Cambrian, Vihula section, sample 5, x 100.
2. *Paltodus deltifer pristinus* (Viira), M element, Cn 1527, L. Ordovician, Pakerort Stage, Varangu borehole, depth 18.60 m, x 100.
3. *Baltoniodus variabilis* (Bergström), Pb element, Cn 1528, M. Ordovician, Kukruse Stage, Kohtla section, bed F, x 65.
4. *Eoplacognathus reclinatus* (Fåhraeus), Pb element, Cn 1529, M. Ordovician, Lasnamägi Stage, Suhkrumägi, x 40.
5. *Prioniodus elegans* Pander, Pb element, Cn 1530, L. Ordovician, Billingen Stage, Voka, x 100.
6. *Amorphognathus tvaerensis* Bergström, Pb element, Cn 1531, M. Ordovician, Kukruse Stage, Kaagvere borehole, depth 296.8 m, x 100.
7. *Oulodus panuarensis* Bischoff, M element, Cn 5518, Llandovery, Raikküla Stage, Kalana quarry, sample Kl-4, x 100.
8. *Oistodus lanceolatus* Pander, M element, Cn 1532, L. Ordovician, Billingen Stage, Mäekalda section, sample 7a, x 75.
9. *Ozarkodina roopaensis* Viira, Pa element, Cn 1398, Ludlow, Paadla Stage, Karala outcrop, x 50.
10. *Ozarkodina crispa* (Walliser) morph, Pa element, Cn 1475, Ludlow, Paadla Stage, Kaugatuma borehole, depth 58.5 m, x 75.
11. *Ozarkodina p. polinclinata* (Nicoll *et* Rexroad), Pa element, Cn 6509, Llandovery, Adavere Stage, Viki borehole, interval 123.3-123.5 m, x 100.
12. *Ozarkodina b. bohemica* (Walliser) ß morph, Pa element, Cn 1533, Wenlock, Rootsiküla Stage, Vesiku borehole, depth 8.65 m, x 40.
13. *Ozarkodina remscheidensis eosteinhornensis* (Walliser) s.l., Pa element, Cn 1506, Ludlow, Kuressaare Stage, Vaivere outcrop, x 40.
14. *Ozarkodina r. remscheidensis* (Ziegler), Pa element, Cn 1445, Přidoli, Ohesaare Stage, Ohesaare Cliff, x 40.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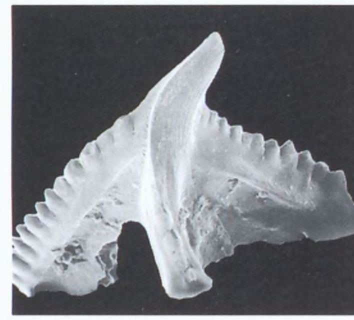
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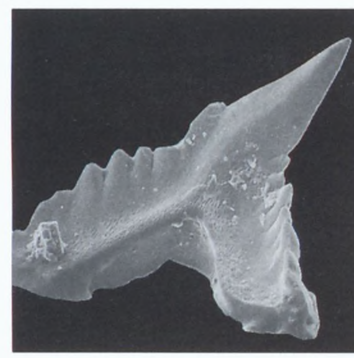
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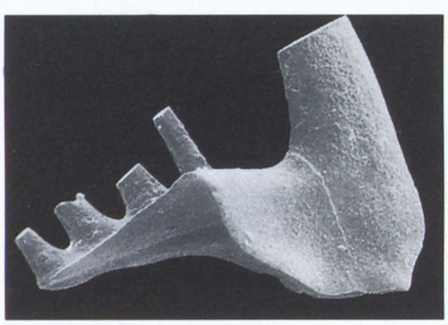
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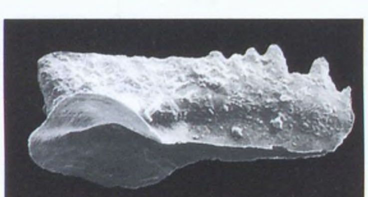
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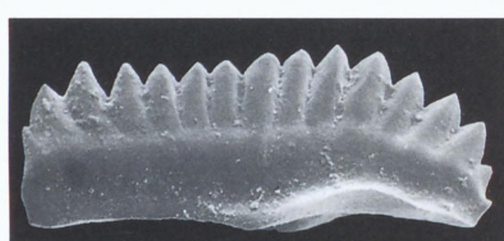
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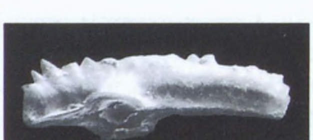
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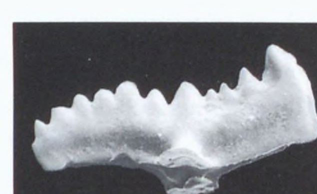
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Table 41. The number of the conodont genera in the Ordovician and Silurian

	Order	Proconodontida		Belodellida		Protopandero-dontida							Panderodontida	Prioniodontida											Prioniodinida		Ozarkodinida		Unknown
	Family	Proconodontidae	Cordylodontidae	Belodellidae	Ansellidae	Oneotodontidae	Acanthodontidae	Drepanoistodontidae	Protopanderodontidae	Cornuodontidae	Dapsilodontidae	Strachanognathidae	Panderodontidae	Balognathidae	Oistodontidae	Prioniodontidae	Periodontidae	Polyplacognathidae	Plectodinidae	Cyrtoniodontidae	Icriodellidae	Icriodontidae	Distomodontidae	Pterospathodontidae	Prioniodinidae	Chirognathidae	Spathognathodontidae	Kockelellidae	
	STAGES *	3(2)		2(2)		8(7)							1	14(11)											5(2)		10(2)		
SILURIAN	Ohesaare												1												1		1	?	
	Kaugatuma												1												1		1		
	Kuressaare												1												1		1	?	
	Paadla												1										1		1		1	1	
	Rootsiküla												1												1		1	2	
	Jaagarahu			1							1		1												1		1	1	1
	Jaani			1				1			1		1										2	2	?		1	1	2
	Adavere			1				1					1								1		2	3	?		1	2	2
	Raikküla			2				1			1		1								2	1	1		?		1	2	1
	Juuru			1				1			1		1								1		1		1		1		1
ORDOVICIAN	Porkuni			1				2					1	1													?		1
	Pirgu			2	1			2			1	1	2	2					1		1								1
	Vormsi			2	1			2			1		2	2					1		1								1
	Nabala			1	1			2			1		2	2							1								1
	Rakvere							1	1				3	1															1
	Oandu							1					1	1					2	1	1								1
	Keila			1				1			1		1	1															
	Haljala			1				1	1	1	1		1	2			1												
	Kukruse							1	1	1			1	2				1											
	Uhaku			1				1	3	1			1	3			1	1								1			
	Lasnamägi			1				1	3	1			1	3			1	1								1			
	Aseri			1				1	3	1			1	1			1	1								1			
	Kunda						1	1	3	1				3	1		1	1											
	Volkhov						1	2	2	1				1			1	1											
	Billingen						2	3	3						2	2													
	Hunneberg		1			1	2	3	3																				
	Varangu		1			1	1	2	1																				
	Pakerort		3				1	?																					
CAMBRIAN		1	3																										

* - the number of the known families, in the brackets - the number in Estonia. Classification of Conodonta after Sweet (1988) and Aldridge & Smith (1993).

the Kuressaare, Kaugatuma and Ohesaare stages – consists mainly of three species and their subspecies *Ozarkodina confluens, O. remscheidensis* and *Oulodus elegans* (Viira 1982, Вийра 1983).

The Ordovician conodont faunas from Estonia belonged to the North Atlantic Province (cold-water realm). The conodont zone succession, generally accepted for this province, is applicable also in Estonia (Table 42).

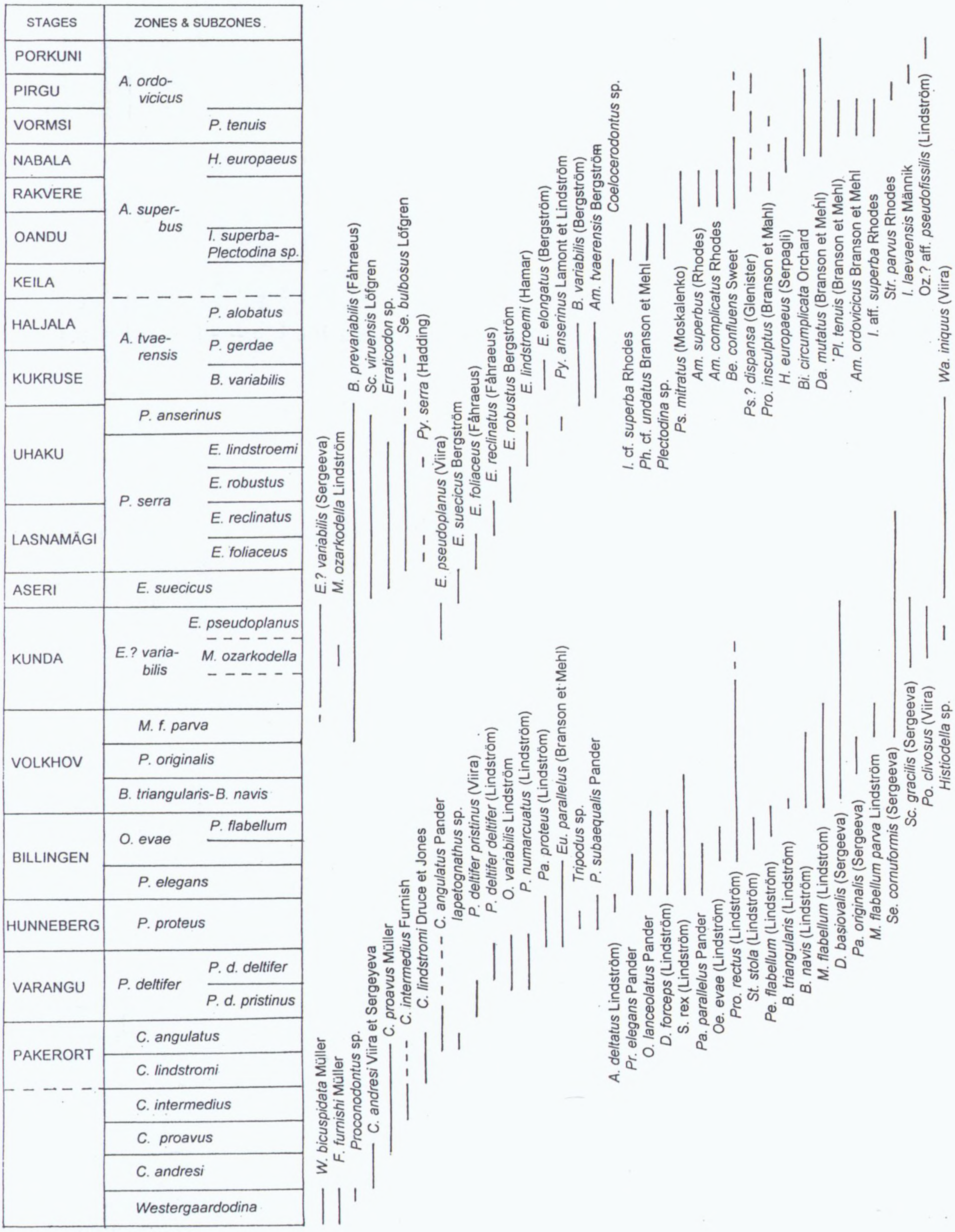
The world-wide conodont zonation for the Silurian System, recently revised, is basic for the local Estonian zonation. As many Silurian conodont species have evident ecological control, the near-shore zones have been additionally introduced (Table 43).

Early vertebrates

Silurian vertebrates (agnathans and fishes) of Estonia (Photo 43:1-3) have been studied relatively well. The first description of an osteostracan *Thyestes verrucosus* was published by Eichwald (1854). Later on several papers and monographs have dealt with the morphology and microstructure of the exoskeletons of agnathans and fishes, with their biostratigraphy, phylogeny and palaeoecology (Pander 1856, Rohon 1892, *etc*., Robertson 1938, *etc*., Hoppe 1931, Gross 1967, *etc*., Märss 1986, *etc*.).

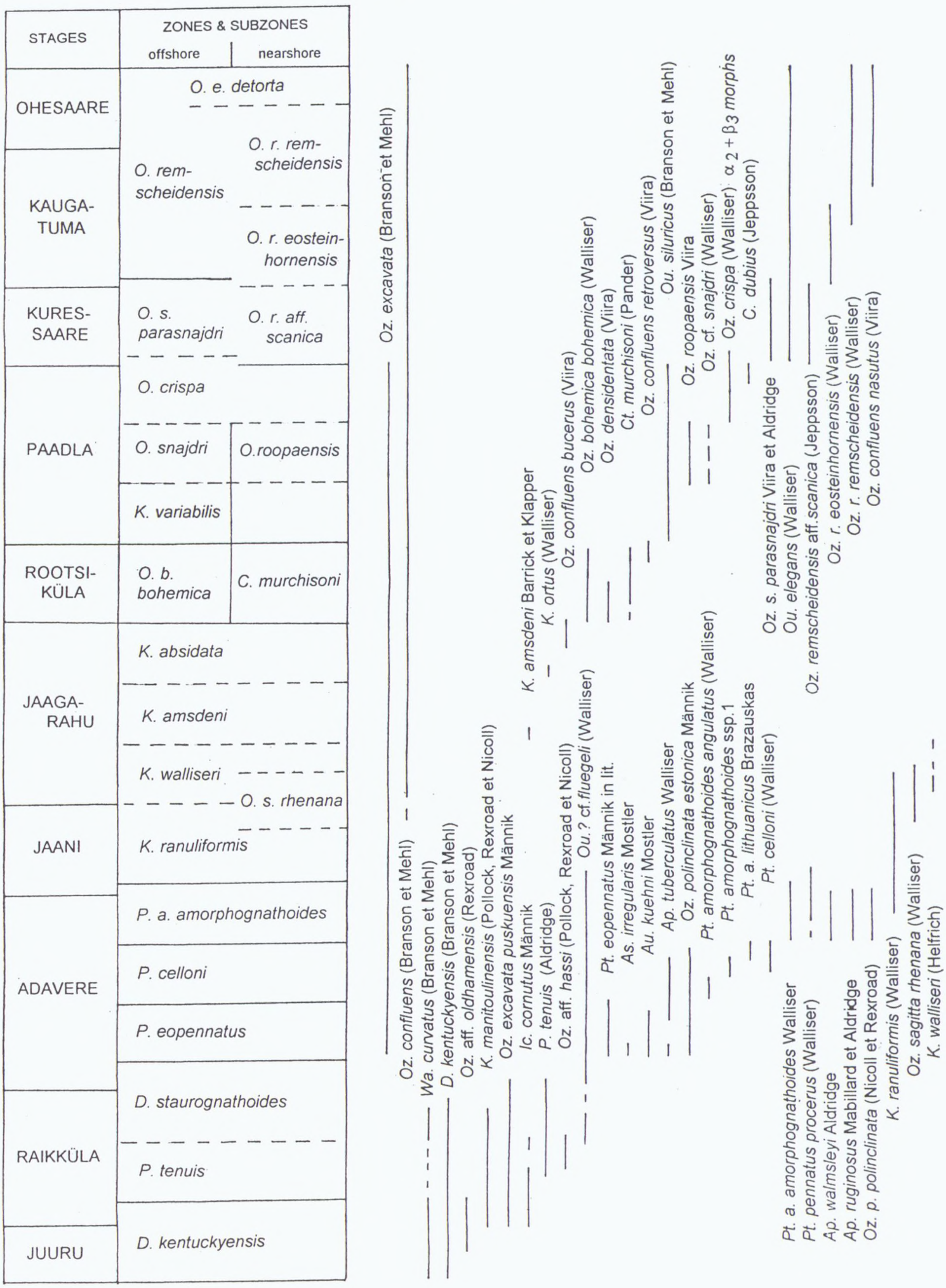
Estonian Silurian vertebrates were a well advanced and diverse group of animals containing several representatives of early agnathans and gnathostomes. The first finds come from the Jõgeva beds of the Raikküla Stage (Middle

Table 42. Distribution of the selected conodont species in the Ordovician of Estonia



Abbreviations of genera: *A. = Acodus, Am. = Amorphognathus, B. = Baltoniodus, Be. = Belodina, Bi. = Birksfeldia, C. = Cordylodus, D. = Drepanoistodus, Da. = Dapsilodus, E. = Eoplacognathus, Eu. = Eucharodus,*
F. = Furnishina, H. = Hamarodus, I. = Icriodella, M. = Microzarkodina, O. = "Oneotodus", Oe. = Oepikodus, Oi. = Oistodus, Oz. = Ozarkodina, P. = Paltodus, Pa. = Paroistodus, Pe. = Periodon, Pl. = Plectodina, Ph. = Phragmodus, Po. = Polonodus, Pr. = Prioniodus, Pro. = Protopanderodus, Ps. = Pseudooneotodus, Py. = Pygodus, S. = Scolopodus, Sc. = Scalpellodus, Se. = Semiacontiodus, St. = Stolodus, Str. = Strachanognathus, W. = Westergaardodina, Wa. = Walliserodus.

Table 43. Distribution of the selected conodont species in the Silurian of Estonia



Abbreviations of genera: *Ap.* = *Apsidognathus*, *As.* = *Astropentagnathus*, *Au.* = *Aulacognathus*, *C.* = *Coryssognathus*, *Ct.* = *Ctenognathodus*, *D.* = *Distomodus*, *Ic.* = *Irciognathus*, *K.* = *Kockelella*, *Oz.* = *Ozarkodina*, *Ou.* = *Oulodus*, *P.* = *Pranognathus*, *Wa.* = *Walliserodus*.

Llandovery). However, in other regions vertebrate remains have been recorded from much earlier strata. A problematic agnathan *Anatolepis* has been described from the Upper Cambrian to the Lower Ordovician rocks from Spitsbergen, Greenland and North America. The oldest confirmed agnathans, closely related to the heterostracans, come from the Lower Ordovician of Australia (Ritchie & Gilbert-Tomlinson 1977). The first thelodonts, another group of agnathans, have been described from the uppermost Ordovician of the Timan-Pechora Region (Karatajūtė-Talimaa 1996). Gnathostomes also appeared in the (?) Late Ordovician, but they became widely distributed beginning from the Early Silurian: acanthodians in the Baltic area (Märss 1986) and China (Pan Jiang 1988), elasmobranchs in Mongolia (Каратаюте-Талимаа и др. 1990; Каратаюте-Талимаа и Новицкая 1992), placoderms in the mid-Silurian of China (Pan Jiang 1988). The earliest osteichthyans - actinopterygians are known from the Late Silurian of Baltic (Märss 1986; Fredholm 1988) and the Central Urals.

Thus, the appearance of early vertebrates was a gradual process which lasted through the whole Ordovician. In the Early Silurian, all main agnathan groups (thelodonts, heterostracans, osteostracans, anaspids, galeaspids) already existed, and during the Silurian, representatives of all higher taxa of fishes (placoderms, acanthodians, chondrichthyans, osteichthyans) appeared. The Silurian was the period of several innovations and radiations in the evolution of early vertebrates (Kaljo & Märss 1991).

From the Silurian of the Baltic area, 55 vertebrate species belonging to 33 genera have been recorded (Table 44). During that period, three main evolutionary stages of vertebrates can be documented.

(1) The Raikküla - Jaagarahu evolutionary stage when single species of thelodonts, an osteostracan, an anaspid and acanthodians occur.

(2) The Rootsiküla - Paadla stage with a wide distribution of thelodonts, anaspids, osteostracans (in the Paadla Stage also heterostracans and actiopterygians).

(3) The Kuressaare to Ohesaare stage with a diverse vertebrate fauna of both agnathans and gnathostomes.

Using the distribution of vertebrate species in Estonia and Latvia, the regional biozonal succession has been compiled (Märss 1982, improved in 1990 and 1996, see also Table 8) which served as a base for the Global Vertebrate Zonal Standard (Märss *et al.* 1995).

The Silurian vertebrates, belonging to the nekton and nekto-benthos, inhabited all facies belts of the Palaeobaltic tropical sea: lagoonal, shoal, open shelf, transitional and depression (Märss 1986, 1989, 1990). The skeletal elements of the same vertebrate species occur both in the carbonate and terrigenous rocks that allows to treat the corresponding terrigenous sediments and fish remains in them as marine and not as continental (Märss 1989). During the Early Silurian and the beginning of the Late Silurian in the Palaeobaltic Basin the agnathans preponderated over gnathostomes, especially in the shallower parts of the sea. Beginning from the Ludlow in the deeper-water open shelf and transitional environments the gnathostomes prevailed, and during the Přidoli gnathostomes predominated in all facies belts.

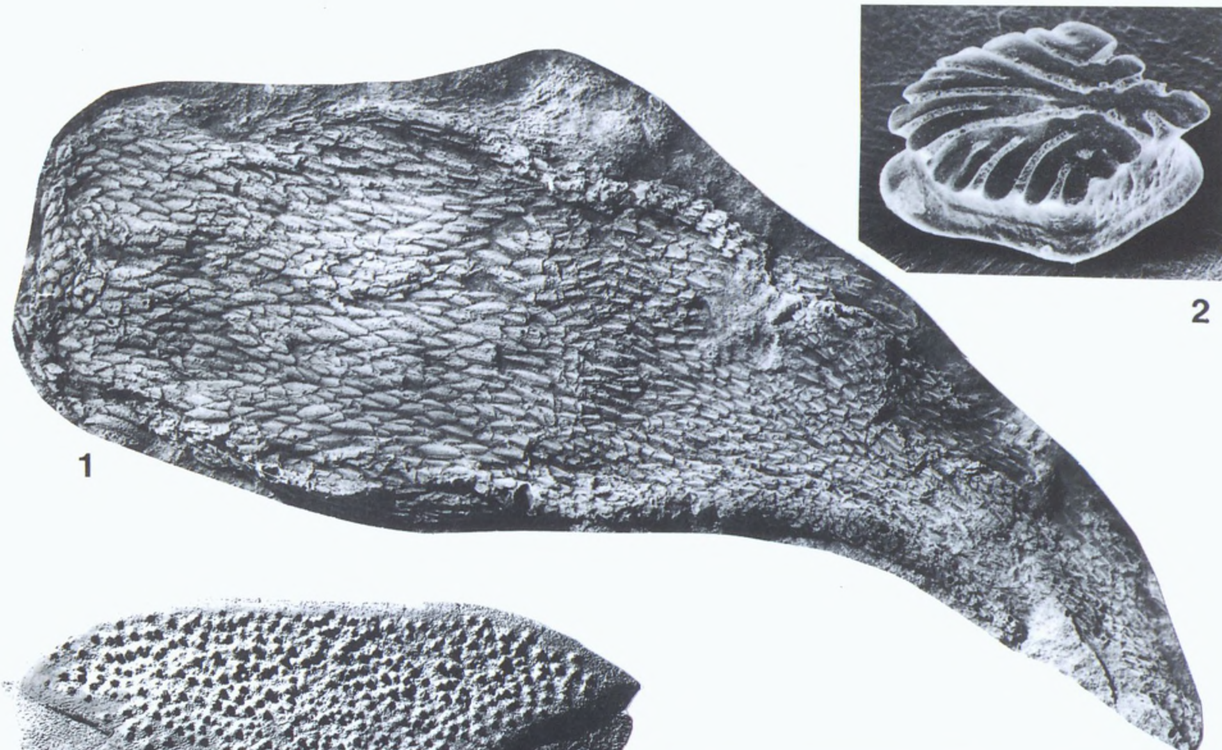
Judging by palaeogeographic reconstructions, the distribution of the Silurian vertebrates in the Baltic region, northern North America and northern Greenland shows that they were mainly tied to the tropics, occurring between latitudes 40 degrees north and south. Agnathans and fishes with different species dwelled also in the temperate zone (*e.g.* finds from southern Siberia and Bohemia). Five faunal provinces can be distinguish on the base of vertebrates: the eastern and northern Europe and Severnaya Zemlya, Siberia, Tuva, China, and North America (at least partly).

Devonian fishes

The study of the Devonian fishes in Estonia started in the fourth decade of the 19th century. During the current century the monographical studies were carried out by Gross (1933, 1940a,b, *etc.*) concerning all fossil fish groups, by Heintz (1928, 1930, 1934), Mark-Kurik (1973; Марк 1953a) and Obrucheva (Обручева 1962, 1966) on arthrodires, by Karatajūtė-Talimaa (1960, Каратаюте-Талимаа 1963) on antiarchs, by Obruchev and Mark-Kurik (1968, Обручев и Марк-Курик 1965) on psammosteid heterostracans, by Vorobyeva (Воробьева 1977) on crossopterygians and by Valiukevičius (Валюкявичюс 1985) on acanthodians

The fishes, known from Estonia, characterize mainly the flourishing period of the Devonian fish faunascoinciding with the Middle Devonian. Biogeographically, they belong to the eastern part of the Euramerica Province (Dineley & Loeffler 1993). All main groups, such as agnathans, placoderms, acanthodians, crossopterygians, dipnoans and actinopterygians, are present showing a great diversity. Dominating forms are psammosteid heterostracans which outside the NW of the East-European Platform are uncommon. Placoderms (arthrodires, antiarchs, Photo 43:6) are also numerous. Real giants (Photo 43:4-5) occur among the representatives ofthe above groups (*e.g.* psammosteids *Tartuosteus, Pycnosteus* and arthrodires *Homostius* and *Heterostius)*, but also in other groups (Crossopterygii, Dipnoi). Acanthodians are frequent members of fish assemblages. Tesserated cephalaspidids and actinopterygians, known by their microremains, seem to be quite common. Rare chondrichthyan remains have also been discovered. Decline of the Middle Devonian fish fauna can be observed roughly since the Gauja Age and further on when the assemblages became much less variable (Tables 45, 46).

During the last decade, biozonation based on different fish groups was worked out and included into the Devonian correlation chart of the East-European (Russian) Platform (Ржонсницкая и Куликова 1990). However, the zonal subdivision of the Devonian in Estonia and Latvia is largely based on publications by Gross (1942, 1950). Gross established the psammosteid, antiarch and some Upper Devonian acanthodian zones (Mark-Kurik 1993c). The acanthodian zonation has been elaborated by Valiukevičius (1994, 1995, Valiukevičius *et al.* 1995, *etc.*). A summary of the recent zonal schemes is given by Mark-Kurik (Blieck *et al.* in press; Table 47). The rather detailed psammosteid and placoderm (antiarch and arthrodire) assemblage zones, especially characteristic of the Middle Devonian, have been established on the basis of macroremains collected from exposures. The acanthodian bio- or interval zones, based on microremains (scales), are more universal as they can be distinguished using samples from both outcrops and drill cores. But the stratigraphical ranges of the Middle Devonian acanthodian zones, particularly those



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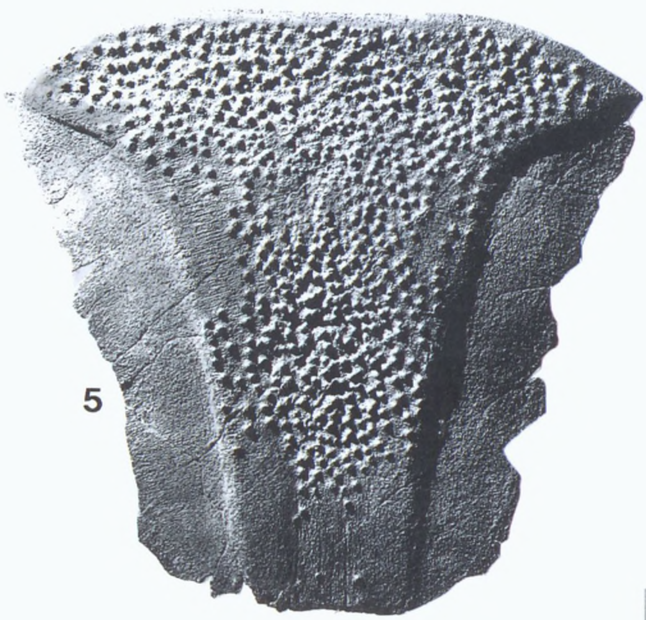

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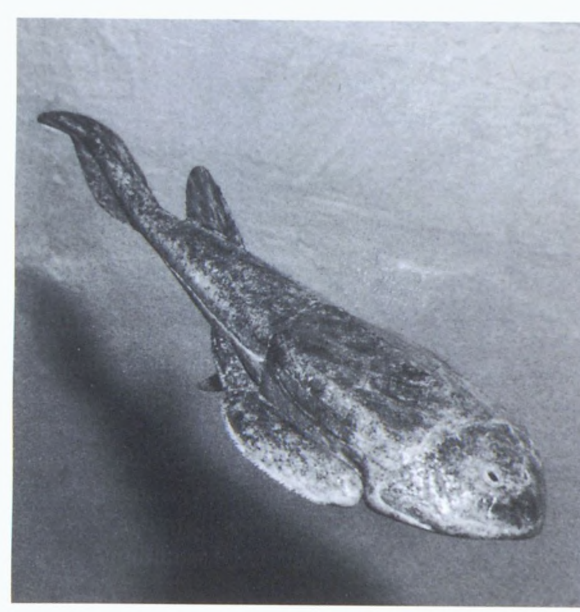

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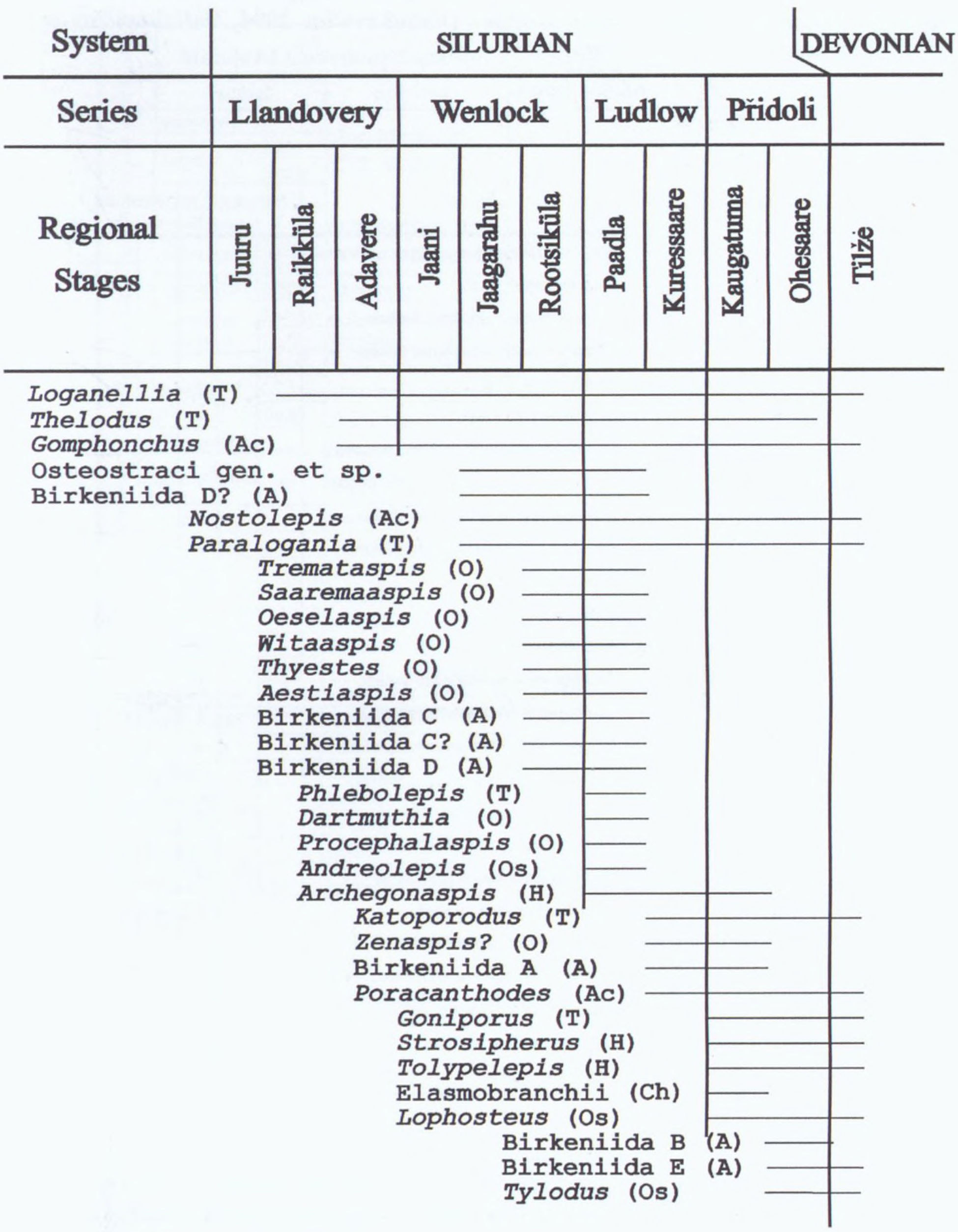
Table 44. Stratigraphical ranges of vertebrate genera in the Silurian of the Baltic area



Abbreviations in the brackets for agnathans: T - Thelodonti, H - Heterostraci, O - Osteostraci, A - Anaspida; abbreviations for gnathostomes: Ac - Acanthodii, Ch - Chondrichthyes, Os - Osteichthyes.

Photo 43. Silurian and Devonian agnathans and fishes:

1. Jawless fish (thelodont) *Phlebolepis elegans* Pander, Pi 6686, Ludlow, Paadla Stage, Himmiste-Kuigu, x 2.
2. Thelodont *Thelodus admirabilis* Märss, scale, Pi 6505, Ludlow, Kuressaare Stage, Sakla borehole, depth 6.2 m, x 60.
3. Thelodont *Katoporodus timanicus* (Karatajūtė-Talimaa), scale, Pi 6893, Přidoli, Ohesaare Stage, Ventspils borehole, depth 373.6 m, x 100.
4. Reconstruction of the Middle Devonian, Givetian, giant jawless fish (psammosteid heterostracan) *Pycnosteus tuberculatus* (Rohon); length of the fish about 2.5 m.
5. Placoderm fish (arthrodire) *Homostius latus* Asmuss, anterior median ventral plate, Pi 609. Givetian, Burtnieki Stage, Karksi, x 0.7.
6. Reconstruction of the Late Devonian, Early Frasnian placoderm (antiarch) *Bothriolepis;* length of the fish about 30 cm.

Table 45. Stratigraphical ranges of selected Middle and lower Upper Devonian fishes (from Сорокин и др. **1981, modified)**

	Aruküla			Burtnieki			Gj	Am	Pl
	vl	kr	tr	hm	kr	ab	gjS		plS
Tartuosteus giganteus (Gross)	+	+	+						
***Pycnosteus palaeformis* Preobr.**	+								
Ganosteus artus Mark-Kurik	+	+	+						
Psammolepis proia Mark-Kurik	+	+							
Cephalaspidida	+			+					
Actinolepis tuberculata Ag.	+	+							
Holonema obrutshevi Mark	+								
Homostius latus Asm.	+	+	+	+	+				
Heterostius ingens Asm.	+	+	+	+	+				
Coccosteus grossi O.Obr.	+	+							
Byssacanthus dilatatus (Eichw.)	+	+							
Asterolepis estonica Gross	+	+							
Gyroptychius pauli Vorob.	+	+	+						
Glyptolepis sp.sp.	+	+	+	+	+	+			
Dipterus sp.sp.	+	+	+						
Orvikuina sp.n.	+								
Cheirolepis sp.sp.	+					+			
Tartuosteus? luhai Mark-Kurik		+	+						
***Pycnosteus pauli* Mark**		+	+						
Nodocosta pauli Gross		+							
Thursius estonicus Vorob.		+							
Dipnoi gen.n.		+							
Hybosteus sp.		+							
Tartuosteus maximus Mark-Kurik			+?	+					
Millerosteus? sp.			+						
***Pycnosteus tuberculatus* (Rohon)**				+	+				
Ganosteus stellatus Rohon				+	+	+			
Psammosteus bergi (Obr.)				+					
Actinolepis magna Mark-Kurik				+		+			
Tropinema haermae (Mark)				+					
Coccosteus markae O.Obr.				+					
Asterolepis sp.1 Kar.- Talimaa				+					
Gyroptychius elgae Vorob.				+					
Glyptolepis ? karksiensis (Vorob.)				+					
Holoptychiidae				+					
Psammolepis sp.sp.				+	+	+			
Nodocosta sp.					+				
Byssacanthus sp.sp.					+	+			
Hamodus lutkevitshi Obr.					+	+			
Panderichthys? sp.					+	+			
Psammolepis abavica Mark-Kurik						+			
Psammosteus sp.sp.						+			
***Watsonosteus* sp.n.?**						+			
Livosteus? sp.						+			
Plourdosteus? panderi O.Obr.						+			
Asterolepis essica Lyarskaya						+			
Microbrachius cf.*dicki* Traq.						+			
Chondrichthyes?						+			
Laccognathus sp.						+			
Moythomasia? sp.						+			
Psammolepis paradoxa Ag.							+		
P. venyukovi Obr.							+		
P. alata Mark-Kurik							+		
P. heteraster Gross							+		
Plourdosteus livonicus (Eastm.)							+		
***Asterolepis ornata* Eichw. sensu Ag.**							+		
Glyptolepis baltica Gross							+		
Laccognathus panderi Gross							+	+	
Megadonichthys kurikae Vorob.in litt.							+		
Psammolepis undulata (Ag.)								+	
Psammosteus praecursor Obr.								+	
Ps. maeandrinus Ag.								+	+
Asterolepis radiata Rohon								+	
***Bothriolepis prima?* Gross**								+	
Panderichthys rhombolepis Gross								+	
***Bothriolepis cellulosa* Pand.**								+?	+
Ctenurella pskovensis (Obr.)									+
Grossilepis tuberculata (Gross)									+
Moythomasia perforata (Gross)									+

Index species are in bold characters. vl, kr, tr: Viljandi, Kureküla and Tarvastu beds of the Aruküla Stage; hm, kr, ab: Härma, Koorküla beds and Abava Substage of the Burtnieki Stage; Gj, *gjS*: Gauja Stage and Sietiņi Member of the Gauja Stage; Am, Amata Stage; Pl, *plS*, Pļaviņas Stage and Snetnaya Gora Member of the Pļaviņas Stage.

Table 46. Stratigraphical ranges of the Middle Devonian acanthodians (Valiukevičius 1994, Valiukevičius pers. comm.)

	Narva			Aruküla			Burt-nieki
	V	L	K	vl	kr	tr	hm
***Cheiracanthoides estonicus* Valiuk.**	+						
Acanthodes? sp.C	<+						
Cheiracanthus crassus Valiuk.	+						
Rhadinacanthus balticus Gross	+	+	+	+			
Acanthodes? sp.B	<+	+	+	+	+	+	+ >
Acanthodes? sp.D	<+	+	+	+	+	+	+ >
Cheiracanthus brevicostatus Gross	<+	+	+	+	+	+	+
Cheiracanthus longicostatus Gross	<+	+	+	+	+	+	+
Ptychodictyon distinctum Valiuk.		+	+			+	
***Ptychodictyon rimosum* Gross**		+	+	+	+	+	+
Cheiracanthus sp.		+	+				
Diplacanthus sp.		+	+	+			>
Acanthodes? sp.A		+	+	+	+	+	+ >
Cheiracanthus intricatus Valiuk.			+				
***Nostolepis kernavensis* Valiuk.**			+				
Cheiracanthoides proprius Valiuk.			+				
Markacanthus costulatus Valiuk.			+	+			
Minioracanthus laevis Valiuk.			+	+	+		
Ptychodictyon sulcatum Gross			+	+	+	+	+
Diplacanthus carinatus Gross			+	+	+	+	+
Acanthodii gen.n. Valiuk.			+	+	+	+	+
Markacanthus alius Valiuk.				+	+	+	+
Rhadinacanthus multisulcatus Valiuk.				+	+	+	+ >
***Diplacanthus gravis* Valiuk.**				+	+	+	+ >
Nostolepis sp.n. Valiuk.						+	+
?*Ptychdictyon* sp.							+

Index species are in bold characters. “<“, “>” show an earlier and later occurrence, respectively. *V, L, K*: Vadja, Leivu and Kernave members of the Narva Stage; vl, kr, tr: Viljandi, Kureküla, Tarvastu beds of the Aruküla Stage; hm, Härma beds of the Burtnieki Stage.

Table 47. Fish zones and their correlation with the standard conodont zones, eastern European miospore zones and Baltic regional stages (from Blieck *et al.* in press, modified)

	STAGES	Standard Conodont Zonation: actual		Standard Conodont Zonation: previous		Miospore zones	Baltic regional stages	Thelodont and heterostracan zones	Placoderm and acanthodian zones	
UPPER	FRASNIAN	*jamieae*		*Anc. triangularis*		SD	Daugava	*Psammosteus megalopteryx*	*Bothriolepis trautscholdi*	
		hassi	Late							
			Early	unzoned						
				asymmetricus	Upper					
		punctata			Middle		Dubniki			
		transitans			Lower		Pļaviņas		*Bothriolepis cellulosa*	
		falsiovalis	Late		Lowermost	BI				
			Early				Amata		*B. prima, B. obr.*	*Devononchus concinnus*
MIDDLE	GIVETIAN	*disparilis*	Late / Early	*disparilis*		IM	Gauja		*Asterolepis ornata*	
		hermanni-cristatus	Late	*hermanni-cristatus*	Upper					
			Early		Lower		Abava		*Watsonosteus*	
		varcus	Late	*varcus*	Upper	EX	Burtnieki	*Pycnosteus tuberculatus*	*Diplacanthus gravis*	
			Middle		Middle					
			Early		Lower		Arukūla	*P. pauli* *P. palaeformis*		
		hemiansatus		*ensensis*	Upper					
	EIFELIAN	*kockelianus*			Lower	RL	Narva: Kernavė	*Schizosteus striatus*	*Coccosteus cuspidatus*	*Nostolepis kernavensis*
				kockelianus						
		australis		*australis*			Narva: Leivu		*Ptychodictyon rimosum*	
		costatus		*costatus*			Narva: Vadja		*Cheiracanthoides estonicus*	
		partitus		*partitus*		PT	Pärnu	*Schizosteus heterolepis*	*Laliacanthus singularis*	
LOWER	EMSIAN	*patulus*		*patulus*		DI	Rēzekne	*Skamolepis fragilis*		
		serotinus		*serotinus*						
		inversus		*laticostatus/inversus*			?			
		nothoperbonus					?			
		gronbergi		*gronbergi*		AB		*Amaltheolepis baltica*	*Gomphonchus tauragensis*	
		dehiscens		*dehiscens*						
	PRAG.	*pireneae*		*pireneae*			Ķemeri	*Rhinopteraspis dunensis*		
		kindlei		*kindlei*						
		sulcatus		*sulcatus*		E	?			
	LOCHKOV.	*pesavis*		*pesavis*						
		delta		*delta*		Z	Stoniškiai	*Rhinopteraspis crouchi*	*Lietuvacanthus fossulatus*	
		woschmidti-postwoschmidti		*woschmidti-postwoschmidti*		G, M, N	Tilžė	*Phialaspis*	*Nostolepis minima*	

Key to the miospore zonation: AB, *Emphanisporites annulatus* Z.; BI, *Acanthotriletes bucerus - Archaeozonotriletes variabilis insignis* Subz.; DI, *Diaphanospora inassueta* Z.; E, *Dictyotriletes eminensis* Z.; EX, *Geminospora extensa* Z.; G, *Emphanisporites zavallatus* var. *gedinniensis* Z.; IM, *Ancyrospora incisa - Geminospora micromanifesta* Subz.; M, *Emphanisporites microornatus* Z.; N, *Cymbosporites proteus* Z.; PT, *Periplecotriles tortus* Z.; RL, *Rhabdosporites langii* Z.; SD, *Geminospora semilucensa - Perotriletes donensis* Z.; Z, *Emphanisporites zavallatus* var. *zavallatus* Z. Abbreviations: *B, Bothriolepis; B. obr., Bothriolepis obrutschewi; P, Pycnosteus;* Lochkov., Lochkovian; Prag., Pragian.

of the late Middle Devonian, are comparatively long.

In some cases, the dependence of the content of the fish assemblages on the character of the bottom of the basin (sandy, muddy) is rather apparent. The representatives of psammosteids and cephalaspidids obviously preferred sandy bottom (Валюкявичюс и др. 1986). In establishing the completeness of the faunal assemblages, the reconstruction of trophic relations was used (Mark-Kurik 1995). Incomplete assemblages may be caused by the taphonomic loss. In the relatively complete assemblages various secondary and tertiary consumers were represented beginning with the top consumer, the largest predator.

The stratigraphical ranges of fish species, mostly the Middle Devonian ones, are given in Tables 45 and 46. Among psammosteids and placoderms (Table 45) the species confined to one or two stratigraphical units of lower rank (members, beds) are rather numerous whereas the acanthodians (Table 46) show a limited number of species with short range.

BEDROCK TOPOGRAPHY

The Estonian bedrock is cuesta-like and more complicated in southern Estonia. The maximal amplitudes of absolute heights of the bedrock surface are close to those of the modern topography, being minimal in the ancient valleys (Harku Valley -145m) and maximal in the Haanja Heights (+166m).

History of research

In Estonia, the peculiarities of the bedrock topography are easier to study than in neighbouring areas, because the Quaternary cover is rather thin, especially in northern Estonia (Fig. 158).

The research into the Estonian bedrock topography goes back to well over a century. The first information on the bedrock topography was issued by Schmidt (1854, 1883) who described the cuesta-like topography of Estonia with abrupt northern and rather flat southern slopes. Later, main attention focused on single relief forms, particularly the escarpments. The Sub-Cambrian Peneplain (Giere 1938, Fromm 1943), Cambrian and Silurian escarpments (Martinsson 1958, Jaanusson 1944a), Vendian (Witting 1910, Nilsson 1913, Büchting 1918), Ordovician (Tammekann 1928, 1940, Martinsson 1958, Künnapuu 1959, Miidel 1992) and Silurian (Jaanusson 1947, Aaloe 1958, Aaloe & Miidel 1967) escarpments were described in particular detail.

The first scheme of the bedrock topography was compiled by Tammekann (1928, 1949). More detailed zonation, based on the lithological complexes, was carried out by Orviku (Орвику 1955). He distinguished several large features, including the Depression of the Gulf of Finland, the North-Estonian Escarpment (Photos 2, 18), the North-Estonian Plateau with the Pandivere Upland, the Middle-Estonian Lowland, the Middle- and Upper-Devonian plateaus, the depressions of Võru-Piusa, and lakes Võrtsjärv and Peipsi. A large quantity of information about the bedrock topography has accumulated in the Geological Survey of Estonia which was summarised by Kajak (Каяк 1966). He was the first to show the isolines of the bedrock topography after every 20 m. To the large features defined by Orviku, he added the West-Estonian Lowland and several medium forms, including the Ahtme Elevation and the Luuga-Narva and Ojamaa lowlands. On a more detailed map compiled by Miidel (Раукас и др. 1971), the isolines were drawn after every 10 m. In 1977, a map of the bedrock topography of Estonia was composed by Tavast on a scale of 1:200 000 (Tavast & Raukas 1978). It was based on the data from nearly 2500 boreholes and dugholes, but also on the materials of geophysical evidence and earlier maps. Some information has also been obtained on the bedrock topography of the coastal areas of the Baltic Sea (Amantov *et al.* 1988, Тавaст и Амантов 1992). In the Gulf of Finland, the Sub-Vendian, Vendian-Cambrian and Ordovician escarpments were distinguished. Owing to the geological mapping of Estonia, there are rather top-quality maps available on the bedrock topography of many regions (Saaremaa a.o.) but, unfortunately, only the bedrock map in the scale of 1:2 500 000 (Kajak 1995) has been published until now.

Genesis of the bedrock topography

The bedrock topography of Estonia has developed under the effect of different geological processes. In the course of long-term continental period, the development of bedrock surface was controlled by erosional processes, and by the beginning of the Pleistocene, a complicated bedrock topography had formed. In the Pleistocene, the bedrock topogra-

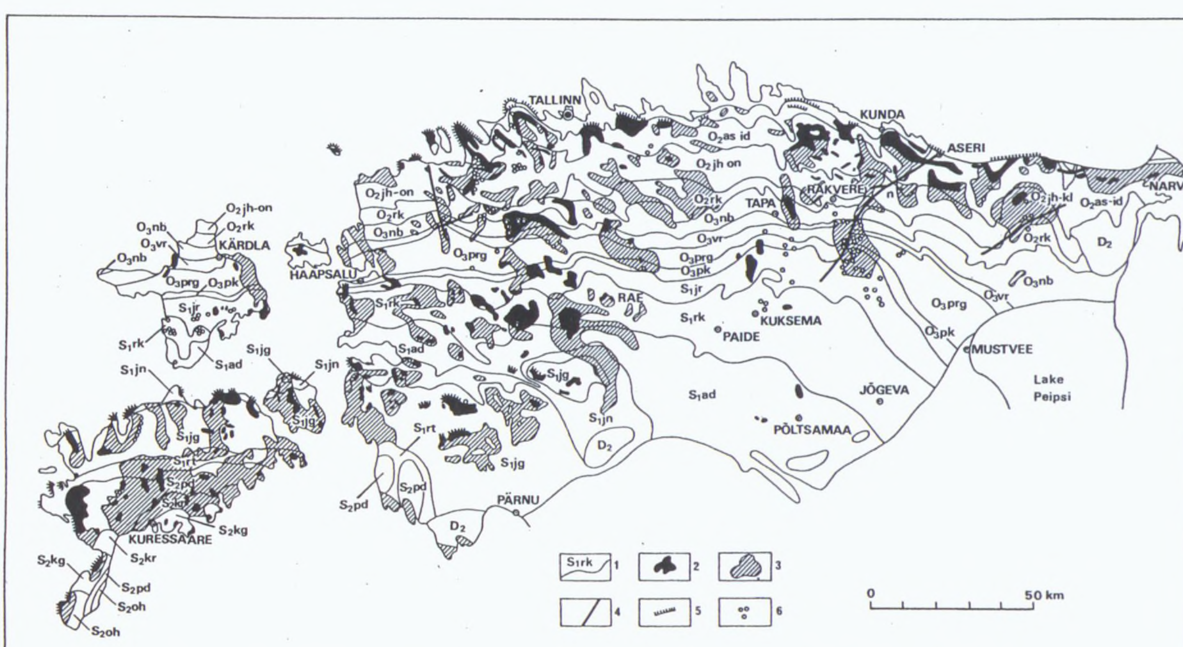


Fig. 158. Distribution of alvars (after Jürgenson & Tavast 1986, with complements): 1 - boundaries between outcrops of bedrock stages with stratigraphical indices of the stages (see Ch. IV 3.4); 2 - alvars; 3 - outcrops of the bedrock covered with a thin (up to 30 cm) layer of Quaternary deposits; 4 - zones of tectonical disturbances; 5- cliffs; 6 - karst forms.

phy was significantly affected by glaciers; in the Holocene the waters of the Baltic Sea (Photo 44) and several terrestrial geological agents played an important role.

The bedrock topography has been affected by both passive and active geological factors, such as tectonic movements, heterogeneity of the lithological composition of the bedrock, erosional, gravitational and other geological processes, eluvial and karst phenomena, cosmic factors and human activities.

Some researchers point out the leading role of the tectonic movements (Doss 1913, Николаев 1962, Ряхни 1973), others maintain that the main relief forming factor was the long-term denudation on the lithologically heterogenous bedrock (Schmidt 1883, Hausen 1913a, Марков 1931б). Many researchers (Giere 1938, Fromm 1943, Tammekann 1949) stress the importance of the fluvial erosion in forming the cuesta-like topography of the bedrock.

A rather big group of scientists gives prevalence to glacial erosion (Grewingk 1879, Lewinski & Samsonovicz 1918, Исаченков 1976). Giere (1932) maintains that the North-Estonian Klint had retreated under the influence of glacial erosion from the line of Suur-Pakri - Lavansaari - Seiskari to the present position.

According to the calculations by Issachenkov (Исаченков 1982) and Makkavejev (Маккавеев 1976), the Pleistocene glaciers had removed a layer of rock up to 50 - 60 m in thickness from the bedrock surface on the southern slope of the Fennoscandian Shield. In the ice lobe depressions, the thickness of such a layer could have reached even 100 metres.

Most of contemporary researchers are of the opinion that the bedrock surface has formed under the influence of different geological processes (Таваст и Раукас 1982).

Two or three stages (Giere 1932, Tammekann 1949, Martinsson 1958) can be distinguished in the history of the formation of the bedrock topography. In the pre-Quaternary period, active tectonical movements and long-term denudational processes prevailed.

In the Pleistocene, the genesis of the bedrock topography was influenced by glacial erosion and erosion by interglacial rivers and the Eemian Sea. Wave erosion of different stages of the Baltic Sea and denudational continental processes changed the bedrock topography in post-glacial time. During all geological epochs, the tectonic movements have played an important role in the formation of the bedrock topography, which is revealed through: 1) the regional sinking or decelerated denudation; 2) the differentiated movements of the blocks of the crystalline basement which formed new morphostructures; 3) geological structures which have already lost their tectonical activity (faults, crevice zones). The latter forms promoted the selective denudation of the bedrock.

Photo 44. In the Holocene, the bedrock topography was significantly affected by the waves of the Baltic Sea. The profile on the Island of Osmussaar displays the heterogeneity of the lithological composition in the lower part of the limestones of the Lasnamägi Stage and upper part of the Uhaku Stage. *Photo by Karl Orviku.*

We have compared the block structure of the Estonian basement, bedrock and contemporary topography (Tavast & Vaher 1982, Raukas *et al.* 1988, Вахер и Таваст 1979) and found that direct morphostructures are infrequent in our area, and such local structures which could correspond in the plan and sign to large forms of bedrock topography are not observed in the Estonian territory.

For example, the most prominent topographic phenomenon of the area - the Haanja Elevation, is located in the eastern part of the Mõniste-Lokno plain-type fold. On the crest of the anticline the basement lies at a depth of 230 m below sea level, descending northwards to 500 m, and southwards to 1000 m below sea level. According to the distribution of Vendian and Lower Palaeozoic rocks in southeastern Estonia, this structure was formed principally in the Late Silurian (Вахер и Таваст 1979).

The southern part of the elevation intersects with the high-dipping slope of the anticline which amplitude accounts for more than 120 m on the floor of the Upper Devonian Pļaviņas beds. The height of the northerly dome on the above-mentioned floor amounts to 30 m. Thus, the amplitude of relatively young movements does not exceed 30 m. It forms only about one third of the relative height of the Haanja Heights in view of its bedrock surface, and less than 15 m of the relative height of topographic elevation. Consequently, these were exogenic processes, not tectonic movements, that dominated in its shaping (Raukas *et al.* 1988).

The question is what role, if any, linear dislocations played in the formation of the bedrock and recent topography. Extremely dense network of boreholes in the area of Estonian oil shale deposit permitted Vaher (Вахер и Таваст 1979) to draw the contour lines for that region at 1 m intervals. However, with one exception, even in this case expression of the linear structures in the bedrock topography was not detected in the recent topography of northeastern Estonia. If these structures were activated by inherited postglacial movements, some attack on the denudated bedrock surface must have occurred though it may well be invisibly minute.

Summarizing the above, we should like to stress that the formation of the bedrock and contemporary topography has been only little affected by postglacial and recent blockwise differential movements.

Although most of linear structures and destruction zones are not reflected in the bedrock topography, they have influenced on the formation of some single forms or relief complexes. As an example serves the slope of the Pandivere Elevation in the destruction zone (Рыук и Таваст 1982). Miidel (Мийдел 1966б,1971) measured the direction of tectonic joints in the North-Estonian river valleys crossing the Ordovician Escarpment. The results were compared with the direction of the river valleys in the places where measurements had been carried out. In all cases, tectonic joints coincided with the direction of the valleys which suggests that the linear tectonic disturbances played an important role in the formation of modern river valleys. In the same areas, the ancient valleys and karst phenomena are often connected with tectonic joints (Heinsalu 1977). Such connection is quite understandable, because the linear erosion and karst processes are naturally more intensive in the destruction zones of the bedrock.

In distinguishing the role of tectonic movements in the genesis of the bedrock topography, we have to mention two different aspects: the indirect and direct. The general rise of the Earth crust created the conditions for the erosion-denudation processes and determined their intensity. Even the dead tectonic structures (joints, zones of disturbances) are highly conducive to the denudation and dissolution processes. The role of the differentiated neotectonic movements in the genesis of the single relief forms of the bedrock topography was rather modest, however, these movements had a great influence on the formation of the bedrock topography over vast areas.

In the development of all active endogenic geological processes the lithological heterogeneity has been and will be of very great importance, since it accounts for uneven glacial erosion, denudation, the specific arrangement of the river drainage, the distribution and morphology of karst forms, *etc.* (Tavast & Raukas 1978, Вахер и Таваст 1979). All meridionally elongated depressions and lowlands (Middle Estonian, Võrtsjärv, Peipsi, the Gulf of Riga and Vidzeme) have developed in the easily erodable Middle Devonian siltstones, clays and sandstones.

The lithological heterogeneity of the bedrock affected not only the genesis of large and medium relief forms, but it also controlled the formation of small ones (Photo 23). Already Schmidt (1854, 1883) called attention to the long-time denudation in the formation of hard rock hillocks elongated from the north-west to south-east. The height of such hillocks is usually less than 10 m. The areas between the hillocks (Winkler 1920, Tammekann 1928) are composed of less durable limestones.

The intensity of erosional processes in pre-Quaternary time, during interglacial and in late- and postglacial times depended on many factors, such as the amount of water, the geological structure and the shape of the catchment area, the inclination of the valleys, the stream velocity, and the climate.

The most impressive example of the pre-Quaternary river erosion is the ancient buried valleys. Most of them probably formed in the Late Palaeogene when the Earth's crust was much higher than at present due to the riftogenesis in the North-Atlantic area (Пуура 1980). A complicated net of valleys was formed in and around the contemporary Baltic Sea.

There are several schemes of Estonia's buried ancient valleys compiled by many researchers (Tammekann 1928, 1949, Каяк 1970, Раукас и др. 1971, Tavast 1992).

Marine and lacustrine erosion affected the bedrock topography during interglacials and late- and postglacial times. Due to neotectonic movements, which were more intensive in the northwestern than in the southern part of Estonia, the shoreline was subject to rather rapid changes here and only during the transgressive phases of the Baltic Sea, when the rise in the water-level more or less equalled the uplift of the Earth crust, the position of the shoreline somewhat stabilized and the erosion of the bedrock was considerable (Кессел и Раукас 1967).

By our opinion, the wave erosion was one of the essential factors responsible for the formation of alvars (Photo 26), in the development of which three main stages can be distinguished: pre-glacial, Pleistocene and post-glacial (Jürgenson & Tavast 1986). During the first stage, erosion and denudation were intensive. During the second stage, erosion and accumulation of glaciers were the most characteristic processes. In the post-glacial stage, the Baltic Sea with its transgressions and regressions participated in the formation of alvars. After the final retreat of the sea from Estonia, various less important continental processes followed.

Human activities have affected upon the bedrock topography considerably during the last decades, especially in the regions where phosphorite, oil shale, clay and limestone are mined. Man has also influenced onto the bedrock topography by dredging rivers and laying out the foundation for constructions. As the bedrock topography is comparatively flat, the construction of high- and railways hasn't brought about remarkable changes in the topography, however, this effect tends to increase. For instance, a 5-m-deep and several-kilometres-long channel with a width of approximately 50 m, has been blasted into the Ordovician rocks at Lasnamäe in Tallinn.

The zonation of the bedrock topography of Estonia is presented on the basis of relative and absolute heights, taking into consideration the lithological composition of the bedrock. In the area under consideration the following large forms were proposed (Tavast & Raukas 1978, Tavast 1992, Figs.159, 160): I - Depression of the Gulf of Finland, II - Viru-Harju Plateau, III - Livonian Lowland, IV - Middle-Devonian Plateau; V- Central-Estonian and Lake Võrtsjärv depressions, VI - Lake Peipsi Depression, VII - Valga and South-Estonian lowlands, VIII - Upper-Devonian Plateau. Also the medium (Pandivere and Sakala elevations, Ordovician, Silurian (Photo 23) and Devonian escarpments), small (elongated eminences and hollows) and tiny forms (ice scratches) are distinguished.

To sum the above up, it should be pointed out that the bedrock topography has formed under the effect of different geological factors. Unlike some other geological factor, it has continuously affected the formation of the Quaternary deposits and landforms, being thus responsible for the inherited nature of the litho- and morphogenesis during the Quaternary as a whole (Tavast 1992).

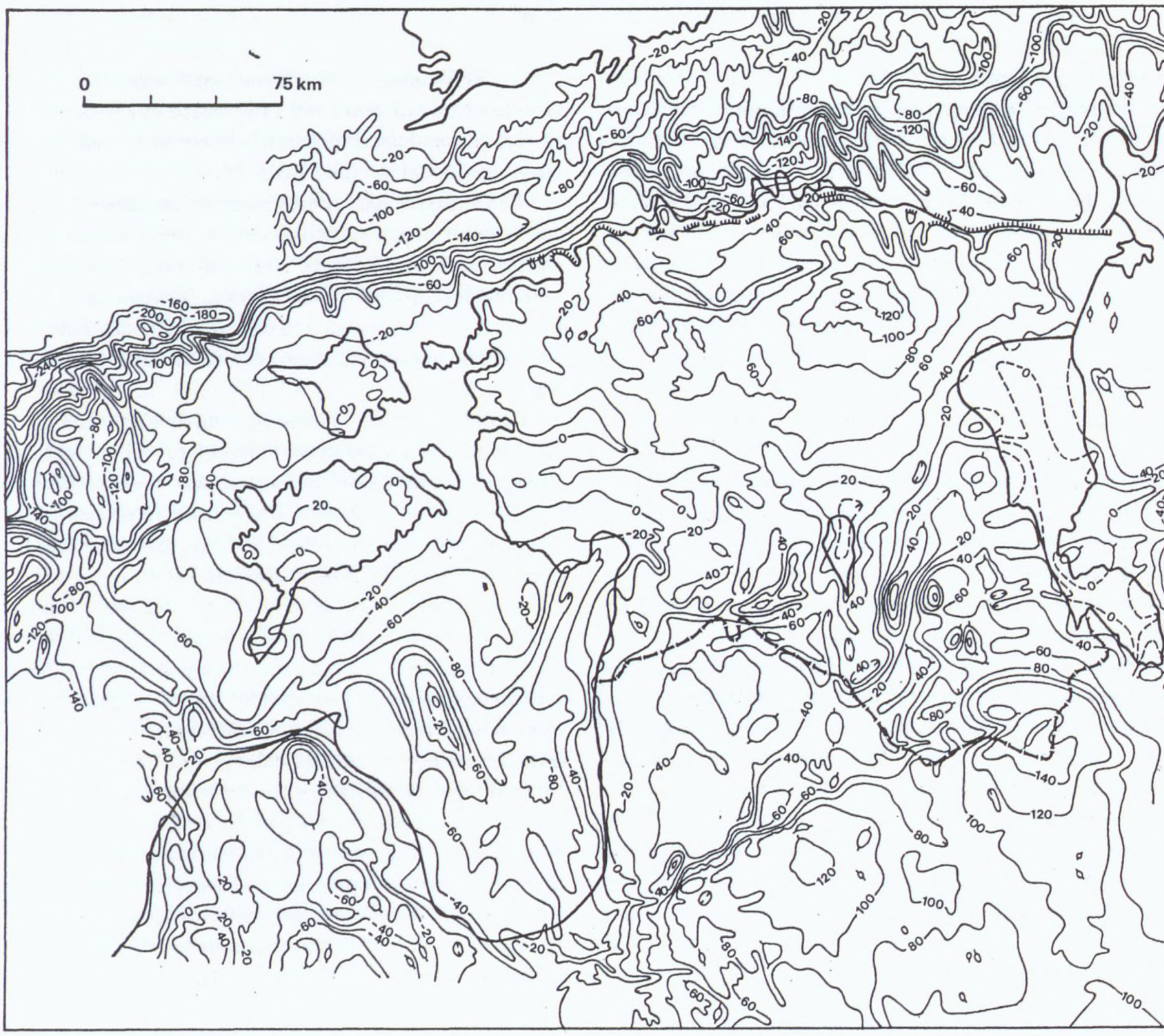


Fig.159. Isolines of the bedrock topography, drawn after 20 m.

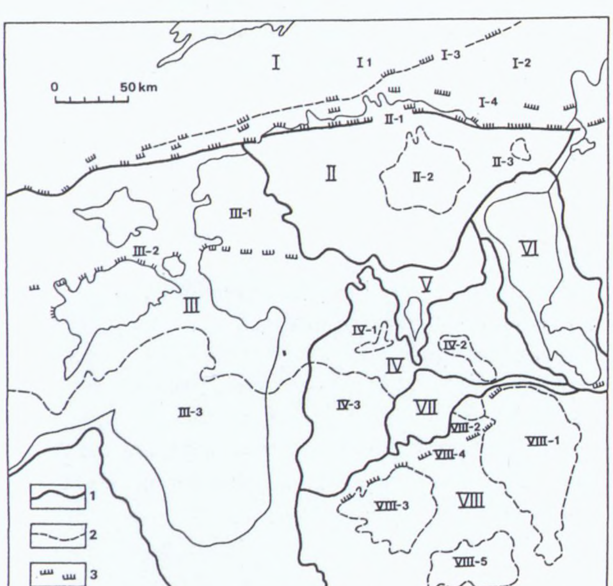


Fig.160. Zonation of the bedrock topography (after Tavast 1992): 1 - boundaries of large bedrock forms; 2 - boundaries of medium bedrock forms; 3 - escarpments. I - Depression of the Gulf of Finland: 1 - Sub-Vendian Peneplain; 2 - Cambrian-Vendian Peneplain; 3 - Vendian Escarpment; 4 - Cambrian Escarpment. II - Viru-Harju Plateau: 1 - Ordovician Escarpment; 2 - Pandivere Elevation; 3 - Ahtme Eminence. III - Livonian Lowland: 1 - West-Estonian Lowland; 2 - Depression of the Gulf of Riga. IV - Middle-Devonian Plateau: 1 - South-Sakala Elevation; 2 - Otepää Elevation; 3 - North-Vidzeme Depression. V - Central-Estonian Lowland and Depression of Lake Võrtsjärv. VI - Lake Peipsi Depression. VII - Valga and South-Estonian lowlands. VIII - Upper-Devonian Plateau: 1 - Haanja-Aluksne Elevation; 2 - Karula Eminence; 3 - Vidzeme Elevation; 4 - Devonian Escarpment; 5 - East-Latvian Lowland.

Quaternary glaciations covered on the East-European Plain a vast territory, extending over polar and subpolar regions in the north and reaching as far as the Don and Dnieper rivers in the south. The deposits of the Lower Pleistocene are absent in Estonia and even the Middle Pleistocene sequence is rather uncomplete. Nevertheless, Estonia was one of the first regions where the theory of continental glaciations was elaborated, and the concepts of the glacial litho- and morphogenesis worked out here have contributed to solution of topical questions concerning the dynamics of ice sheets and formation of glacial landforms and deposits.

During all glaciations, Estonia was affected by the Baltic and Peribaltic ice streams which moved at different rates during different glaciations and stadials of glaciations. For example, the lithology of till beds and the orientation of clasts in tills suggest that the ice flow direction during the Late Ugandi (Warthe) and Valgjärve (Early-Middle Weichselian) glaciations was mainly from north-west to south-east, during the Early Ugandi (Saale) and Late Sangaste (Elster II) glaciations from north to south. There were naturally different local movements depending upon the bedrock topography.

Areas of accumulation and erosion remained relatively stable through time. Ancient valleys, interlobate massifs, lee-side areas of bedrock elevations, as well as escarpments oriented transverse to the ice movement, acted as areas of accumulation (Таваст и Раукас 1982). Intensive erosion took place on bedrock elevations and in ice lobe depressions, and the Quaternary cover is correspondingly very thin.

Alternation of till beds with interglacial and interstadial sediments has been controlled by the cyclically changing climate. At the beginning of all glaciations, the centre of glaciation was in the Scandinavian mountains owing to which circumstance a northwest-southeast direction of the ice movement was prevalent. As climate deteriorated further, the centre of the glaciation extended eastwards and a north-south movement of the ice became dominant. This direction is recorded by the distribution of indicator boulders in the tills. At the end of the glaciations when climate improved, the glaciation centre was transferred once again to the Scandinavian mountains, and the direction of the ice movement was again northwest-southeast (Raukas 1961).

The bedrock topography exerted an influence over the distribution of ice-marginal depositional zones which are mainly connected with the slopes of bedrock elevations and with the depressions in the bedrock. Various types of glacial landforms are also related to certain elements of the bedrock topography, which facilitates the determination of their genesis.

Drumlins (Photo 3) most frequently form large crag and tail formations on the distal side of bedrock elevations, *e.g.* the Saadjärve Drumlin Field where before the drumlinization thick older deposits occurred (Каяк 1965б), and in depressions where the glacier desintegrated into lobes which moved at different speeds (Rõuk & Raukas 1989).

End moraines are usually distributed on the proximal slopes of elevations (Photo 45) blocking the movement of glaciers and thus causing the accumulation of till. Marginal eskers (narrow deltas) are mainly spread in bedrock depressions. Radial eskers occur under different topographic conditions, frequently in ancient valleys. Glaciofluvial deltas are characteristic of regions with the bedrock surface inclined towards the ancient glaciers (northern Estonia). Hilly glacial relief (Photo 46) is widely distributed in areas with a rapidly changing bedrock topography (Таваст и Раукас 1982). In southeastern Estonia, the formation of hilly landscapes was greatly affected by Middle Pleistocene sediments which in the inner part of the heights are much thicker than in the surrounding lowlands (Каяк 1965a, Kajak 1995).

The evidence obtained by analysing the internal structure of the basement and bedrock, the thickness of different facies within the sedimentary cover complexes, and the maps showing the structure and contours of different key beds, provide the basis for the conclusion (Raukas *et al.* 1988) that the tectonic framework has considerably and repeatedly changed with

Photo 45. End moraines often occur on proximal slopes of bedrock elevations. Tamsalu-Naistevälja push moraine (North Estonia) with poorly rounded particles of the local bedrock. *Photo by A. Miidel.*

Photo 46. Hilly glacial relief is widespread in areas with a rapidly changing bedrock topography. Otepää Heights. *Photo by B. Murd.*

time. The block structure of the basement is weakly reflected in the bedrock and glacial topography. Linear elements, such as faults in the basement, are identified as the most active and long-lasting forms, but even they are rather poorly reflected in the bedrock and present topography. For instance, the Haanja Heights, which is the highest region in the Baltic Republics, is situated on the tectonically active Mõniste basement rise, but low rates of neotectonic uplift and the relatively thick Quaternary cover show that the tectonic movements have not been decisive in the formation of this highland area (Таваст и Раукас 1982).

Although the concept of the extinction of Pleistocene glaciers by way of stagnant ice fields separated from the margin of the active glacier was known already before the turn of the century, there nevertheless, still exist disagreements as to the extension and thickness of such dead ice fields (cf. *e.g.* discussion in Karukäpp & Raukas 1976).

The vast majority of investigators believe that the formation of stagnant ice fields was determined, first of all, by the change in the dynamics of the glaciers due to climatic conditions: ablation exceeded the inflow of the ice from the accumulation zone.

Since the hilly topography is mainly distributed in southern Estonia (Photo 45), it was assumed for a rather long time that the glacial relief of southern Estonia was formed chiefly due to the effort of stagnant ice, whereas that of northern Estonia was shaped by active ice (Каяк 1963). At that, it was supposed (Ряхни 1963б) that the glacier lobes in northern Estonia continued to be active practically until their final melting. Nowadays these assumptions have been refuted and the significant role of both stagnant and active ice in both regions mentioned is quite clear (Карукяпп 1979).

In a very generalized form, Estonia's territory may be divided into lobe depression regions with a rather flat topography reflecting the stadial and phasial halts of the glacier margin, and ice-shed regions of insular heights with a rather complicated topography which have been formed during the course of a long time interval under the influence of active, passive and dead ice (Аболтиньш и др. 1989).

Most likely, the waters of the Holsteinian Sea did not reach Estonia. Continental deposits of Holsteinian (Likhvin) age have been established in the southwestern (Karuküla section) and southeastern (Kõrveküla section) Estonia (Liivrand 1984).

The Karuküla organic deposits, overlain by a thin (1 m) reddish-brown till of the last glaciation, rest on the grey till of Middle-Pleistocene age. According to Liivrand, the Holsteinian organogenic deposits at the Karuküla site are erratics. According to Kajak (1995), these deposits are glaciotectonically compressed and disturbed, but in autochtonous bedding.

Description of the section (from top downwards):

Layers	Thickness in metres
Reddish-brown till, upper part is weathered	1.00 - 2.00
Yellowish unsorted sand with inclusions of organic matter	0.10 - 1.25
Forest peat with pieces of wood	0.15 - 0.70
Forest peat/*Phragmites* peat transition layer	0.05 - 0.15
Phragmites peat	0.25 - 0.40
Gyttja	0.05 - 0.60
Yellowish-grey silty clay	0.10 - 0.35
Unsorted sand or grey till	0.50 +

Pollen zones in the Karuküla section after Reet Pirrus (Орвику и Пиррус 1965) and Liivrand (1984, 1991) are the following (Fig. 161):

- K_1 - lower part of gyttja-*Betula (Betula nana)* and *Pinus.* Rise of *Picea, Alnus* and *Ulmus* curves;
- K_2 - upper part of gyttja-*Picea* and *Alnus* maxima. *Quercus, Ulmus, Trapa natans* and spores of *Osmunda* are present. Immigration of *Tilia*;
- K_3 - *Phragmites* peat - *Tilia, Quercus* and *Ulmus* maxima. *Picea* is frequent;
- K_4 - lower part of forest peat - *Picea, Abies* and *Carpinus* maxima;
- K_5 - upper part of forest peat - *Betula* and *Pinus* with *Picea. Alnus* and broad-leaved species are present.

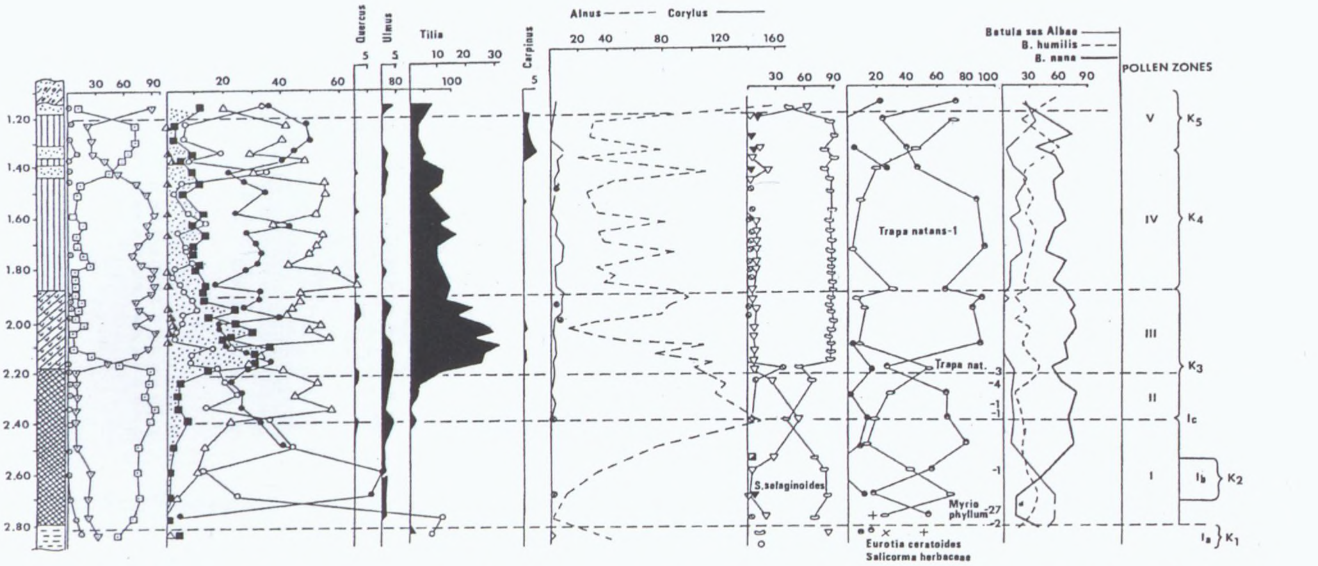


Fig. 161. Simplified palynological diagram of the Holsteinian continental sediments of the Karuküla section (after Liivrand 1984). For legend see Fig. 162.

The pollen zones of the Karuküla section (Fig. 161) correspond to the pollen zones of Holsteinian deposits in Latvia and Lithuania (Liivrand 1984, 1991). They all demonstrate a wide spread of conifers, an early appearance of *Picea* and *Alnus* (maxima occurred already before the climatic optimum) and a very limited occurence of *Corylus*.

Holsteinian deposits in the Kõrveküla section occur within an old buried valley near Tartu. They are represented by gyttja and loam, covered by reddish-brown till and underlain by glaciofluvial deposits. The palynological spectrum coincides with that of the Karuküla section (Лийвранд и Саарсе 1983).

As mentioned above, both marine and continental Eemian deposits are found in Estonia. Interglacial deposits were supposed to occur in the basin of the Gulf of Finland already long before they were discovered in the Rõngu section (Orviku 1939). Thomson (1934), in his work dealing with the finds of mammoths in northern Estonia, pointed out that all the remains of these animals (molar teeth at Paljassaar, Pirita, Ülemiste and Lüganuse; incisiors and shin-bones at Ihasalu) occurred well preserved in the fore-klint area. This is indicative of their short transportation, and refers indirectly to the occurrence of interglacial layers north of the klint. Basing on the drilling data by Mickwitz (1908) and Öpik (1929) on the islands of the Gulf of Finland where submorainic organogenous layers were reported, Zans (1936) correlated the layers with the Portlandia Sea (Skærumhede) sediments of the Eemian Interglacial in Denmark (Jessen & Milthers 1910).

Abundant places of gas emanation in the bottom deposits of the Gulf of Finland were also indicative of submorainic organogenous layers, *e.g.* between the Viimsi Peninsula and Malusi Island at Viinistu (Eplik 1935), on Keri Island and in several other places. On the Island of Keri, gas was used in household and in the lighthouse.

Drilling on the Island of Prangli (Fig. 91) brought some clarity to the stratigraphy of Pleistocene deposits in the Gulf of Finland (Kajak 1961). Interglacial deposits composed of greenish-grey silts and comprising subfossil molluscs and plant remains with vivianite were discovered at a depth of 66.0 - 77.7 m in seven boreholes. They are underlain by waterlain glacial deposits and two beds of brownish till with different thicknesses (6 and 15 m), and overlain by four grey tills (see the description of the section). Interglacial deposits occur in an area of about 2.5 km^2 and are best preserved in borehole 6 (see the description below) which was chosen for a stratotype section (Raukas & Kajak 1995).

All the four Järva tills are separated by thin layers of waterlain glacial deposits, in places (between the upper two tills) by varved clays with plant remains which has enabled Kajak (1961) to speak about interstadial sediments.

In the area of the Gulf of Finland, the stratigraphic subdivision of the Pleistocene deposits is hampered by the limited distribution of interglacial and interstadial deposits and abundant erratics. Interglacial deposits in a secondary position stand out because of their differing lithological and palaeontological composition in the sections studied. The recognition of erratics derived from marine sediments is aided by a study of their hypsometric position. Marine Eemian interglacial deposits in the vicinity of St. Petersburg, Lake Ladoga and on the coast of Luga Bay are located above the coastline of that interglacial, which is a clear indication of their secondary position (Малаховский и Саммет 1982). Only in the Prangli section they seem to be *in situ.*

Pollen diagrams, compiled and interpreted by Liivrand (Лийвранд 1972, Лийвранд и Вальт 1966, Liivrand 1984, 1991 a.o) show the whole cycle of development of vegetation between the Late Ugandi (Wartha) and Early (or Middle) Järva (Weichselian) glaciations (Fig. 162). Three complexes are most distinct:

(1) spore-and-pollen spectra of late-glacial deposits of the Late Ugandi glaciation characterizing varved clays and the lower part of marine sediments;

Description of borehole 6 on Prangli Island

Layer	Thickness in metres
Unsorted sand with gravel and pebbles, m_{IV}	0.00 - 6.00
Sandy loam with varved lamination, $l_{III}vr$	6.00 - 20.80
Grey varved clay, l_{III} vr	20.80 - 25.30
Dark-grey dust-like greenish till, $g_{III}vr$	25.30 - 32.00
Till as above, but richer in crystalline rocks, g_{III} vr	32.00 - 43.35
Grey, greenish loam, l_{III} sv	43.35 - 45.50
Dark-grey dust-like greenish till, g_{III} vl	45.50 - 55.00
Dark-grey greenish sandy till, g_{III} vl	55.00 - 63.15
Light- and dark-grey greenish varved clay, $l_{III}kl$	63.15 - 63.75
Grey sand, of medium grain size, $l_{III}kl$	63.75 - 65.85
Greenish-grey dust-like loam with weakly pronounced horizontal and lenticular lamination. In the upper part (65.85 - 67.60 m) coarse quartz sand with the grains of vivianite arranged in layers. Sandy loam at a depth of 69.0 - 71.9 m, fragments of molluscs at a depth of 71.50-74.40, plant remains at a depth of 74.40 - 76.51 m, $m_{III}pr$	65.85 - 78.25
Brown varved clay, l_{II} ug_3	78.25 - 79.45
Yellowish-grey sand with varying grain-size, l_{II} ug_3	79.45 - 83.15
Greyish-brown till, g_{II} ug_3	83.15 - 94.80
Brown varved loam, l_{II} ug_2	94.80 -107.70
Greyish-brown till, g_{II} ug_1 which rests directly on the Proterozoic crystalline rocks.	107.70 -123.00

(2) spore-and-pollen spectra of marine Prangli (Eemian) interglacial deposits;
(3) spore-and-pollen spectra of early glacial deposits and the upper part of marine deposits.

Late-glacial deposits of the Late Ugandi glaciation are characterized by the following pollen zones (Fig. 162):
LUg_1 (Ms_1) - *Betula nana* L., abundant herb pollen (up to 70 %). The herbs are dominated by *Artemisia* (up to 60 %) and *Chenopodiacea* (up to 30 %). *Eurotia ceratoides* and *Polycnemum* are constantly present. The quantity of pine pollen increases at the end of the zone. The formation of varved clays and the lowermost part of grey loam is referred to that zone.
LUg_2 (Ms_2) - Pollen zone of tree-like birch and pine. Substantial decline in *Betula nana* L. and herbaceous plants. The zone indicates a short-term climatic warming.
LUg_3 (Ms_3) - Birch pollen zone. Birch pollen dominates the spectra (95%) consisting mainly of *Betula nana* L. (90%) which refers to approaching climatic cooling.

After the Late Ugandi glaciation, the climate was cold and very dry (cryoxerotic stage).

On the Island of Prangli, the marine interglacial deposits are characterized by the same pollen zones (Fig. 162) which Jessen and Milthers (1928) differentiated in the Eemian interglacial deposits in western Europe (zones c - i) and Grichuk (Гричук 1961) in the Mikulinan interglacial deposits on the East-European Plain (zones M_2 - M_8).

c(M_2) - Birch and pine pollen zone. Pollen of tree-like birch prevails. B. *nana* L. and B. *humilis* Schrank occur in small amounts.
d(M_3) - Pine and birch pollen zone. Spruce disappears, hazel and afterwards also alder appear.
e(M_{4a}) - Pine and birch pollen zone. Oak and elm also occur. The proportion of hazel and alder increases.
f(M_{4b}) - Oak and elm pollen zone. The lower maximum of hazel and alder.
f(M_5) - Willow pollen zone. The second half of hazel and alder maximum.
g(M_6) - Hornbeam pollen zone. The proportion of hazel and alder decreases.
h(M_7) - Spruce pollen zone. Abrupt decline in thermophilous species.
i(M_8) - Pine pollen zone. Hazel, alder and hornbeam continue to occur in very small amounts.

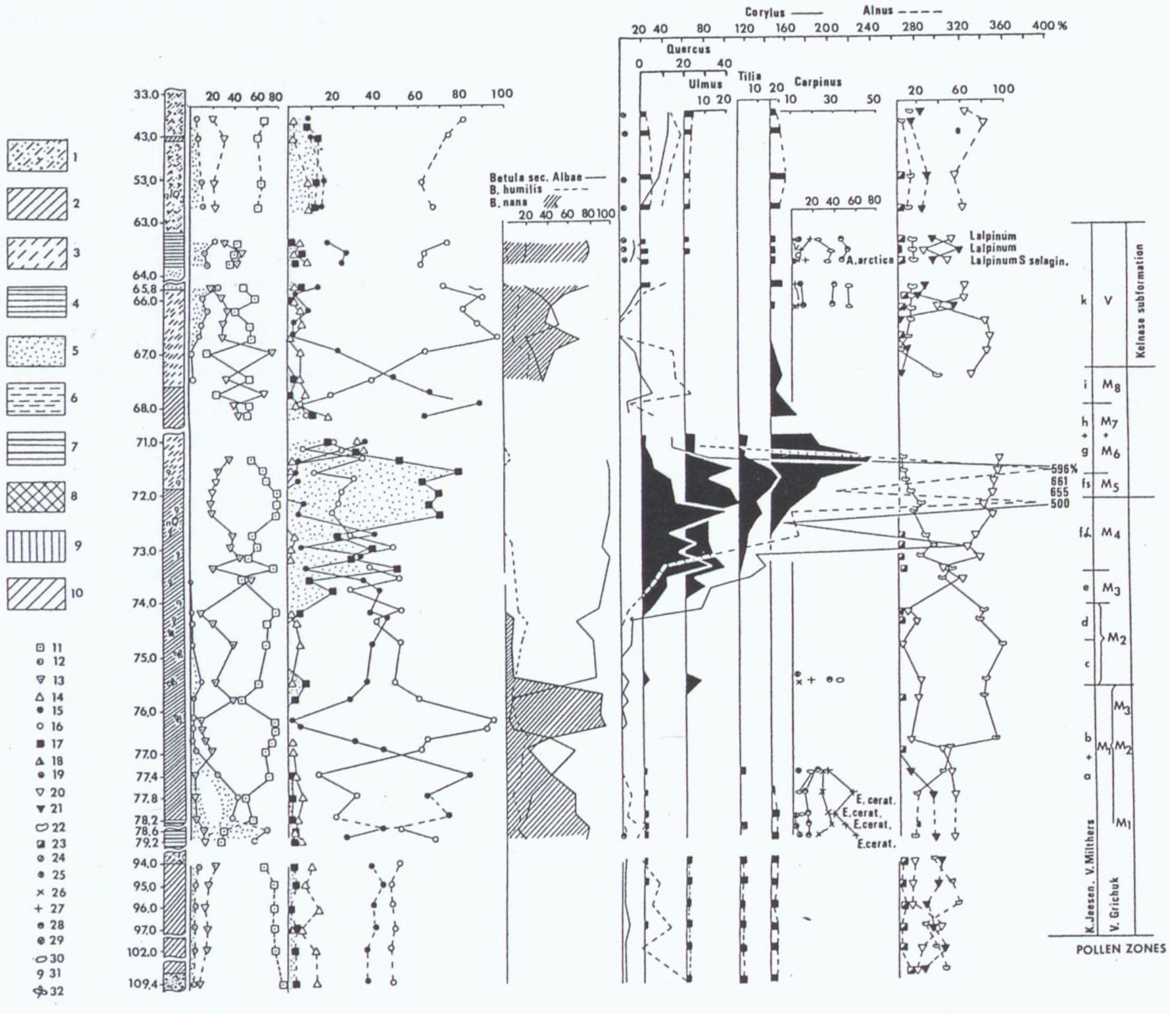


Fig. 162. Simplified palynological diagram of the Eemian marine sediments of the Prangli section (after Liivrand 1984): 1 - till; 2 - loam; 3 - sandy loam; 4 - clay; 5 - sand; 6 - silt; 7 - varved clay; 8 - sapropelite; 9 - forest peat; 10 - *Phragmites* peat; 11 - composite pollen of trees; 12 - composite pollen of herbaceous plants; 13 - spores; 14 - fir; 15 - pine; 16 - birch; 17 - composite pollen of broad-leaved trees; 18 - *Abies*; 19 - willow; 20 - *Bryales*; 21 - *Sphagnum*; 22 - ferns; 23 - *Lycopodium*; 24 - *Equisetum*; 25 - herbs; 26 - chenopods; 27 - *Artemisia*; 28 - *Graminae*; 29 - heather; 30 - *Carex*; 31 - molluscs; 32 - plant remains.

The Eemian interglacial is characterized by two equally well-pronounced climatic stages - thermoxerotic and thermohygrotic. In the former stage, oak and elm were widely spread, whereas willow, hornbeam and spruce pollen maxima are typical of the latter stage.

According to Liivrand (1984, 1991), the further development of the vegetation is shown by the appearance of herbaceous plant associations and tundra species which are indicative of the effect of the starting climatic cooling of the Järva (Weichselian) glaciation (Fig. 162)

To this period is attributed the formation of the upper part of marine sediment complex and that of grey varved clays of the Kelnase Subformation (Table 13).

$Q_{III}\,jr_1kl$ - Birch pollen zone. Strong representation of *Betula nana* L. Herbaceous plants are represented by sedges and loose-bunch grasses. In varved clays the pollen of *Artemisia arctica* (Cham.) Wallr. and the spores of *Lycopodium alpinum* L. and *Selaginella selaginoides* (L.) Link occur. The climate was cold and humid (cryohygrotic stage).

Diatom analyses, performed on the interglacial deposits from borehole 4 (Черемисинова 1961, Знаменская и Черемисинова 1962), discovered two stages in the development of the Eemian Sea which correspond to two diatom complexes. One of them characterizes the lower part of the deposits and the beginning of transgression. In the composition of the diatom complex there have been identified cold-favouring fresh-water relict forms, such as *Cocconeis disculus* (Shum.) Cl., *Diploneis domblittensis* (Grun.) Cl. a.o., and the fresh- and salt-water forms *Pinnularia* sp., *Epithemia* sp., *Neidium* sp., but also marine shallow-water forms, such as *Hyalodiscus scotius* (Ktz.) Grun., *Actinocyclus Ehrenbergii* Ralfs, *Grammatophora* sp. and others. According to Čeremisinova, the co-occurrence of offshore-marine species with cold-water relict forms marks the start of the sea-water's intrusion into the glacial lake. The late-glacial pollen zones a+b correspond to this period.

The second complex of diatoms is typical of the period of interglacial transgression and is related to the middle part of marine loams at a depth of 60.5 - 67.5 m below sea level. The diatom complex indicates normal salinity of water, but there are shallow-water forms, such as *Melosira sulcata* (Ehr.) Ktz., *Actinocyclus Ehrenbergii* (Bail) Ralfs, *Hyalodiscus scoticus* (Ktz.) Grun, *Grammathophora* sp. a.o. Besides, there also occur typical thermophilous Eemian species: *Synedra Gaillonii* (Bory) Ehr., *Navicula abrupta* Greg., *Coscinodiscus antiquus* Grun., *C. granulosus* Grun., *C. perforatus* Ehr. According to the pollen evidence, this period corresponds to the interglacial unit characterized by pollen zones M_2-M_8.

Sea water intruded into the Gulf of Finland early, already in the post-glacial of the Late Ugandi Glaciation and marine conditions existed here throughout the whole Eemian interglacial. Figure 163 shows the two principally different constructions of the distribution of the Eemian Sea (Raukas 1991a).

According to Grichuk, the Baltic Sea was connected with the North Sea *via* the Skagerrak, Kattegat and Danish Sounds through the present lake system of Vänern and Mälar in Central Sweden and the area of the current Kiel Canal on the Jutland Peninsula (Герасимов и Величко 1982). Like Lavrova (Лаврова 1961), he assumes a connection between the Eemian and White Sea basins through the system of shallow sounds and the lakes of Onega and Ladoga. This means that the Eemian and Boreal transgressions must have been synchronous, as demonstrated by dating techniques including the ESR-method (Molodkov & Raukas 1988).

According to the second reconstruction, the contours of the Eemian sea closely coincided with the Litorina Sea limit and therefore this basin of water could have been linked with the ocean only via the Skagerrak, Kattegat and Danish Sounds. Transgressive waters would have inundated the Lake Ladoga depression and small areas of the Vistula River valley (Благоволин и др. 1982). To our mind, the first version seems more reliable.

In terms of pollen zones, the marine interglacial sediments on Prangli Island are in good agreement with the other well-known Eemian interglacial deposits in the St. Petersburg (Mga section, a.o.) and other regions. In several other parts of the Gulf of Finland, the seismoacoustic evidence indicates the three-fold division of the Pleistocene deposits (Kiipli *et al.* 1993), analogical to that discovered on Prangli Island. This is also reflected in the occurrences of tills of two glaciations with marine Eemian deposits in many supra-aquatic (islands of Eksi, Rammusaar, Malusi) and sub-aquatic (banks of Kuradimuna, Bezymyannaja, Moksei a.o.) drumlins, but also in the area of the Island of Mosčnyi. The layers under consid-

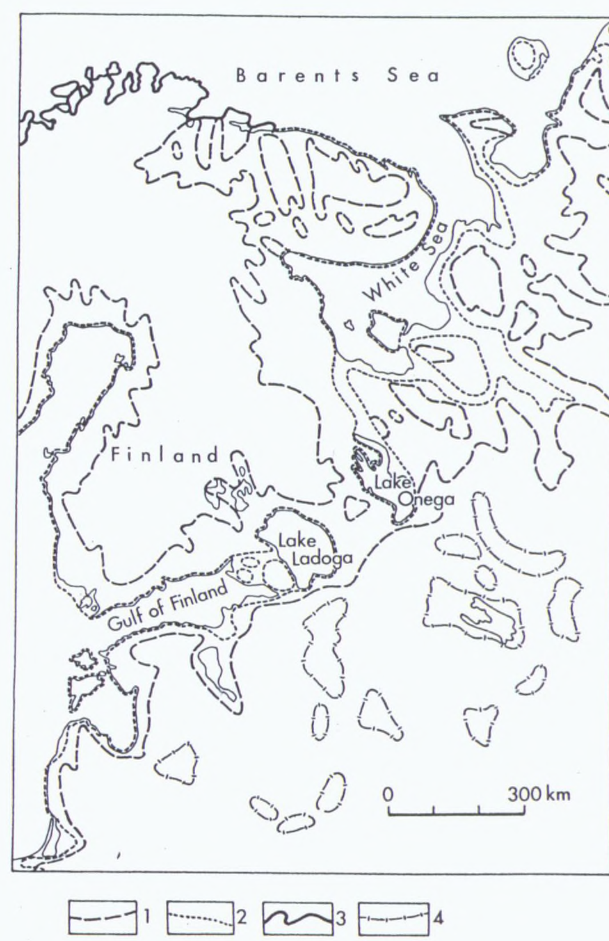


Fig. 163. Eemian shoreline and submerged areas : 1 - after Grichuk (Гричук 1961) and Lavrova (Лаврова 1961); 2 - after Blagovolin, Leontjev and Muratov (Благоволин и др. 1982); 3 - contemporary shorelines; 4 - Eemian inland bodies of water according to Grichuk (after Герасимов и Величко 1982) with complements.

eration are of almost horizontal bedding; the top of the intermorainic layers is located at a height of 60-70 m and the base at a depth of 75-90 m below sea level.

Very complicated and still unsolved is the post-Eemian geological history of the territory under consideration. Contradictory opinions have been expressed on the Early Weichselian stage. Finnish researchers maintain that in the Early Weichselian their territory was free from ice; in the Baltic countries the vast majority of investigators are of the opinion that at that time there was a thick ice sheet, almost as large as during the Late Weichselian.

The oxygen-isotope curves from deep-sea sediments show that the volumes of water, stored in glaciers, were almost equal in the first and second halves of the Late Pleistocene. This does not mean that the development of the glaciers was synchronous in different areas. On the basis of geomorphological indications it is rather difficult to reconstruct palaeoglaciological parametres, and further, the potentials of geochronological methods are too limited to allow unambiguous conclusions.

In the former Soviet Union, several investigators (I. Krasnov, E. Zarrina, A. Raukas, L. Serebryannyi) maintained that the glaciation culminated on the East-European Platform in the Early Weichselian when the Brandenburg (Bologoe) marginal moraine was formed, others (M. Makaryčeva, N. Čebotareva, A. Veličko a.o.), on the other hand, either deny an Early-Weichselian advance of the ice sheet or consider it to have been of a rather limited distribution. According to Nilsson (1973), the glaciers did not reach Scania in southern Sweden at that time. In Finland, a Lower-Weichselian till is represented by a dark-grey clayey variety (Rainio & Lahermo 1984), correlated with the Suintio Till (Bouchard *et al.* 1990). The so-called purplish-grey till in southern Estonia is referred to the Lower Weichselian and so is the lower grey till overlying the Eemian in the fore-klint area and in the North-Estonian Plateau, and also the related waterlain glacial deposits (Раукас 1978). Lower-Weichselian tills are also found in Latvia and Lithuania. In the new stratigraphical scheme of Estonia (Table 131) the purplish-grey till belongs to the Lower Weichselian Valgjärve Subformation (Raukas & Kajak 1995). Near Valgjärv, this bed overlies Eemian sediments (Kajak 1995).

Some authors maintain that instead of a single increase in the volume of the ice sheet and its melting in the study area there were numerous rapid fluctuations of the ice margin in the Early Weichselian. Based on pollen diagrams, these authors have differentiated in northwestern Russia three distinctly differing cycles in the development of the vegetation in the Early Weichselian (Малаховский и Спиридонова 1981), reflecting independent warm oscillations. The first, Upper-Volga cycle was characterized by the prevalence of a boreal vegetation. In the succeeding, Tosna cycle alongside with the above-mentioned complex, nemoreal species were spread, whereas in the third, Berezai cycle a hypoarctic vegetation dominated. These interstadials are separated from each other by Kurgolovo, Kileshi, Bologoe and Jedrovo stages, whose pollen spectra show the prevalence of nonarboreal plants. At first, spores dominated and then the proportion of herb pollen increased. From a geological and geochronological point of view it may, however, be concluded that these interstadials are not yet sufficiently grounded, and that their pollen spectra are not clear enough.

The radiocarbon ages are beyond the limits of the method, and the spore and pollen data are difficult to use in the correlations.

Detailed studies have been performed on classical Early-Weichselian interstadials in the Netherlands and Denmark. In Denmark, the Brørup interstadial deposits, characterized by pollen spectra of birch and pine forests with spruce (Andersen 1961) were studied beyond the boundaries of the Weichselian glaciation, however, not far from its margin. The preceding Rodebæk interstadial deposits, which are embedded between solifluction layers in the same region, are characterized only by pollen of dwarf birch, juniper and herbs. Further south, in the Netherlands, spruce and pine forests with birch and hazel were present during the Brørup interstadial, whereas during the Amersfoort interstadial, correlated with the Danish Rodebæk, there were only birch and pine forests. Quite often the Rodebæk (Amersfoort) interstadial is connected with the Brørup. In western Europe, after the Brørup interstadial, the Odderade interstadial has been distinguished with its stratotype being located in Schleswig-Holstein, Germany. The Brørup and Odderade interstadials in northwestern Europe were warmer and of longer duration than other Weichselian interstadials with boreal forests. They are correlated with the isotope substages 5a and 5c in the deep-sea chronology.

It is worth mentioning that in the Netherlands in Western Europe, the Middle Weichselian interstadials (Moershoofd, Hengelo and Denekamp) were established long ago. Their age succession has been determined on the basis of radiocarbon datings of an order of 30,000 - 50,000 years. According to spore and pollen data they were characterized by a cold-favouring vegetation.

In the City of St. Petersburg, in a section on Grazhdanski Prospect, a still warmer Middle Valdaian interval has been established (Малаховский и др. 1969). According to E. Spiridonova's pollen analysis, this section displays three climatic warmings and three coolings. The warmings are characterised by the culmination of pine, spruce and tree-like birch. However, at the same time, a great quantity of herb (20%) and dwarf birch pollen shows the absence of closed forests. Interpretation is also complicated by the redeposited interglacial pollen of hazel (up to 8%), alder (up to 15 %), broad-leaved and other species of trees, which undoubtedly contribute to the rise in the arboreal tree pollen percentage. During climatic coolings the share of dwarf birch (50-60 %) and herbs (up to 50 %) increased remarkably. Evidently, tundra-like periglacial vegetation was widespread at that time. The section has yielded an radiocarbon date of 40,380 ± 800 (ЛУ-22) years, corresponding to the end of the first warming. Many sections in the East-European Platform correlate with the key section on Grazhdanski Prospect in terms of radiocarbon and palynological data. These Middle Valdaian (Weichselian) sections are not complete; they are mostly located in the vicinity of the marginal zone of the Valdaian glaciation.

In such a complicated situation, it would be rather difficult to correlate these sections with those in western Finland (Vimpeli, Oulainen), which some authors refer to the Eemian interglacial, and others to a Weichselian interval. To our mind, a reliable conclusion has been drawn by Donner (1983, 1984) who considers them as representing an undisturbed Early-

Middle Würmian interstadial. They are correlated with the Jutland interstadial in Sweden, the Peräpohjola interstadial in northern Finland, and the Brørup interstadial in Denmark. Beside spore and pollen data, the conclusion is also supported by the fact that the *in situ* section of Oulainen, with its continental organogenous deposits, is undoubtedly located below the maximum level of the Litorina Sea which means that it lies also below the Eemian Sea level. The Vimpeli section is located at a height of 117-120 m above sea level, *i.e.* 10 m higher than the Eemian marine section at Norinkylä. Therefore, it cannot be Eemian either, but should be attributed to the Oulainen interstadial.

The radiocarbon ages do not contradict this conclusion. The sample of wood from the Vimpeli section yielded the ages ≥43,000 (Hel-1404) and ≥50,300 (Su-925), peat ≥43,000 (Hel-1405) and the fraction of humus from the same sample ≥40,300 (Hel-1406). The sample of gyttja from Oulainen section gave the ages ≥48,000 and 63,000 (+5500, -3200) years (GrN - 7982). Eight TL sand samples above the interstadial deposits in the Oulainen section yielded an average age of 94,000 ± 4000 years and three samples from the lower part - 121,000 ± 5000 years (Donner 1984).

From a stratigraphical point of view, the location of the Vääna-Jõesuu section (Fig. 91) in the Vääna klint bay, 20 km west of Tallinn, has remained unclear. In that section, Weichselian deposits with a total thickness of 71.1 m were studied (Раукас и Лийвранд 1971).

The description of the section:

Layers	Thickness in metres
Yellowish-grey sand, of varying grain size	0.0 - 11.50
Sandy gravel with pebbles	11.50 - 13.45
Glaciolacustrial silty clay with indistinct varved structure	13.45 - 13.80
Grey compact till, rich in crystalline rocks	13.80 - 53.10
Dark-grey silty clay, with organic matter	53.10 - 56.00
Grey compact till, rich in crystalline rocks	56.00 - 71.10
Below follows Cambrian clay.	

Taking into consideration its location in the fore-klint area, only 35 km southwest of the Island of Prangli, and also the similar geological structure and petrographic composition of tills, the Vääna-Jõesuu section is well correlated with the section of Prangli Island. The layer with organic matter lies here higher than on Prangli Island which might indicate its younger age. This is confirmed by pollen evidence.

In the Vääna-Jõesuu section, the quantity of pollen of broad-leaved trees (oak, hornbeam, linden and elm) amounts to 80% in the sum of trees. The amount of hazel and alder reaches 300 %. At the first sight, this concentration is entirely comparable with the quantity of thermophilous tree pollen in interglacial deposits. However, a distinct successional development of vegetation is not observed here. Pollen of the different thermophilous species occurs throughout the intermorainic interval, but the absolute content varies, being less in the lower part of the interval. As an exception serves linden, the concentration of which is lower than that of the other broad-leaved species, and in the lower half of the interval, it is represented only by occasional grains. To solve this problem, E. Liivrand used variograms for tills and intermorainic layers. In the tills redeposited pollen of thermophilous species was discovered in different amounts, but with similar composition. In the tills the pollen is practically entirely redeposited, and in the present case, prevailingly from the underlying Eemian layers. It was demonstrated that the pollen composition of interglacial thermophilous species practically did not change throughout the 70-metre-thick Weichselian section. Interglacial deposits were subjected to the most intensive abrasion during the accumulation of intermorainic clays, in which the preservation of pollen is much worse than in tills. Besides, in the same sections one can observe the inversion of spectra which is indicative of redeposition. During the accumulation of the lower till layer, the Upper Eemian layers with abundant pollen of hornbeam were eroded. Afterwards the lower layers rich in oak pollen were also subject to erosion.

The occurrence of an extremely high amount of redeposited interglacial pollen in intermorainic clays at Vääna-Jõesuu hampered the study of *in situ* pollen of the periglacial vegetation which could have persisted here under the ice-free conditions. According to E. Liivrand, this is shown also by the maximum of herbaceous plant pollen (up to 18 %), accompanied by an increase in the pine and birch pollen, spores of *Sphagnum* and a small content of redeposited pollen of thermophilous species (only 5-12 %).

Diatoms (see Раукас и Лийвранд 1971) are most likely also redeposited from the Eemian deposits, although thermophilous species, typical of this sea, have not been recorded.

G. Nedesheva found in the tills and intermorainic clays foraminifera, sometimes in great amounts. Taking into account their ecological needs, it was concluded that the intermorainic clay accumulated in a shallow sea basin where the water temperature did not exceed +3°C and the salinity was below normal (less than 30‰). But, since similar species have been found in great amounts also in the underlying till, at a depth of 58 m, the foraminifera may well have been redeposited (Раукас и Лийвранд 1971).

The intermorainic interstadial and stadial clays of the Vääna-Jõesuu section may be referred to the Middle Weichselian, although direct evidence in favour of this conclusion have not yet been found. On the basis of their stratigraphical position and spore and pollen characteristics, it is difficult to connect them with the sections of Oulainen and Vimpeli in Finland, or with intermorainic sections in southern Estonia and in the Leningrad District.

The problems concerning the deglaciation of the last ice sheet and the occurrence of Gotiglacial interstadials and stadials will be discussed in the next chapter.

DEGLACIATION HISTORY

General data about the glacial dynamics

Estonia and the Gulf of Finland south of the Salpausselkä ice-marginal formations was freed from the continental ice in Gotiglacial time. The final phase of the Gotiglacial was characterized by a marked differentiation of the radial flow in the thin marginal parts of the ice sheet, resulting in the formation of tension zones, and areal and linear concentrations of drift. Selective erosion and accumulation under the conditions of convergent movement of the ice lobes towards the subglacial uplands led to the formation of the Haanja and Otepää glacial accumulative insular heights (Raukas & Karukäpp 1979).

A dynamic system of fractures developed during the final stage of the deglaciation; along these fractures glaciofluvial deposits were subsequently accumulated. Eskers are practically absent in the areas of the Daniglacial phase of deglaciation, and they are not typical of the early part of the Gotiglacial either. Eskers often occur within the zone of glacial erosion where the average thickness of glacial drift is small, for example, on the Pandivere Upland.

As the thickness of the ice diminished, the movement of the marginal lobes was highly controlled by the underlying bedrock topography (Таваст и Раукас 1982). The lobes reduced in size because the supply from the central parts of the ice sheet decreased. As a result, the length of the margin of the retreating ice sheet diminished towards the end of the Gotiglacial.

Ice-marginal formations

Marginal positions of the ice sheet in the present topography are marked by interrupted chains of end moraines and glaciofluvial formations (Fig. 164D). Most frequently they seem to represent events of temporary stagnations of the ice margin when the glacial regime was close to equilibrium. More often than in the southern part of the Baltic area, extensive proglacial bodies of water were present in front of the receding ice margin. This is reflected in the relatively common occurrence of glaciofluvial deltas (Photo 47) and marginal eskers, in the absence of typical sandurs, and in the wide dis-

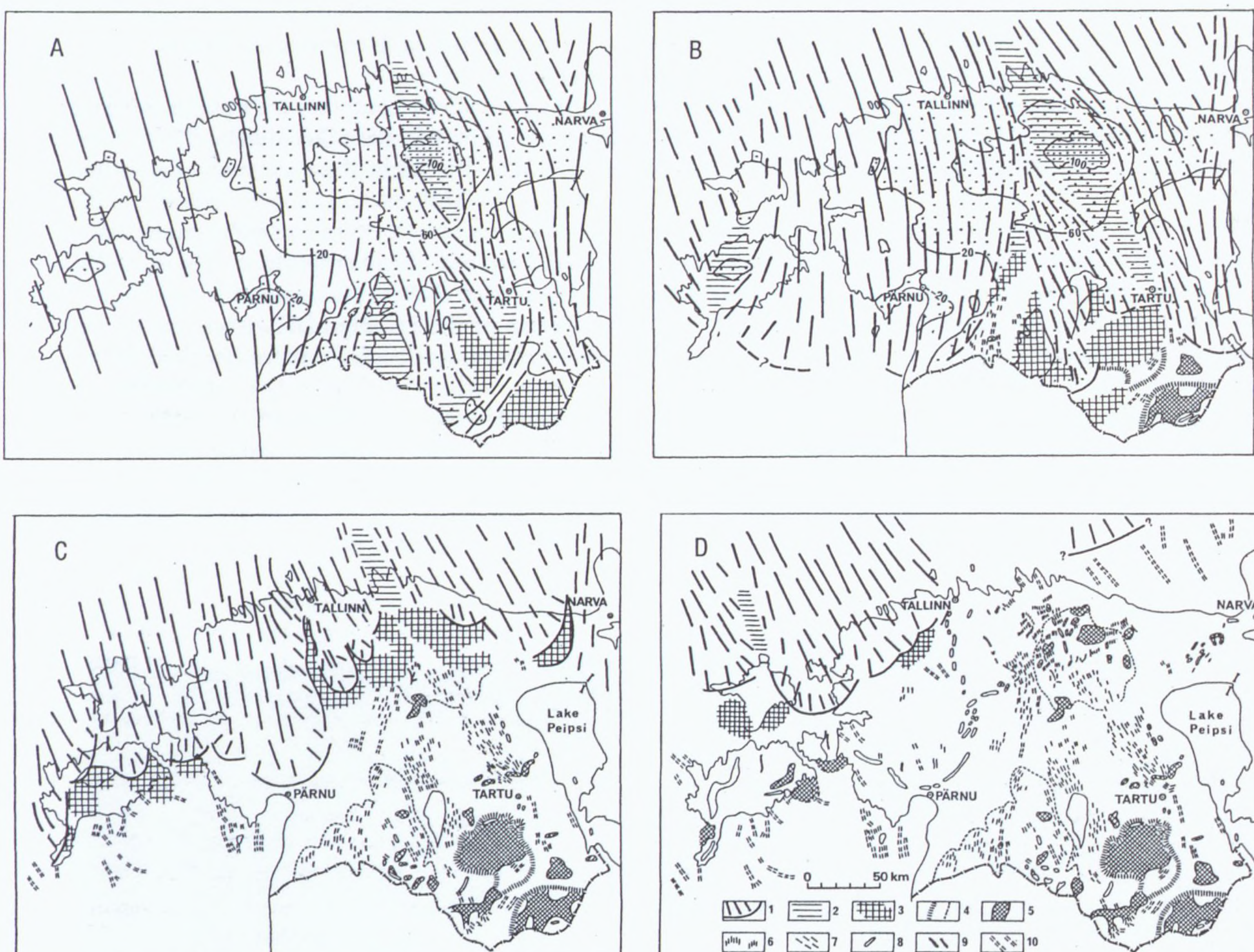


Fig. 164. Glacial dynamics and deglaciation of the Estonian territory against the background of the bedrock topography (dotted surfaces) with indications of its elevation above sea level (lines with numbers) after Raukas & Karukäpp 1979 with complements: A - stagnation of the glacier on the Haanja and Otepää heights; B - marginal formations of glacial lobes in southeastern Estonia; C - Pandivere Stadial of deglaciation; D - Palivere Stadial of deglaciation: 1 - margin of the active glacier and direction of the ice movement; 2 - ice-shed area; 3 - dead ice and formation of hummocky topography; 4 - foot lines of accumulative insular heights (a) and bedrock uplands (b); 5 - hummocky topography; 6 - marginal belts of hummocky topography; 7 - drumlins; 8 - end moraines and marginal eskers; 9 - radial eskers; 10 - orientation of elongated landforms.

tribution of glaciolacustrine sediments. The primary relief was often reshaped by wave action and shore processes of the proglacial lakes.

The concentrated unequal subglacial and marginal accumulation was typical of the early stages of the Gotiglacial in southern and southeastern Estonia (Raukas & Karukäpp 1979), where the Haanja and Otepää glacial accumulative insular heights were formed (Fig. 164A,B). On the basis of the composition of tills belonging to different glaciations or glacial stages, the orientation of elongated clasts in till or linear elements of topography, the direction of the ice flow, and the stages of development of insular heights have been established (Hang & Karukäpp 1979, Raukas & Karukäpp 1979).

During the subglacial stage, the several glaciodynamic structures, including ice-folds, fractures and shear plains overloaded by till and erratics from subglacial surface, were formed. The convergent direction of ice flows caused intensive accumulation around the subglacial elevations as initial centres. Both, the movement direction and rapid accumulation led to stagnation of great portions of thick glacier ice in the central areas of insular heights, which initiated the second, englacial stage of development of the insular heights. This stage marked the beginning of the glacial stagnation and retreat of the active ice in southern and southeastern Estonia.

High relief of specific large landforms of complicated structure in the centre of the heights (Photo 46) is the result of glaciotectonics of extremely great horizontal and vertical stresses in the beginning of the englacial stage of morphogenesis, causing horizontal displacement and considerable upthrusting of sediments. Towards the end of this stage, the area of stagnant ice gradually rose; active accumulation and relief formation took place on the widening transition belt between active and stagnant ice (Photo 48). The glacial deposits still contained great blocks of buried dead ice (Fig. 164B).

The above-mentioned stage was followed by peripheral marginal accumulation (Аболтиньш 1972) and the final, dead ice stage when intensive glaciokarst, slope and erosional processes took place (Карукяпп 1985). Various crevasse fillings, in some places covered with flow till and other solifluctional deposits, were formed.

Photo 47. Typical sandurs are absent among ice-marginal formations in Estonia, but glaciofluvial deltas formed in proglacial lakes, are relatively abundant. Pühamäe Glaciofluvial Delta in northern Estonia. *Photo by A. Raukas.*

Linear ice-marginal formations are typical of the Pandivere and Palivere zones.

In the Pandivere zone, they are represented by push end moraines (Photo 45), glaciofluvial deltas (Photo 47) and marginal eskers. The height of the formations ranges from some metres up to 40-50 metres (Sinimäed at Vaivara, Photo 34) depending on the intensity of accumulation and ablation, and on the abundance of drift in the glacier. The North-Pärnumaa marginal esker chain has the greatest, more or less continuous length of 150 km (Fig.164C). It was eroded and partly levelled by younger bodies of water. Marginal features of the Pandivere Stadial also occur on the northwestern slope of the Pandivere Upland. East of the Pandivere Upland the ice margin followed the line Vaivara - Laagna - Narva.

The Palivere line (Fig. 164D) is clearly marked by mar-

Photo 48. Traces of glacial push in the Tulimäe Hill in the proximal part of the Otepää Heights near Valguta. *Photo by R. Karukäpp.*

ginal eskers in northwestern Estonia and by extensive glaciofluvial deltas near and east of Tallinn (Raukas 1992b). The Upland of Western Saaremaa and Sõrve end moraines are the westernmost continuation of the Palivere line. The eastern part of the line is submerged by the present Gulf of Finland.

A number of different views has been expressed on the course of deglaciation and correlation of the marginal formations in the area. There is no disagreement as to the ice margin positions in western Estonia where continuous chains of marginal formations of various types are present. On the other hand, the reconstructions suggested for the St. Petersburg district and for the areas covered by the waters of the present Gulf of Finland differ significantly (Заррина и Краснов 1965, Серебрянный и Раукас 1966). Analysis of the Gulf of Finland's bottom topography, based on large-scale depth charts, reveals a series of subparallel marginal features (Карукяпп и Васильев 1992) which do not seem to fit in with the conventional schemes. Usually, these features show smaller dimensions than the radial forms of glacial topography, represented by megadrumlins and eskers. However, the available information does not yet allow any reliable reconstruction of the submarine continuation of the Palivere line.

Ice-dammed lakes

Research into the deglaciation history is in many respects based on the analysis of the raised beaches of proglacial lakes. The coastal formations of local ice lakes are usually small, up to 2 m high (Raukas 1992a), and unclear in outline owing to the plain topography, the occurrence of erosion-resistant rocks and deposits in the distribution areas of the lobes, and to the small dimensions, shallowness and brief existence of the lakes, on account of which the action of the waves was inconsiderable (Raukas 1986).

Coastal formations of the Baltic Ice Lake, which formed after the retreat of the continental ice to the north and northwest from the Pandivere Upland, are clearer in outline (Ramsay 1929, Квасов и Раукас 1970). The littoral formations of the different phases of the Baltic Ice Lake are represented by glaciofluvial deltas.

The delta surfaces in northern Estonia were mostly formed at altitudes of 38 - 50m (Männiku) or at about 20 m (Potsepa) and above 70 m (Kemba, Voose). Their vertical distribution shows a more or less clear concentration at certain levels of altitude (Пярна 1960). In most cases, the delta surfaces are even (Photo 47) and gently inclined towards the distal margin, or slightly convex. Proximal slopes of the highest deltas mark the contact with the glacier during the existence of local ice lakes (Kemba, Mustamäe).

Deglaciation of the Pandivere Upland and the area north of the upland is related to two types of glaciofluvial deltas (Карукяпп и Таваст 1985) where the upper (70-80 m) level (Kemba, Valgejõe) has contact slopes with the glaciers and inclination of bedding to the distal southern slope of the delta. The lower (45 - 48 m) level formed as the result of the streams flowing from the Pandivere Upland to the north and transporting glaciofluvial deposits back in the opposite direction — into the lowering ice lakes. The latter are of no significance for palaeoglaciological reconstructions, but indicate the local altitude of the water level before the Palivere Stadial.

Lithostratigraphical evidence

The various tills of different ages from the last glaciation indicate the stadial-oscillatory character of the deglaciation. The distribution of tills and index boulders in them implies that there were at least four stadials during the last glaciation in Estonia. Each of the stadials is characterized by a certain direction of continental ice flow, differing from that of the other stadials and, accordingly, by a certain combination of index boulders (Раукас 1963a). The Haanja, Otepää and Sakala stadials are characterized by the prevailing southeasterly direction of the ice movement, the Pandivere Stadial by the southerly or even southwesterly direction (west of the Pandivere Upland) and the Palivere Stadial again by the southeasterly direction (Fig. 164) with the corresponding composition of index boulders, mainly rapakivi from southwestern Finland, granites and rapakivi from the Åland Archipelago, Baltic red quartz-porphyries and olivine diabases from Satakunta, all of which are practically absent south of the ice-marginal formations of the Palivere Stadial (Raukas 1992b).

The presence of till-covered layers of interstadial character, the distribution of index boulders and a number of other lithostratigraphical observations provide evidence for a significant event of ice front oscillations. For instance, at Männiku in the southwestern part of Tallinn, glaciofluvial varved clays of the Pandivere Stadial are covered by glaciolacustrine varved clays with a thickness of up to 20 m (Раукас и Ряхни 1966, Карукяпп и Мийдель 1972), indicating a glacier readvance.

Biostratigraphical evidence

Interstadial or interphasial layers between the different till beds occur at many sites both in northern and southern Estonia. Palynological data from these layers are available, for example, on the Island of Prangli (Раукас и Ряхни 1966, Раукас 1978). Unfortunately, they contain a lot of redeposited pollen that hampers the correlation of sections and the establishment of palaeogeographical conditions during the interstadial events.

Of greater consequence in dating of deglaciation of the area are the pollen data from till-covered Pleistocene deposits which suggest severe climatic conditions throughout the Late-glacial. Local vegetation just set in to develop and the concentration of redeposited pollen in sediments is high. That is why the Bølling sediments in Estonia do not reveal any clear palynological characteristics (Pirrus & Raukas 1996).

For dating of different steps of the deglaciation, it would be essential to know whether till-covered Bølling deposits occur in Estonia or not. Such deposits may be present in southern Estonia, but hardly in the north. In the section of Haljala (Мянниль и Пиррус 1963), a pollen assemblage suggesting a brief interval of warming, possibly Bølling, has been reported from a sandy interlayer at a depth of 10.5-11.2 m. However, in view of the circumstance that the pollen might have redeposited into the sandy sediment, the interpretation of the pollen data remains uncertain, and the more that there has not been established any other corresponding warm interval elsewhere in Estonia.

Deposits of Older Dryas age occur both in northern and southern Estonia. They are represented by varved clays, silts and sands with horizontal varve-like bedding. Compared with Allerød deposits, their pollen floras are poorly preserved, suggesting redeposition. Upwards the amount of redeposited pollen usually diminishes. The lower boundary of the Older Dryas is undefined in Estonia (Каяк и др. 1976).

Allerød deposits (about 10,800 - 11,800 yr BP) are present even in the northernmost parts of Estonia (Pudisoo, Haljala, *etc.*). This suggests that Estonia was almost totally ice free by the beginning of Allerød, except the northwestern part (Пиррус и Раукас 1969). Allerød deposits differ by the large amount of organic remains they contain and are therefore considered a stratigraphic reference horizon throughout the Baltic States (Kabailiene & Raukas 1986). In the Allerød, the whole Baltic area experienced considerable warming which highly contributed to the rapid ice retreat.

The Upper Dryas deposits (10,000 - 10,800 yr BP) in Estonia are characterized by a high content of herbs (40 - 60% of the total amount of spore and pollen) and dwarf birch. The Younger Dryas cooling gave rise to the development of tundra vegetation once again and undoubtedly promoted activization of the ice cover close to Estonia.

The accumulation of organic sediments in the lakes of southeastern Estonia started only at the end of the Younger Dryas about 10,300 -10,200 yr BP, substituting the accumulation of sand, silt and clay which had been clearly prevailing during the Late-glacial. The oldest organic sediments have been dated at Saviku ca. 10 200±90: TA-328 yr BP (Сарв и Ильвес 1975).

In the light of the pollen evidence, the retreat of the ice margin from the Haanja (Luga) position started during Bølling time and the deglaciation of the Estonian territory was completed during the second half of the Allerød (Пиррус и Раукас 1969). Hence, the deglaciation of the Gulf of Finland took place towards the end of the Allerød and in the beginning of the Younger Dryas.

Chronostratigraphy

In Estonia, deglaciation has been dated with conventional varve chronology, noncorrected radiocarbon chronology and TL methods. Up to now, all the above-mentioned methods are prone to big errors and uncertainities. None of them can, therefore, be assigned universal validity. The majority of ^{14}C dates of intermorainic or submorainic sequences are younger than one would expect on the basis of the conventional methods. A good example is the submorainic sections at Petruse (12,670±200; 12,080±120) and Viitka (10,950±80) in the hilly area of southeastern Estonia which was freed from ice at least 13,000 years ago. This may result from redeposition, caused by glaciokarst processes (Raukas & Karukäpp 1994). At the same time, depending on the "hard water" effect, some organic layers above the uppermost till have given ^{14}C dates of 13,000 - 14,000 yr BP (Пиррус и Раукас 1969). In dating of geological objects with the TL method, the most complicated problem is the establishment of the zero point for the time of the formation of the sediments studied. All the dates obtained are substantially older than would be expected on the basis of traditional deglaciation chronology.

Somewhat better results have been obtained using conventional varve chronology. To establish a local chronology, Rähni (Ряхни 1963a) studied varved clays in northeastern Estonia. The clay series investigated by Markov (Марков 1931a) from the Luga and Neva basins provided a good basis for further correlations. A connection between the varve diagrams from northeastern Estonia and from the Luga Basin in the Leningrad District was established by using the so-called drainage varves. These marker horizons served as the basis for the dating of the ice retreat from the Pandivere Upland (Раукас и др. 1971).

The attempts to create a local chronology for western Estonia on the basis of varved clays have not yet proved successful, although according to Pirrus (Пиррус 1968) and other authors, clays suitable for varve chronological studies are present there.

The attempts to determine the age of the Salpausselkä ridges by means of varved clays have been hampered by the difficulties of correlating the floating Finnish chronology with the Swedish time scale. A connection between the Swedish and Finnish chronologies was proposed by Strömberg (1990) on the basis of step-by-step correlations of varves from east-central Sweden via Åland to southwestern Finland. According to these correlations, and including the recent revisions of the Swedish time scale (Cato 1987), Sauramo's zero year should be dated at 8693 BC (ca 10,650 yr BP).

Markov and Krasnov produced a large number of varve graphs from sites in the Leningrad District (Markov & Krasnov 1930) and in southern Karelia (Марков 1931б). According to these data, there was a constant retreat of the ice margin in the northern Onega area with an average rate of 160 metres a year. Unfortunately, the northwesternmost proximal equicess by Markov in Karelia lies about 100 km distalwards of the southernmost equicess by Sauramo (1923, 1929). By calculations of the rate of retreat, Markov thus arrived at the conclusion that the ice retreat from Petroskoi (Petrozavodsk) took place in the year -3300 in Sauramo's system. According to the conventional dating of Sauramo's zero year, this would correspond to approximately 13,900 yr BP.

Extending Sauramo's chronology to the Leningrad area, Markov (Марков 1931a) estimated that the retreat of the ice from St. Petersburg to Viipuri would thus have taken about 250 - 350 years.

Markov (Марков 1931a) was not able to correlate varved clays from the Luga Basin near Kingissepp (Jamburg) with those from the Neva Basin. The clay sequences in this area are characterized by the occurrence of two drainage varves, one at the level of 79-80 years and the other at the level of 111 years from the base of the varve sequence. They are explained as reflecting two events of ice-lake drainage from the Neva to the Luga Basin, the connection between the basins being regulated by the ice margin leaning against the klint edge near the Koporje Village.

According to Rähni (Ряхни 1963a), the Pandivere ice-marginal position is correlated in time with the drainage of the Neva Ice Lake at Koporje. The first drainage event can be dated at 12,080 and the second at 12,050 yr BP (Raukas *et al.* 1969). When the two large ice-marginal water bodies joined up, synchronous drainage varves were formed in the clay sequences of northeastern Estonia and the Luga Basin (Ряхни 1963a). The temporary closing of the connection between the two bodies of water was caused by an oscillation of the ice margin 12,080 - 12,050 yr BP during the Pandivere event. In 1984, Karukäpp reinvestigated a section of the Luga Basin to check the varve correlation with northeastern Estonia (Ряхни 1963a) by palaeomagnetic methods (Бахмутов и др. 1987). As the result, it was established that the drainage varves in the Luga Basin and northeastern Estonia differ in the composition and, therefore, the direct correlation on the basis of the drainage level is not possible (Бахмутов и др. 1987). The palaeomagnetic data from the studied clay sections did not provide an indisputable basis for a revision of the earlier established age of the Pandivere Stadial (Raukas *et al.* 1969).

Concluding remarks

The territory of Estonia was freed from the continental ice in Gotiglacial time. Against the background of a gradual climatic warming there probably occurred remarkable cooling periods, which caused halts or even new advances (Palivere Stadial) of the degrading ice cover, marked in nature by distinct ice-marginal formations. According to traditional approach, Estonia was finally cleared of the continental ice about 11,000 years ago, but before the glaciers temporarily reinvaded the West-Estonian Archipelago and northwestern Estonia. This Palivere Belt is well marked with end moraines, marginal eskers and glaciofluvial deltas and can be traced from the distribution of index boulders (Raukas 1992b).

It should be pointed out that the revised Swedish varve chronology (Cato 1985, Lundqvist 1986) calls for some readjustment of the dates obtained earlier. It may be suggested that the recession from the Estonian ice marginal zones could have started some half thousand years earlier than hitherto assumed (Donner & Raukas 1989, Карукяпп и др. 1992). However, the data available today do not allow any definite revision of the old deglaciation chronology (Raukas 1986).

EVOLUTION OF THE BALTIC SEA

Estonia has been a maritime nation from time immemorial. The oldest Estonian towns belonged to the venerable Hanseatic League. A great number of Estonians earned their living out at sea and many of them never returned. In consideration of the vital role the sea has played in the life of the Estonians, the investigation and management of coastal areas is a prime priority in Estonia.

Already the ancient inhabitants of the coastal area noticed that in a course of time submarine boulders and small islets dangerous for seafarers crop out from the water, former islands join the land and fishing villages withdraw from the sea-shore. In the 13th century, a big portion of the present old Tallinn was under the sea, with water reaching as far as the town wall, now half a kilometre away from the sea-shore.

The first investigators erroneously associated these phenomena with a gradual fall of the water level in the sea basin (Козакевич 1848). In the second half of the century, the tectonical uplift of the area was a generally accepted truth, however, up to now opinions differ as to the grounds of recent Earth's crust movements.

In the history of the Baltic the following five stages have been distinguished in Estonia:

1) The Baltic Ice Lake in the Allerød and Younger Dryas (Fig. 165 A);

2) The practically fresh-water Yoldia "Sea" at the end of the Younger Dryas and at the beginning of the Pre-Boreal (Fig. 165 B);

3) The Ancylus Lake in the second half of the Pre-Boreal and Boreal (Fig. 165 C);

4) The Litorina Sea in the Atlantic and in the beginning of the Sub-Boreal (Fig.165D);

5) The Limnea Sea in the Sub-Boreal and Sub-Atlantic covering the last 4000 years.

Taking into account similar ecological conditions and general development, the transitional Echeneis and Mastogloia phases are referred to the Ancylus and Litorina stages, respectively. The last episode, called the Mya Sea, is characterized by conditions similar to those at present. Therefore the recognition of an independent Mya Stage in the history of the Baltic in Estonia is not justified.

The evolution of the Baltic Sea is characterised by alternation of transgressions and regressions. For a long time it was believed that transgressions were more or less synchronous all over the Baltic. However, recent studies have shown that due to the different tectonical situation, they culminated at different times in different parts of the Baltic (Кессел и Раукас 1984). Similarily, the velocities and tendencies of the shoreline displacements varied. If in the central and northern Baltic the ancient beach deposits and coastal formations are located above the contemporary sea level, then in the southern Baltic the synchronous formations are up to 40-50 metres below the water level (Fig. 166).

The diatom and mollusc evidence shows that the rise in the salinity and water temperature was not simultaneous either, which complicates stratigraphical correlations of near- and offshore sediments in different parts of the Baltic.

For instance, in Blekinge (southern Sweden) saline conditions established some 8500 years ago, after the waters of the Baltic were reunited with the ocean. Diatoms show that in western Estonia saline influence began about 7800-8000 years ago. A clearly brackish mollusc fauna enclosed in Litorina deposits, appeared there about 7000 years ago (Кессел 1965).

In general, the epicontinental character of the Baltic Sea is reflected in all processes responsible for the coastal morphology and near- and offshore sedimentation. Tides were insignificant. Much more important were seasonal variations and short-term changes in water level induced by strong winds. Longshore drift was predominant in Estonia's northeastern and southwestern coastal waters. Elsewhere it was obstructed by the rugged shoreline and shallowness of water. In those areas material of local origin accumulated. On the basis of the composition and morphology of shingle and gravel grains, the mechanism of the formation of beach deposits can be easily elucidated and the initial rocks defined (Кессел и Раукас 1967). Contrasts in sediment calibre may be correlated with contrasts in wave energy related to the past and present nearshore gradients.

Owing to the moderate crustal uplift of the area, the shoreline remained stable through transgressions. The related coastal formations observable in a vast area are up to 6-7 m thick (Photo 49) and morphologically clear (Раукас 1966). This means that the transgressive shorelines of the Baltic Ice Lake (B_{III}), Ancylus Lake (A_I) and Litorina Sea ($L_I + L_{II}$) can serve as key horizons for palaeogeographical reconstructions.

The treatment of the Baltic stages varies with areas and authors and, therefore, they are not always comparable with each other. The units distinguished on the basis of littoral diatoms are variable because changes in salinity were expressed differently in different areas. Any precise stratigraphical subdivision on the basis of the faunal evidence is difficult (Photo 50) because this evidence mainly comes from shallow-water deposits which seldom form continuous sequences. In actual fact, the subdivisions in different countries are based on mixed criteria and the results are not inspiring. A combination of morpho-, litho-, bio- and chronostratigraphy has lead to confusion (Хюваринен и Раукас 1992). The main stages in the Baltic Sea history are generally known from the beginning of the century, but they have never been properly defined as stratigraphic units.

Ignatius *et al.* (1981) subdivided the Baltic off-shore sediments into three main lithological units, repeatedly observed in cores: (1) glacial varved clays; (2) massive sulphide-rich transitional clays and (3) massive or finely laminated postglacial muds with a relatively high organic content. The transitional clay unit can be divided into two subunits: the lower part is characterized by a strong sulphide colouring, while the uppermost part is formed of grey clay with occasional dark streaks. In all likelihood, the deposition of post-glacial muds started at the beginning of the Litorina Stage when brackish-water conditions were established. In a broad sense, the transitional clay unit correlates with the Ancylus Stage, while the uppermost transition clay (sometimes referred to as "upper Ancylus") correlates with the Mastogloia Stage, separating the Ancylus and Litorina stages (Хюваринен и др. 1992б). The boundary between the transition clays and postglacial muds is well-defined and consistent with the stratigraphical

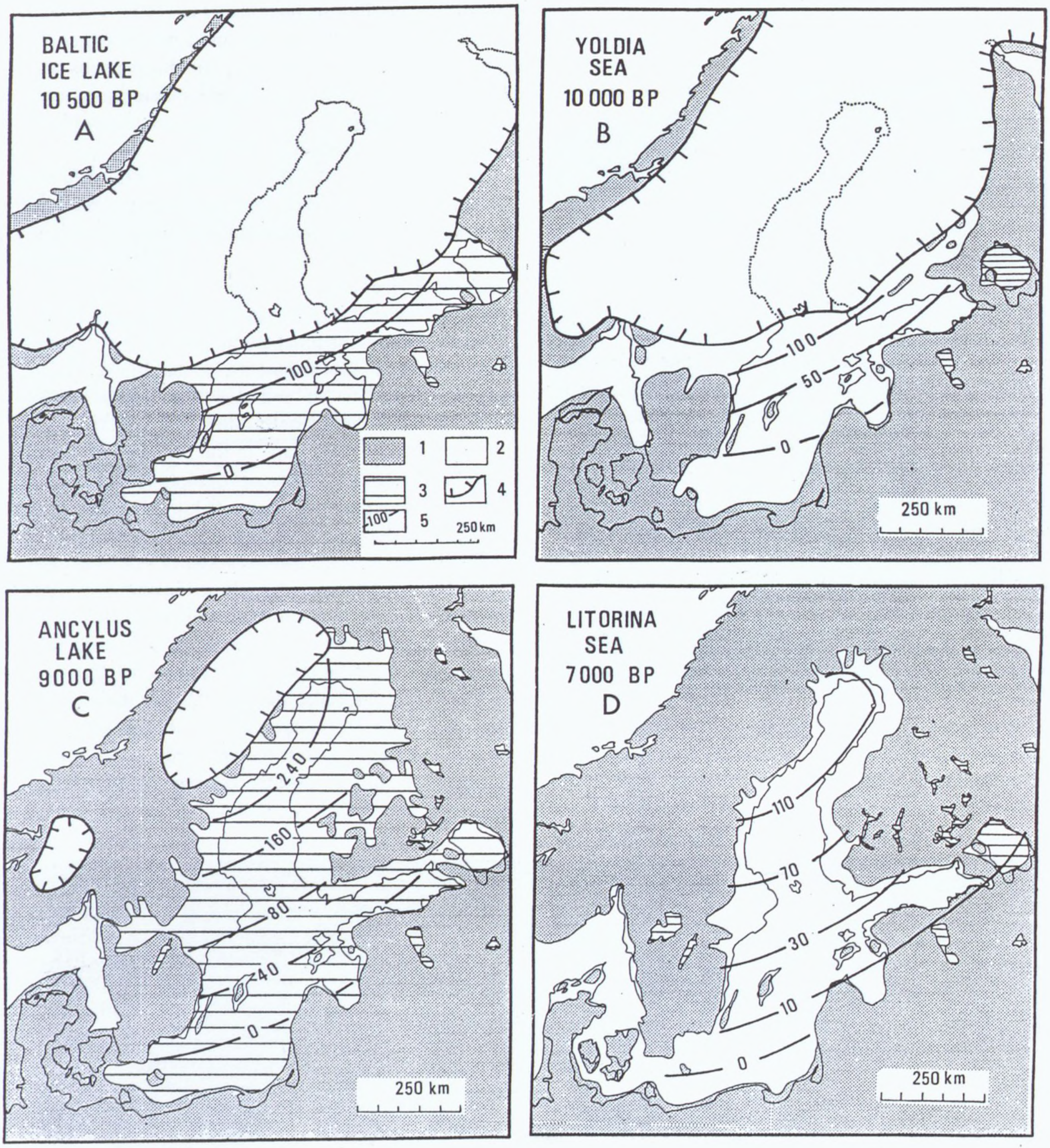


Fig. 165. Palaeogeographical reconstructions of the main stages of the Baltic Sea history after Hyvärinen (1991): 1 - dry land; 2 - sea; 3 - lake; 4 - ice margin; 5 - isolines for shore level altitudes in metres above the present sea level. A - Baltic Ice Lake; B - Yoldia Sea; C - Ancylus Lake; D - Litorina Sea.

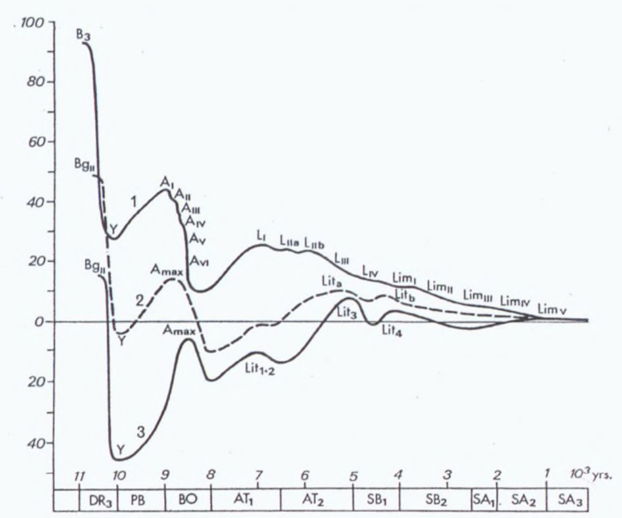


Fig. 166. Water level changes at Kõpu, Island of Hiiumaa (1), in western Latvia (2) and western Lithuania (3) with clear regressions of the Baltic Ice Lake and the Ancylus Lake.

Photo 49. A coarse-grained shingle barrier spit of the Ancylus transgression (A_I) at Iru is characterised by a great thickness of deposits. *Photo by A. Miidel.*

horizon. According to Ignatius *et al.* (1981), it marks the boundary between the Ancylus and Litorina stages. The fact that some of the lithological units are diachronous across the Baltic Sea, while others are clearly synchronous, can cause misunderstanding when used in conjunction with climatic chronozones (Winterhalter 1992).

On the Estonian shelf, Lutt (Лутт 1992) differentiated seven more or less distinct lithological units which could be included to four main sedimentation stages of late- and post-glacial time (Kiipli *et al.* 1993): (1) basal part: formation of glacial and glaciofluvial deposits; (2) stage of Late-glacial vast ice-dammed lakes; (3) stage of post-glacial great lakes; (4) stage of marine sedimentation: formation of mostly stratified pelitic and silty sediments, containing brackish-water fauna and flora. The off- and near-shore deposits are characterized by numerous unconformities and rapid facies changes (Лутт 1992) and often gaps in the sections are prevailing. Therefore, in terms of palaeogeographical conclusions, the classical supraaquatic sections with buried organic sediments and lagoonal deposits often impart much more information.

The Baltic Sea was formed some 12,000 years ago after the readvance of the ice margin from the northern slope of the Pandivere Upland in northern Estonia (Fig. 167), as a result of which the isolated big ice-dammed basins west and east of the elevation joined up (Квасов и Раукас 1970).

The **Baltic Ice Lake**, the first stage in the history of the Baltic, had coastal formations at five different levels of which the Palivere and B_{III} levels (corresponds to the formation of Salpausselkä I ridges in southern Finland) were of transgres-

Photo 50. Coastal formations in Estonia often contain subfossil mollusc shells. *Cerastoderma glaucum* shells in the Limnea Sea beach deposits at Järve, Island of Saaremaa. *Photo by Karl Orviku.*

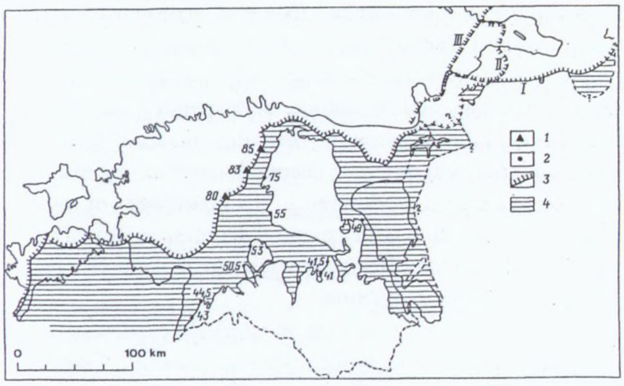


Fig. 167. The initial point of the formation of the Baltic Sea after the retreat of the glacier from the Pandivere Upland. Roman numbers show the location of the ice margin in the St. Petersburg Region according to different scientists: I - K. Markov; II - I. Krasnov and E. Zarrina; III - M. Spiridonov with co-authors. Legend: 1 - elevation of aqueoglacial forms above the present sea level, m; 2 - elevation of coastal forms a.s.l.; 3 - ice margin; 4 - area occupied by ice-dammed lakes. *Compiled by J. Vassiljev on the basis of data published by H. Kessel, E. Lõokene, K. Markov, K. Pärna, A. Raukas and E. Rähni.*

sive character (Fig. 166). It was assumed for a long time on the basis of varve chronology and radiocarbon dating (Raukas *et al.* 1969) that the Palivere Stadial (Fig. 168) occurred about 11,200 years ago. But the results obtained by dating the Salpausselkä ridges (Donner & Raukas 1989) by means of revised Swedish varve chronology (Cato 1985, Lundqvist 1986), in which 430 years is added to the age of the drainage varve in the Swedish time scale, suggest that the Palivere zone was formed about 400-500 years earlier than hitherto supposed (Raukas 1992b).

The retreat of the continental ice cover across the lowland at Mt. Billingen in Central Sweden established a connection between the Baltic Ice Lake and the ocean in the west, causing a rapid lake level drop (Fig. 166). In Estonia, near Pärnu, it was estimated at about 25-30 m (Talviste 1988). This event occurred according to the revised Swedish varve chronology 10,690 varve years BP (Cato 1987), and is considered as the beginning of the Yoldia Sea Stage (Fig. 165B). Svensson (1989) has proposed 10,300 BP as the time for the final drainage of the Baltic Ice Lake. In connection with the abrupt drop of the water level, there should be clear changes in the erosion-accumulation areas and in the grain-size of the bottom deposits but, unfortunately, these phenomena have not yet been precisely studied. Only in a few seismic profiles and boreholes, the boundary between late-glacial and post-glacial clays is distinct.

After the drainage of the Baltic Ice Lake, slightly saline conditions could have existed in the so-called Preboreal Yoldia Sea near the Swedish coast, but not in the open off-shore waters and in the eastern Baltic (Raukas 1991b). In terms of shore displacement, the Yoldia Sea in the eastern Baltic was regressive due to the rapid crustal uplift (Fig. 166). Unfortunately, the lowest point reached by the Yoldia regression cannot be precisely determined. The clearest coastal forms of the Yoldia Sea (Y_{II}) in northern Estonia are probably located several metres higher than the maximum level of the succeeding Ancylus Lake in the same area (Raukas 1995d).

The hydrology and salinity of the Yoldia Sea have been subject to much discussion. A detailed account of the ecological conditions in the area of the Baltic-Atlantic straits in western and central Sweden was provided by Freden (1988). Recent diatom studies in the Kalmarsund area (Håkansson in Svensson 1989) suggest that there was only a brief (100-150 years) influx of salt water during the middle part of the Yoldia Stage, which did not reach the open Baltic or the Gulf of Finland (Хюваринен и др. 1992). According to Gudelis (1979), there were only fresh-water basins of various sizes near the Lithuanian coast in the Pre-Boreal.

Due to the limited connection with the ocean, and abundant meltwater supplied from the nearby ice sheet (Fig. 165B), the salinity of the Yoldia Sea was low, and the brackish-water malacofauna, including rare dwarf forms of *Portlandia (Yoldia) arctica,* were preserved in a rather small western part of the Baltic. The single specimens of *P. arctica,* found in tills and marine deposits in southern Baltic (Kliewe & Janke 1978) and in Latvia, have probably been redeposited from the Eemian sediments. These are known to occur in the coastal zone of Latvia in tills younger than the Eemian marine deposits (Dreimanis 1970).

The Yoldian offshore sediments in the area under consideration are typically barren of diatoms, or may contain a very sparse flora consisting mainly of *Melosira islandica*, which is the dominant species in the richer floras of the Ancylus Stage upwards (Хюваринен и др. 1992).

The presence of Yoldian "brackish-water" diatomaceous taxa in some coastal regions and littoral sediments of the Baltic States and Leningrad District together with fresh-water forms is probably due to redeposition processes. This means that the Yoldian offshore diatom assemblages are mainly fresh (Raukas 1995d).

The crustal uplift, being more rapid than the eustatic sea-level rise, closed the connection between the Yoldia Sea and the ocean, initiating the **Ancylus Lake** Stage (Fig. 165C). Opinions differ as regards the time of this event, however,

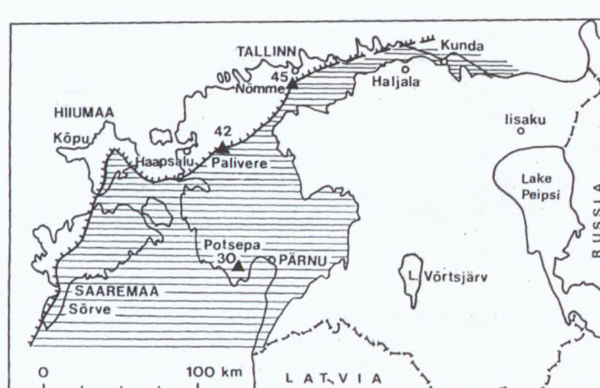


Fig. 168. Palaeogeographic scheme showing the maximum extent of the glaciers of the Palivere Stadial after Kessel and Raukas (Кессел и Раукас 1982) with complements. The hatching denotes the area under the Palivere Ice-Dammed Lake. For legend see Fig. 167.

most frequently 9600 BP is offered. Although, the ice sheet had already strongly receded by that time, its influence on the rapid Ancylus Lake transgression was considerable. The maximum of the transgression was asynchronous in different regions, and in the area of the Gulf of Finland it culminated somewhat 9200-9000 BP (Хайла и Раукас 1992).

The transitional stage of the **Echeneis Sea** between the Yoldia and Ancylus, recognized originally by Thomasson and later by Sauramo and several other authors, is obviously based on erroneous correlations and interpretations of diatom stratigraphy (*e.g.* Svensson 1989, Hyvärinen & Eronen 1979). As there seems to be no valid evidence for this stage, it should be dropped (Кессел и Раукас 1988, Kessel *et al.* 1988).

The Ancylus Stage derived its name from the fresh-water snail *Ancylus fluviatilis* whose remains were found in the beach deposits on the islands of Gotland and Öland and on the coast of Estonia. The diatom flora of the Ancylus period consists of oligohalobous species, the so-called Ancylus forms. *Melosira islandica* ssp. *helvetica* is the dominant species in planktic assemblages. In the off-shore lithostratigraphy, the Ancylus stage is represented mainly by massive, sulphide-stained clays. It is, however, difficult to draw a boundary between the Yoldia and Ancylus off-shore sediments.

According to the biostratigraphic data, the salinity of water in the Ancylus basin near Estonian coast could not have exceeded 3‰, because *Ancylus fluviatilis* is not able to inhabit the water with a salinity higher than that (Kessel & Raukas 1979). The Ancylus transgression was of the order of 10-15 m in the area of the Gulf of Finland, but in the southern parts of the Baltic its amplitude was larger.

Due to a strong erosion of the new outlet channel through the Danish Straits, a rapid regression started over the whole of Ancylus Lake. After a few hundred years, the lowering of the water level slowed down and around 8 500 BP the rising ocean level reached an equiniveau-position with the level of the Ancylus Lake. This event, the establishment of a sound connection between the two bodies of water, may be regarded as the end of the Ancylus Lake Stage.

The end of the Ancylus Lake and the beginning of the Litorina (Mastogloia) Sea is marked in the bottom sediments with a distinct boundary (Ignatius *et al.* 1981), but the reason, responsible for such a great change of sedimentary conditions was until recently unclear. Winterhalter (1992) has explained it with the inflow of saline water, which caused massive flocculation and deposition of suspended mineral particles. Due to the decrease of particulate material in suspension, sunlight could penetrate deeper down resulting in a rapid increase in organic productivity. The abundance of organic matter in conjunction with a marked decrease in fluvial transported mineral matter led, by his opinion, to a drastic change in the character of deposition.

To our mind, this is not the most likely explanation, because according to the diatom flora and mollusc fauna, the salinity changes both in the Ancylus Lake and Litorina Sea were not rapid at all. So, in the Arcona Depression and at Rügen Island, the saline conditions were established at about 8000 yr BP (Kliewe & Janke 1978), on the southwestern coast of Finland at about 7400-7300 yr BP (Eronen 1974), and at the head of the Gulf of Bothnia only at about 7000 yr BP (Eronen 1982).

The regression, following the maximum of the Ancylus transgression (Fig. 166), is often reported as having been rather rapid (Eronen & Haila 1982, Svensson 1989) and, according to Estonian material, had evidences of "catastrophic" outflow similar to the Baltic Ice Lake drainage. Already in the sixties, Kessel and Raukas (Кессел и Раукас 1967) identified a low Ancylus level (A_{VI}), the deposits and relief forms of which have only partly preserved in the contemporary topography and are buried under the transgressive deposits of the Litorina Sea. It was established that the water level dropped about 20 m. According to Tavast (1995), in the Partsi gravel pit on Hiiumaa Island (Fig. 169) the Ancylus mollusc fauna, represented by *Ancylus fluviatilis* (5.5 %), *Lymnaea baltica* (60.7 %), *Bithynia tentaculata* (18.4 %), *Valvata piscinalis* (6.7 %), *Pisidium amnicum* (6.2 %) and *Sphaerum nitidum* (2.5%), is covered with a typical brackish-water Litorina fauna, including *Cerastoderma glaucum* (78.6%), *Macoma baltica* (13.6%), *Hydrobia ulvae* (5.1%) and *Littorina littorea* (1.1%), which provides a clear evidence of considerable paleogeographical changes in the Ancylus/Litorina transitional period. A *Lymnaea baltica* shell sample from Ancylus sediments has yielded an ESR calendar age of 8860 ± 70 yr BP correspond-

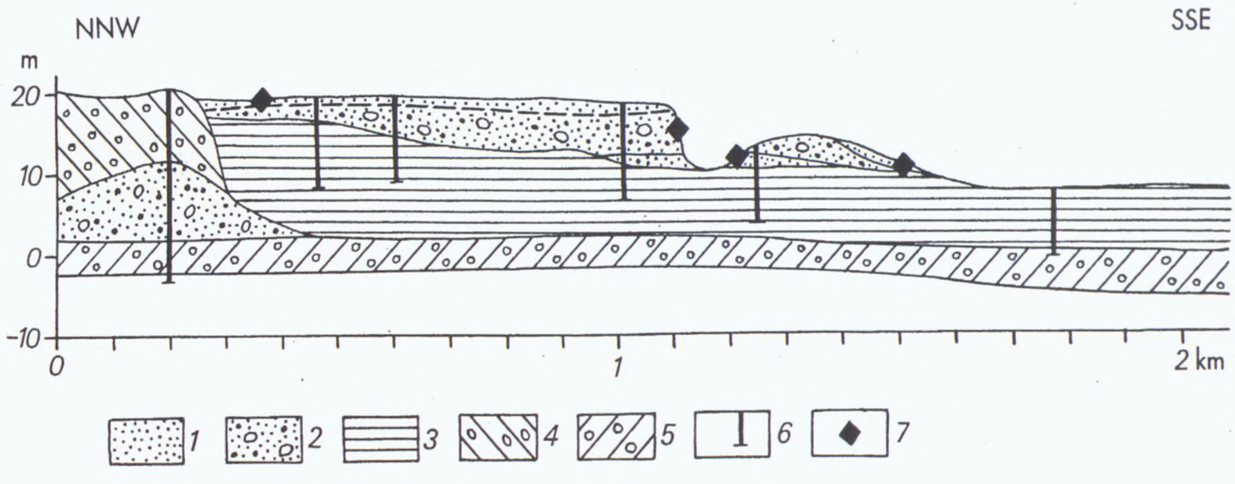


Fig. 169. Geological cross-section of the Partsi gravel pit in the Island of Hiiumaa after Raukas *et al.* 1996: 1 - sand; 2 - gravelly sand; 3 - varved clay; 4 - beige basal till; 5 - grey basal till; 6 - boreholes; 7 - sampling places of mollusc shells.

ing to about 8000 conventional non-corrected radiocarbon yr BP. A *Cerastoderma glaucum* shell sample from the Litorina sediments has given an ESR age of 6310 ± 720 calendar years BP or about 5500 ^{14}C yr BP (Molodkov 1995).

If during the course of 300 years the water level in the Ancylus Lake rose 15-20 m above the ocean level (Eronen & Ristaniemi 1992), then the annual rate must have been 5-6 cm. During the same time span, the ocean level rose at a rate of about 1 cm per year (Pirazzoli 1991).

According to Svensson (1991), a rapid regression of 8-10 m on Gotland followed shortly after 9300 yr BP and the rate of regression during the next 600 years was only around 0.5 m per century. At the same time, on the Island of Hiiumaa, where the transgressive sediments of the Ancylus Lake at Kõpu are fixed at a height of 42-45 m (Kents 1939) and the regressive sediments at Partsi at a height of 10-15 m, the regression amplitude was much higher (including neotectonical uplift 30-35 m). In view of this, the regression rate between 9000 and 8000 yr BP should have been there at least 3.0-3.5 cm annually (Raukas *et al.* 1996).

The **Litorina Sea**, named after the gastropod genus *Littorina*, is the stage of maximum salinity in the history of the Baltic Sea. It is separated from the Ancylus Lake by the **Mastogloia Sea**, a transitional stage of low salinity, often regarded as a sub-stage of the Litorina Sea in a broad sense (Hyvärinen *et al.* 1988). The salinity of the sea water in this basin near the Estonian coast varied between 8 and 15‰ (Kessel & Raukas, 1979).

Shore displacement curves of various authors show a varying number of fluctuations during Mastogloia-Litorina-Limnea time in the course of the last 8500 years. The nature of these fluctuations has been the subject of a great deal of discussion. If these fluctuations were not local, they should have been eustatic in origin and reflect the cyclicity in the global sea-level change. Taking into account a single marine Holocene transgression (Tapes/Litorina) with a broad culmination between about 8000 and 6000 BP on the western coast of Norway (Kaland 1984), a similar sea-level change in the Baltic seems to be most reliable (Hyvärinen 1979). All other short-term relative sea level fluctuations were probably local in origin and they were caused by climatic and tectonic factors (Raukas 1991b).

The transgression culminated at different times depending on the local rate of isostatic uplift (Fig. 166). Along the western sector of the Finnish coast, west of Helsinki, the eustatic rise never exceeded the isostatic uplift. In this area the interval 8000-6000 yr BP is characterized by a very slow regression, indicating that the eustatic and isostatic movements were nearly equal (Хюваринен и др. 1992). After about 6000 to 5000 yr BP, a more or less uniform regression seems to have prevailed also all over the Estonian coast. In Lithuania, the regression of the Litorina Sea in the Early Sub-Boreal about 4300 yr BP was interrupted by a new slight transgression which caused the sea level to rise a few metres (Gudelis 1979). In some papers, a weak Sub-Boreal transgression is reported also in other areas of the eastern Baltic. In Estonia this transgression is not fixed.

The **Limnea Sea** is the final stage in the history of the Baltic. It was first recognized in 1886 by G. Lindström on the basis of the disappearance of the genus *Littorina* and the introduction to the Baltic of the fresh-water mollusc *Lymnaea ovata* (Drap.) f. *balthica* Nilss. The lower boundary is gradual. It is proposed to define this boundary at 4000 years BP (Раукас и др. 1992), in agreement with the usage common in Sweden (*e.g.* Freden 1979). According to Kessel (Кессел 1958, 1961, 1965), *Lymnaea ovata* (*Radix peregra* f. *baltica*) immigrated to the coastal waters of Estonia about 4000 yr BP and already at the beginning of the Limnea Stage, 2-37% of all mollusc shells in the North-Western Archipelago of Estonia belonged to this species; *Lymnaea stagnalis* appeared about 2500 and the typical fresh-water species *Bithynia tentaculata* about 1700 years ago (Кессел 1965). The index species of the Litorina Sea - *Littorina littorea* and *L. Saxatilis*, passed away from the Estonian coastal waters only about 1500 years ago when the salinity dropped below 7-8‰, critical for those species (Кессел 1958). Lagoon and shallow-water deposits of Limnea age are usually characterized by weakly brackish diatom assemblages similar to recent assemblages. The Limnea shorelines, recognized in Estonia, are all regressive. The salinity was about 10‰ in the western part and about 5‰ in the eastern part of the Gulf of Finland. Around 2500 years ago, the salinities were still 2 to 3‰ higher than today (Кессел 1965).

In spite of more than a hundred-year-long investigations of the Baltic Sea sediments and shorelines in Estonia, some conclusions are still rather hyphothetical and contradictory. So, for the establishment of real water-level fluctuations and shoreline displacements for selected time slices we need much better knowledge of the actual rates of tectonic and glacio-isostatic movements of the Earth's crust which undoubtedly differed in every concrete area in time and space. As to the chronologically fixed events, truthful information is needed concerning the possible correlation between ^{14}C, ESR, OSL, varvo-chronological and tree-ring data. Besides, agreements in using ^{14}C dates (corrected and/or noncorrected against tree-ring data) should also be reached.

To reconstruct palaeoenvironment, much greater knowledge of the climatic conditions (circulation of currents, storm frequencies, wave activities, *etc.*) in the past is needed. On the contemporary seashores of Estonia, differences between low and high water level are about 1-1.5 metres, however, in the event of exceptionally fierce storms they may reach 2-3 metres. It means that contemporary beach ridges may be a couple of metres higher above the normal water level and some distance inland from the shore. Due to slope processes, wind and water erosion, it is sometimes difficult to establish the actual heights of beach ridges.

For the further understanding of the suspectibility of the Baltic Sea ecosystem to global changes and elaboration of strategies for the integrated management of the sea and coastal zone, the reconstruction of the basin development since deglaciation will be of the first-rate importance. The study of sea level displacements is an important prerequisite to successful modelling of palaeoenvironmental processes and to better understanding of former ecosystems. Owing to the moderate uplift of the Earth's crust, which differed remarkably with the areas, the coastal relief forms in Estonia are diverser and more pronounced than in neighbouring areas. In view of the above, the Estonian coasts are of great scientific significance since, on the one hand, they allow a glimpse into the late- and postglacial geological history of the Baltic Sea and, on the other hand, make an excellent laboratory where geomorphological and environmental processes and the effects of the changes taking place on land and in near-shore areas can be studied in particular detail.

During the last decades, Estonian beaches have repeatedly suffered heavy storm damage (Photo 51). Therefore, the management of the coastal zone should be carefully planned taking into consideration the possible sea level rise to be expected in the future due to the "greenhouse effect".

Photo 51. During the last decades, Estonian beaches have repeatedly suffered heavy storm damage. Beach at Rocca al Mare (west of Tallinn) in 1993. *Photo by J. Kask.*

HOLOCENE TERRESTRIAL PROCESSES

River activity

The rather young drainage system in Estonia has formed and developed during the last 13,000 years. The formation of river valleys was closely related with the deglaciation of the territory, the evolution of the Baltic Sea and unequal glacioisostatic uplift. Besides, other factors affecting the development of river valleys included the bedrock topography, lithology of Palaeozoic rocks, geomorphology and lithology of Quaternary deposits. The location and orientation of rivers was also controlled by tectonic joints (Tammekann 1926, Teichert 1927a, Мийдел 1966а,б; 1971, 1982). It is particularly obvious with the rivers cut in the bedrock.

On the ground of geomorphology and geology, Orviku (Орвику 1960в) divided the Estonian river valleys into three groups, distinguishing between the rivers in the catchment area of: (1) the Gulf of Finland, (2) the Väinameri and the Gulf of Riga, and (3) lakes Peipsi and Võrtsjärv.

The rivers of the above-mentioned groups differ clearly in the shape of the longitudinal profile (Fig.170). The North-Estonian rivers have convex longitudinal profiles. In the lower courses, the stream gradient is great, often more than 2-3, occasionally even 5-8 m/km (Мийдел 1963). In many cases, 30-50% of the total fall of rivers takes place within short sections (Photo 52) forming only 3-25% of the total river length. The longitudinal profiles of the West-Estonian rivers flowing into the Gulf of Riga and Väinameri are straight, with gradients remaining more or less constant at the whole length of the rivers (Fig.170). Concave longitudinal profiles are typical of the South-Estonian rivers flowing into the southern part of L. Peipsi and L. Võrtsjärv. For instance, in the longitudinal profile of the Piusa River, the lower course forms about 46% of the river's total length, but only 13% of the total fall takes place within this section (Liblik 1966). Numerous studies (Орвику 1960в, Мийдел 1963, 1966б, Hang & Miidel 1987) have shown that the longitudinal profiles of Estonian rivers reflect main elements of the bedrock topography, glacial landforms and deposits, furnishing thus an excellent example of a young drainage system with the unevennesses of the profiles not yet levelled.

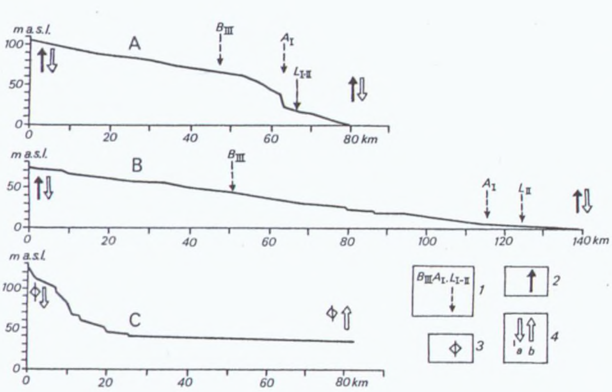


Fig. 170. Longitudinal profiles of Estonian rivers (Орвику 1960в): A - Valgejõgi (N Estonia): B - Pärnu (W Estonia); C - Väike Emajõgi (S Estonia). 1 - transgression limits of the Baltic Ice Lake (B_{III}), Ancylus Lake (A_I), Litorina Sea (L_{I-II}); 2 - rapid uplift; 3 - slow uplift; 4 - bottom erosion (a), considerable fluvial accumulation (b).

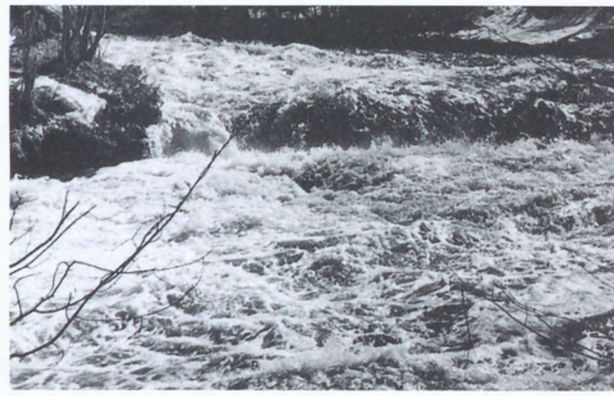


Photo 52. In many cases a considerable amount of the total fall of North-Estonian rivers takes place within short sections. Spilling over the klint edge, the rivers often form waterfalls and rapids. Nõmmeveski. *Photo by A. Miidel.*

The thickness of alluvial deposits in the North- and South-Estonian river valleys differs notably. In northern Estonia, the thickness of fluvial deposits increases from the upper course (1-3 m) towards the middle course (6-7 m), from there onward it decreases steadily. In the lower course, the deposits are only 1-2, occasionally 4-5 m thick (Мийдел и Раукас 1965, Мийдел 1966б). The valleys in southern Estonia have a rather thick alluvial cover in the lower courses - up to 10-15 m together with peat (Fig. 171; Каяк 1959, Мийдел 1966б).

It is interesting to follow the changes in the relation of overbank and channel facies along the rivers. However, it is sometimes very difficult to distinguish between these facies, particularly in places where the river valleys have been cut in fine-grained glaciolacustrine deposits. In these cases, river deposits and facies are lithologically very similar to the source material. Nevertheless, the general tendencies have been determined. In the valleys of northern Estonia, the proportion of channel deposits increases from the middle courses downstream - from 40-50 to 60-70% of the total thickness. By means of detailed grain-size analysis it has sometimes been possible to distinguish between point bar and channel lag subfacies (Hang & Niedzialkowska 1994). Oxbow deposits are of limited distribution. In the lower courses, where incision dominated in the Holocene, palaeochannels are only partly filled with sediments. They are often empty and dry. In the middle reaches, oxbow deposits are of wider distribution. In the grain-size they are very similar to overbank deposits which makes their identification rather problematic.

In the South-Estonian valleys, the role of different facies is just reversed (Fig. 171). The share of channel facies diminishes downstream and in the lower courses it forms less than 20-30% of the whole fluvial sequence. The channel facies is at its thickest (4-5m or 50-60%) in the middle reaches of the rivers (Hang 1995). The overbank facies is characterised by a high content of fen peat. In the lower reaches, its thickness ranges from 5 to 10 m (Каяк 1959, Мийдел 1966б, Мийдел и Тавaст 1981, Miidel *et al.* 1995).

The grain-size and mineral composition of alluvial deposits is immediately controlled by the stream gradient, the bedrock and Quaternary deposits. The main influx of sedi-

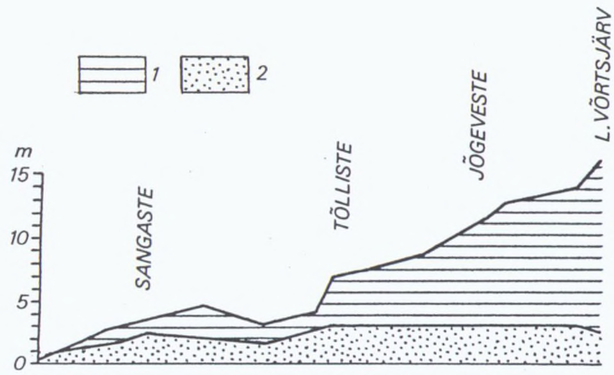


Fig. 171. Overbank (1) and channel (2) deposits in the Väike Emajõgi Valley (S Estonia). Compiled after Kajak (Каяк 1959).

ments into the channel is by creep and sliding from the undercut valley fill of various genesis, and from older river terraces. Thus, the grain-size and mineralogy of river deposits are strongly influenced by local geology. It is reflected most distinctly in the character of channel deposits. In northern Estonia, channel deposits often contain boulders, cobbles and shingles of the Palaeozoic carbonate rocks. The channel deposits of specific character are formed in the valleys, cut in till. As the finer fractions are washed out, coarse material (gravel, boulders, cobbles) remains in the channel. The further downcutting is hampered by the accumulated layer of very coarse material. Rapids with boulders of crystalline rocks typically occur in those channels. In the middle courses of rivers, where mainly fine-grained glaciolacustrine sediments are eroded, the channel deposits contain an abundance of fine-grained material with clay and silt fractions (sometimes more than 60%). In the overbank deposits, fine sand and coarse silt prevail. The mineral composition of alluvial deposits also reveals a great influence of local geology. In both channel and overbank deposits among light minerals (fraction 0.1 - 0.25 mm) quartz and feldspars dominate. The content of carbonate minerals is usually less than 5%. In the lower courses, just downstream the North-Estonian Klint, glauconite is present in notable amount. Among the heavy minerals, amphiboles, garnet, magnetite and ilmenite are prevailing. This kind of mineral association is typical of the deposits occurring in the area of the last glaciation. It should also be mentioned that the overbank and channel facies do not significantly differ in mineral composition (Мийдел и Раукас 1965).

In southern Estonia, where Devonian terrigenous rocks crop out, the alluvial deposits consist mainly of sands with a variable grain-size. For example, the overbank sediments of the Piusa River are dominated by fine (0.1-0.25 mm, 55%) and medium sand (0.25-0.5 mm), while the channel facies consists prevailingly of medium (55%) and coarse (0.5-1.0 mm, 25%) sand.

According to Aivo Lepland (unpublished data), guartz (over 90%) and feldspars are the prevailing light minerals (fraction 0.1-0.25 mm) in both channel and overbank facies. Among heavy minerals, amphiboles, garnet and magnetite dominate. Compared to northern Estonia, the content of zircon, tourmaline, staurolite and leucoxene in the South-Estonian rivers is clearly higher. This is explained with the influence of Devonian terrigenous rocks in which these minerals are common.

The morphology and development of the river valleys in Estonia has been discussed in a number of studies (Tammekann 1926, Künnapuu 1957, Arold 1960,1971, Linkrus 1963, Hang *et al.* 1964, Liblik 1966, Eberhards & Miidel 1984, Hang & Miidel 1987, Miidel & Raukas 1990,1991, Hang 1995, Raukas & Miidel 1995, Каяк 1959, Орвику 1960в, Мийдел 1966б, 1967, Мийдел и Таваст 1981 a.o.). The results of these studies have shown that the morphology and evolution of river valleys in northern and southern Estonia was substantially different. In the north, the valleys are better developed in the lower courses where deep, V-shaped valleys, short canyons and flat-floored valleys prevail. Their depth ranges from 15 to 35 m. The valleys are at their deepest a little downstream of the North-Estonian Klint. In the upper and middle courses, either V-shaped or only channel valleys have developed in the zones of marginal glacial formations, or flat-floored valleys occur in the plains of various genesis. Usually the depth of these valleys does not exceed 10 m. Morphological valley is missing in the peatlands, which are common in the upper and middle courses of the North-Estonian rivers. The development of these valleys has been significantly affected by the Baltic Klint - a steep Palaeozoic escarpment which rises up to 56 m above sea level. The Baltic Klint emerged from the sea step-by-step; earlier in the east after the drainage of the Baltic Ice Lake and later in the west, during the regression of the Ancylus Lake. The klint with its cap from resistant carbonate rocks turned into a permanent base-level for rivers. Together with the numerous related waterfalls (Photo 19) the klint prevented the erosion from penetrating further inland (Мийдел 1967). This explains why under the conditions of the lowering sea-level and continuous land uplift numerous terraces were formed only in the fore-klint reaches of the North-Estonian valleys.

The number of the terraces in the lower courses (Photo 53) ranges from 3 (Pühajõgi Valley, Tammekann 1926) to 11 (Valgejõgi Valley, Linkrus 1963). They are represented by small segments, measuring 30 - 300 m in length and 10 -100 m in width. The surface of the terraces is more or less inclined downstream and towards the channel. They usually have a thin alluvial cover of coarse sand and gravel, the overbank facies is lacking. The altitude and the time of formation of the terraces have been correlated with the numerous shorelines of the Baltic Sea (Tammekann 1926, Künnapuu 1957, Arold 1960, Linkrus 1963, Hang *et al.* 1964, Мийдел 1967 a.o.). It is worth of mentioning that not a single Baltic Sea stage or level has a corresponding terrace in the North-Estonian valleys. Consequently these, mainly erosional (rock-cut and fill-cut), terraces belong to the type of unpaired terraces. The spectrums of North-Estonian river terraces are opened towards the mouth and, having a fan-like shape, point to the continuous rejuvenation caused by the land uplift and lowering base-level (Fig.172, Мийдел 1967, Hang & Miidel 1987).

The land uplift had a different effect on the development of rivers in northern and western Estonia. The isobases of the transgressive shoreline of the Baltic Ice Lake indicate a remarkable difference in the uplift (nearly 20 m) between the upper and lower courses of the rivers flowing to the northwest. For the rivers, flowing to the south-west or even west, it is less than 8 m. In the first case, the distribution of the uplift along rivers caused a relatively strong tilting of the surface to the south-east. As the monoclinally bedded bedrock had a slight southward inclination and the topography was flat, shal-

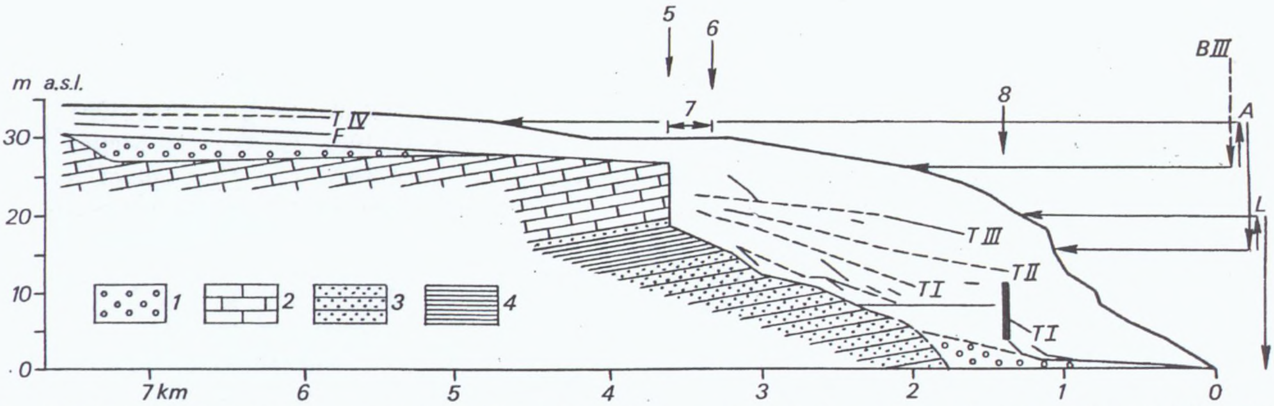


Fig. 172. Geology and geomorphology of the Jägala River: 1 - till; 2 - carbonate rocks; 3 - sandstones; 4 - argillite; 5 - waterfall; 6 - the mouth of the Jõelähtme River; 7- canyon; 8 - the dam of the hydropower station. F - floodplain, T_I..T_{IV} - terraces, arrows indicate the phases of the Baltic Sea and their horizontal extension (B_{III}- the Baltic Ice Lake, phase lll, Y - Yoldia Sea, A - Ancylus Lake, L - Litorina Sea). *After Eberhards and Miidel (1984) with supplements.*

low valleys with wet boggy floodplains and without a single terrace on the slopes, developed in the middle reaches (Мийдел 1963, 1966б). It is quite possible that the development of rivers was characterized by a modest incision or even by a slight aggradation and shifting of the channel zone back and forth, more or less at the same level (Raukas & Miidel 1995). With the rivers flowing into the Gulf of Riga and Väinameri, the uplift along channels was relatively moderate and more or less equal. The river activity in western Estonia depended mainly on eustatic sea-level changes and local geology. In both cases, the regression of the Baltic Ice Lake brought about considerable elongation of the rivers (Орвику 1969).

Two stages of development have been distinguished in the evolution of the North-Estonian river valleys (Hang & Miidel 1987). The first stage started with the formation of glaciofluvial deltas in the Baltic klint bays east of Tallinn and lasted until the rising of the klint out from the sea. This moment marks the outset of the second stage which is still in progress. In the timescale, the boundary of the stages varies with the regions. In the western part, as far as the Jägala River, it coincides with the regression of the Ancylus Lake and the following drop in the Baltic Sea level. In the eastern part, the second stage commenced earlier, after the retreat of the Baltic Ice Lake. During the first stage, fluvial activity was probably modest, while the second stage was a period of intense downcutting, leading to the formation of deep valleys in the lower courses of the rivers.

In southern Estonia, V-shaped valleys are prevailing in the upper and middle courses where they have cut into the complicated topography of the Haanja and Otepää uplands. Due to the changing glacial topography, V-shaped valley sections alternate with flat-floored ones. The depth of the valleys varies between 15-30 m and the width ranges from 200 to 500 m. Descending from the uplands to the surrounding plains, valleys reach their maximum depth (30-45 m). Towards the river mouth, the depth of the valleys gradually decreases. However, at the mouths, valley bottoms lie 6-15 m lower than the present water-level of lakes Peipsi and Võrtsjärv (Каяк 1959, Liblik 1966, Мийдел 1966б, Мийдел и Тавааст 1981). As an average, these valleys are 0.7-1 km, occasionally up to 2 km wide. The wide floodplain is subject to paludification in the middle and lower courses where the thickness of the resultant fen peat reaches 5-10 m (Орвику 1960в, Liblik 1966, Мийдел 1966б, Miidel *et al.* 1995).

The number of terraces left on the valley slopes by river activity (Hang *et al.* 1964, Liblik 1966, Hang 1995, Hang *et al.* 1995) ranges from six (Võhandu Valley) to 16 (Ahja Valley). The terrace levels form some groups, 2-4 levels in each (Fig. 173). In the Piusa Valley, the highest terraces are 500-600 m wide (Liblik 1966) and up to 300 m long. However, their relative height is only 2-2.5 or even less metres. The terraces were cut mostly into glaciolacustrine and glaciofluvial deposits, but in some cases also into till and bedrock. In the rock-cut terraces, the alluvial cover is thin (1-1.5 m) consisting of coarse sand and gravel with pebbles. It is sometimes difficult to distinguish alluvial deposits from glaciolacustrine or glaciofluvial ones. Nevertheless, the terraces are undoubtedly erosional.

The development of rivers in Estonia started from the south according how the territory was freed from the ice cover. The large Võru Valley came into being some 12,600 yr BP, when the glacier of the Otepää Stadial came to a halt along the line Kulje - Talabsi islands - Elizarovo, and the Pihkva Ice Lake I was formed (Раукас и Ряхни 1969). The meltwaters flowed westwards towards the Gauja Basin. Many geologists have recognized the Võru Valley as a marginal meltwater spillway (Hausen 1913b, Hang *et al.* 1964, Раукас и Ряхни 1969, Раукас и др. 1971, Квасов 1975, Miidel & Raukas 1991 a.o.).

Photo 53. The sea-level lowering and continuous land uplift have promoted formation of terraces in the lower courses of the North-Estonian rivers. The Selja River. *Photo by A. Miidel.*

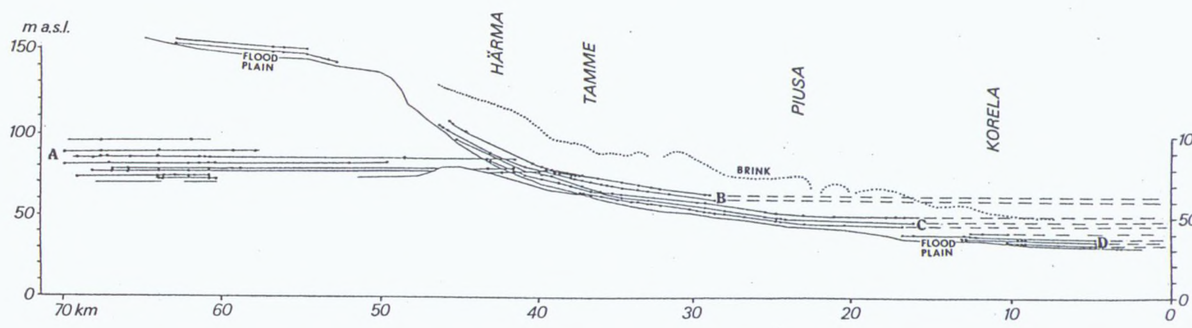


Fig. 173. Terraces (B, C, D) in the Piusa River valley. Glaciolacustrine terraces of the budget A are located in the ancient Võru-Petseri Valley (after Liblik 1966).

However, after Lepland (1991) and Hang (1995), fine-grained glaciolacustrine deposits with horizontal bedding in the consistence of more or less horizontal terraces in the Võru Valley indicate glaciolacustrine origin of those terraces. Accordingly, it was supposed that there was no meltwater flow from the east to the west along the Võru Valley, and this depression was occupied by ice-dammed lakes in which the water-level gradually sank. The formation of river terraces began in the Piusa, Võhandu and Ahja valleys with the lowering of lake-level down to the altitude of 75-70 m. The further retreat of the glacier and lowering of the water-level in ice-dammed lakes led to the incision in the valleys. When the glacier margin stopped on the line Kaiu - Gdov, the outflow from the Peipsi Ice Lake I took place via the Väike-Emajõgi Valley to the south (Раукас и др. 1971). Further retreat of the glacier to the north led to the formation of the Emajõgi Valley and later the Viljandi Valley (Раукас и др. 1971), through which meltwaters flowed from the Peipsi ice-dammed lake to the west.

After Miidel and Raukas (1991), the terrace formation processes in southern Estonia completed in the Bølling. If the formation of the youngest (lowermost) terraces are correlated with the Männikvälja - Iisaku marginal formations (Fig. 174), belonging to the Pandivere Stadial, the formation of terraces should have ended about 12,500 yr BP (Hang 1995, Hang *et al.* 1995). The varve chronology (Раукас и др. 1971) shows that the terrace formation process developed extremely quickly - during the course of 200-300 years (Hang 1995).

The low position of the valley bottoms in the southern part of lakes Peipsi and Võrtsjärv indicate the continuous downcutting in the river valleys after the terrace formation due to the lowering base-level. As a result, in the south the lake depressions dried up and the rivers cut down even as deep as 13-16 m below the present lake-level (Каяк 1959, Орвику 1960в, 1969, Hang *et al.* 1964, Мийдел 1966б, Мийдел и Таваст 1981 a.o.). Probably it took place during the Younger Dryas or later (Орвику 1960в, Раукас и Ряхни 1969, Hang *et al.* 1995). As a result of the uneven land uplift, the base-level rose. Alluvial deposits started to accumulate and re-deepened valleys were subject to paludification. In the Optjok River, it commenced during the Early Boreal (Miidel *et al.* 1995), in the mouth of the Võhandu River during the Boreal (Пиррус и Тасса 1981) and in the mouth of the Emajõgi River in the Late Atlantic (Thomson 1939b, Сарв и Ильвес 1975). The process is still in progress. It has been supposed that the floodplains in the South-Estonian valleys were formed in the Holocene. Dating of oxbow sediments from the Piusa and Võhandu valleys indicates that the meandering of rivers and the formation of accumulative floodplains with organic deposits started in the Pre-Boreal or Early Boreal together with the rise of the base-level (Fig. 175).

As stressed above, the river valleys in northern and southern Estonia have developed in different ways (see Fig. 176).

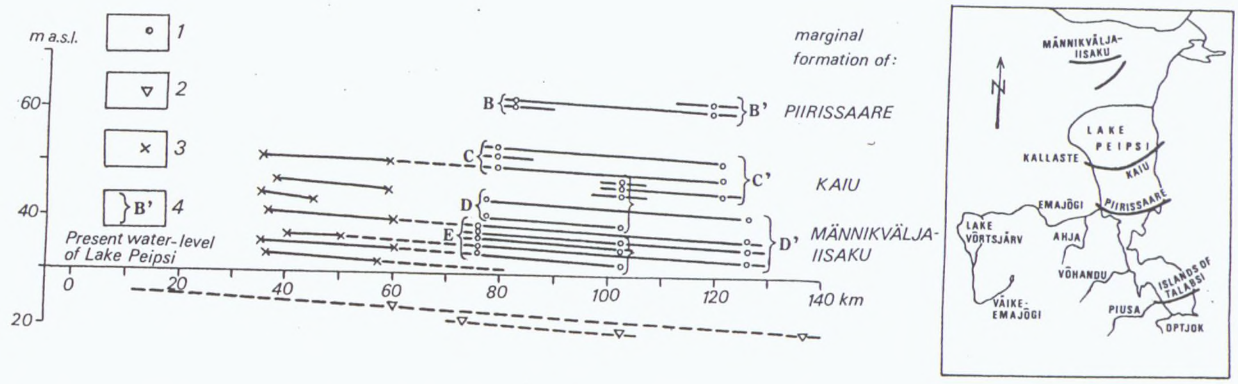


Fig. 174. Correlation of the altitudes of the river terraces and Late-Weichselian shorelines in the Peipsi Depression (after Hang 1995): 1 - river terraces; 2 - valley bottoms; 3 - shorelines (after Liblik 1966); 4 - group of terraces with similar ages and their correlation with the ice-marginal formations (B - E - terraces of the Ahja River, B'-D' - terraces of the Piusa River). The location of ice-marginal belts and valleys is shown on the scheme.

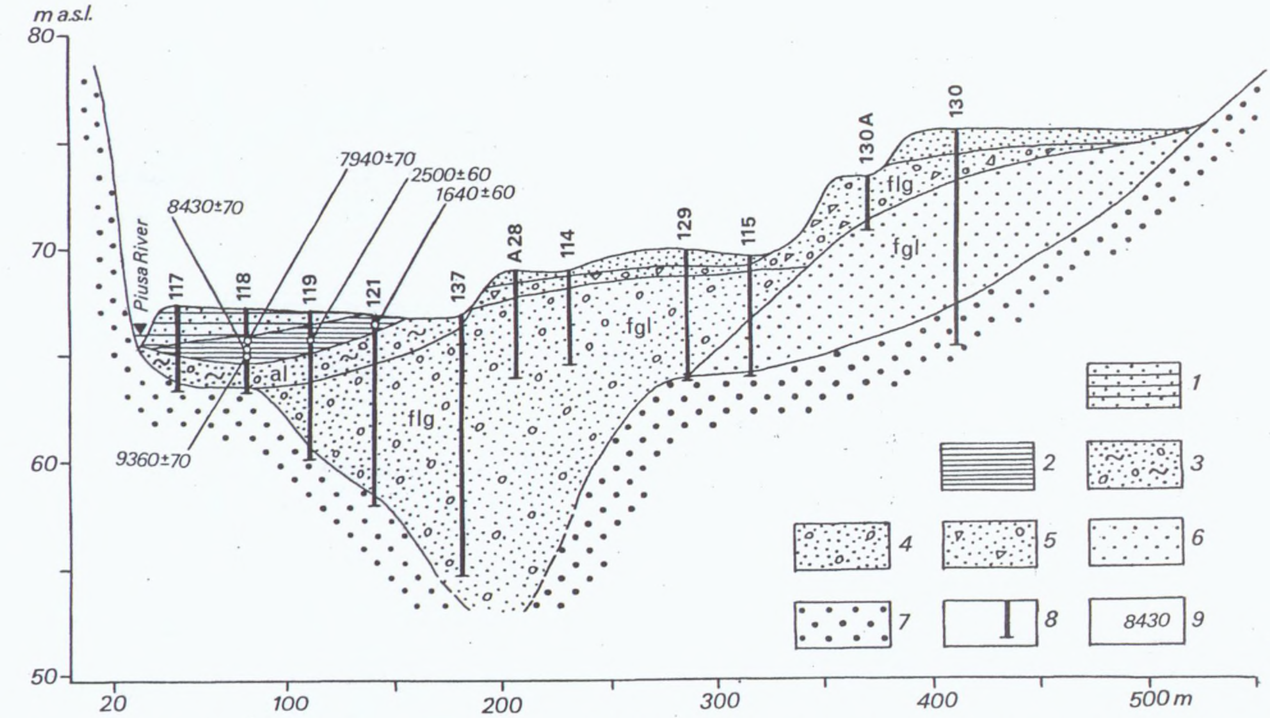


Fig. 175. Cross-section of the Piusa River, near Härma (after Hang 1995): 1 - overbank deposits; 2 - oxbow deposits; 3 - channel deposits; 4 - glaciofluvial sand and gravel; 5 - glaciofluvial gravel with pebbles; 6 - glaciolacustrine sand; 7 - terrigenous Devonian rocks; 8 - borehole; 9 - ^{14}C datings (yr BP).

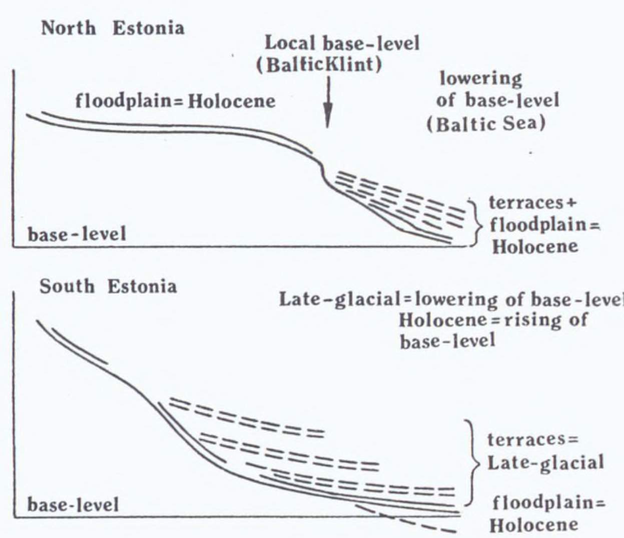


Fig. 176. Principal scheme of age relations of terraces and floodplain with base-level changes in Estonian rivers (after Eberhards and Miidel 1984, with supplements).

Lake Peipsi

Lake Peipsi (3555 sq km) on the border with Russia, ranks fourth in size among the European lakes. Its average depth is 8 m and maximum depth 15.3 m. The lake consists of three main parts: the northern section - Peipsi proper is connected with the southernmost part L. Pihkva through the narrow, strait-like L. Lämmijärv.

Lake Peipsi lies in a vast ice-lobe depression which in the Late Weichselian was occupied by ice-dammed lakes. The contours and water-level of these bodies of water were controlled by the northward retreat of the glacier and opening of new outlets.

During the Otepää Stadial some 12,600 years ago (Раукас и др. 1971), when the ice margin retreated to the line Kulje - Lisje - Talabsi islands - Elizarovo, a big ice-dammed lake was formed in the southernmost part of the lake depression (Раукас и Ряхни 1969, Fig. 177a). This was Pihkva Ice Lake I with the water-levels 95, 85 and 75 m a.s.l. From this lake, meltwaters flowed to the southwest along the Võru Valley (known also as the Piusa-Võru-Hargla Valley) into the proglacial lakes of the Gauja Basin (Fig. 177a). In the Võru Valley, group A terraces were formed at altitudes between 71.5 and 95 m (Liblik 1966). According to Kvasov (1979, Квасов 1975), this lake was only the western part of the large Privalday Ice Lake which came into being about 2000 years earlier and had now an outlet via the Võru Valley.

Lately it was stated (Lepland 1991, Hang 1995, Hang *et al.* 1995) that the terraces in the Võru Valley are glaciolacustrine, not glaciofluvial in origin because they con-

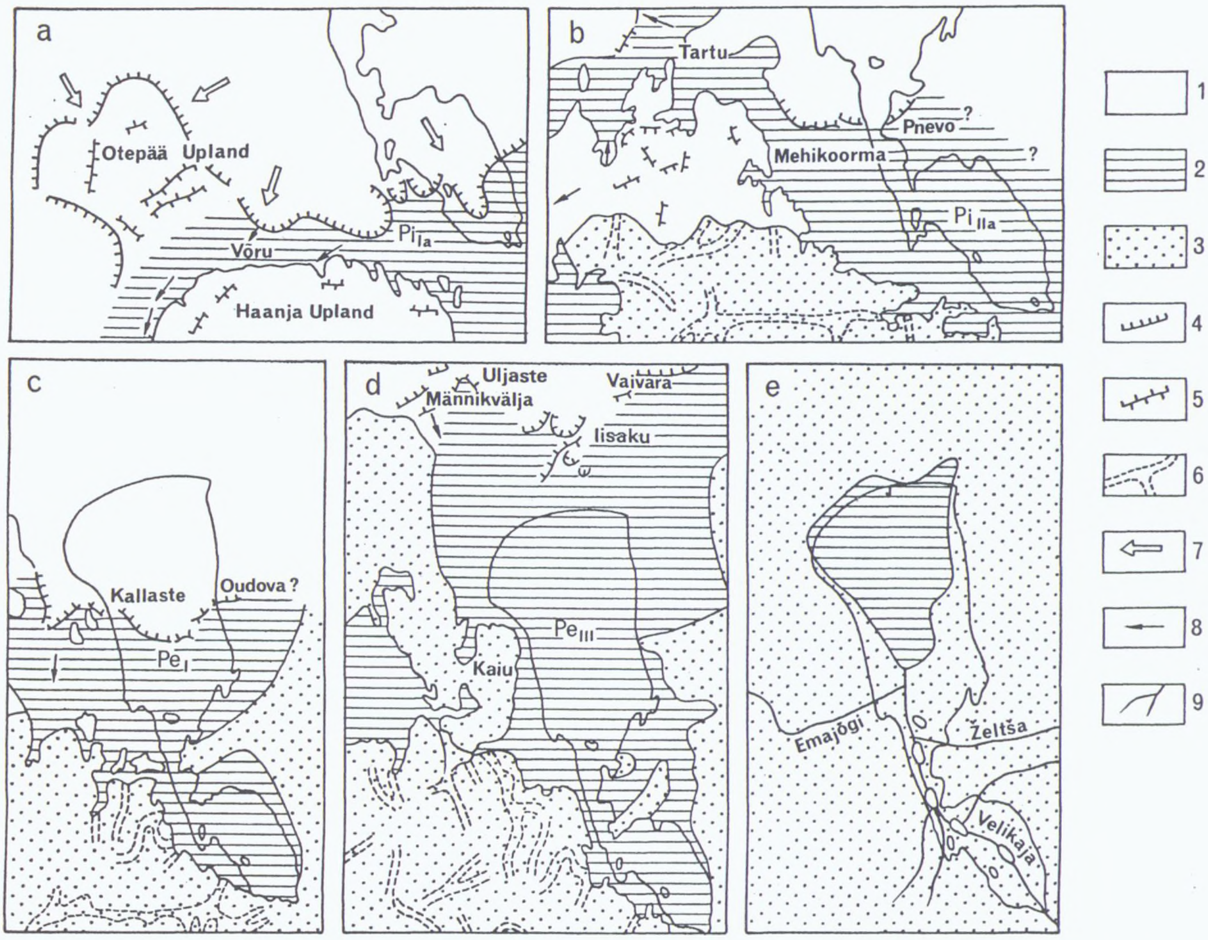


Fig. 177. The development of Lake Peipsi (Раукас и Ряхни 1969): a) Pihkva Ice Lake I (phase $Pi_I a$); b) Pihkva Ice Lake II (Phase $Pi_{II} a$); c) Peipsi Ice Lake (Phase Pe_I); d) Peipsi ice-dammed lake (phase Pe_{III}); e) Lake Small Peipsi. 1 - glacier; 2 - proglacial lake; 3 - dry land; 4 - margins of the active glacier; 5 - field of dead ice; 6 - meltwater valleys; 7 - direction of the ice movement; 8 - direction of the meltwater flow; 9 - rivers.

sist of horizontally bedded fine-grained silts and sands and the terraces have no inclination towards the supposed flow direction. In all likelihood, the Võru Valley was a wide strait which connected ice-lakes in the east and west (Fig. 177a). Further studies are needed to estimate the origin of the terraces in the valley.

Afterwards, when the lake level dropped to 75 m, meltwater flow to the west ceased. It is possible that the outflow was restored along a new strait via the upper courses of the Ahja and Võhandu rivers. However, this connection existed only for a short period.

When the glacier retreated to the line Mehikoorma - Pnevo - Remda, where a dump moraine was formed, the water level sank to 62-60 m a.s.l. and Pihkva Ice Lake II came into being (Раукас и Ряхни 1969, Раукас и др. 1971, see also Fig. 177b). The inflow was from the surrounding heights, occupied by dead ice. The formation of the Ahja and Võhandu valleys started (Fig. 174). There was probably no outflow. The next belt of ice marginal formations has been established on Piirissaar Island and Knyazya Gora (east coast of L. Peipsi, Fig. 178). The water level was at a height of 60 m a.s.l. (Раукас и Ряхни 1969). According to Hang (1995), the terraces of group B in the Ahja and Piusa valleys were synchronous with the Piirissaar glaciofluvial delta and Knyazya Gora end moraine line. Thus, at Knyazya Gora the water level must have been somewhat higher than 60 m a.s.l. The outflow was probably via the Väike-Emajõgi Valley (Fig. 174). After Hausen (1913b), at that time there existed a large Pihkva Ice Lake with its water level 75 m a.s.l. From this lake meltwaters flowed to the west through the Võru Valley and to the southwest via the Valga-Valmiera sandur into the Gauja basin. According to Kvasov (1979, Квасов 1975), the waters of L. Novgorod, a remnant of the split L. Privalday, flowed into Pihkva Ice Lake through a short valley in the vicinity of the town of Porkhov. Later, the connection between Novgorod and Pihkva ice lakes was via a spillway in the middle course of the Luga River. However, Kvasov maintains that Kemba and Voose ice lakes on the northwestern slope of the Pandivere Upland were also synchronous with L. Novgorod, but actually they were formed later when the glacier had retreated northwards (Пярна 1960, Раукас и Ряхни 1969, Раукас и др. 1971).

The further withdrawal of the glacier northwards with a following new advance to the line Kaiu - Gdov about 12,250 yr BP (Fig. 177c) led to the formation of Peipsi Ice Lake I (Раукас и др. 1971). Different views have been expressed as to the height of the lake level. According to Raukas and Rähni (Раукас и Ряхни 1969), in the north-west the lake level was at a height of 86 m a.s.l., at Kaiu 75 m and at Mehikoorma 40 m a.s.l. (Fig. 178). As is known, Peipsi Ice Lake corresponds to the South Peipsi Ice Lake by Hausen (1913b) where the water level was only 36-37 m a.s.l., but Hausen failed to establish any shoreline there. Hang (1995) thinks that the ter-

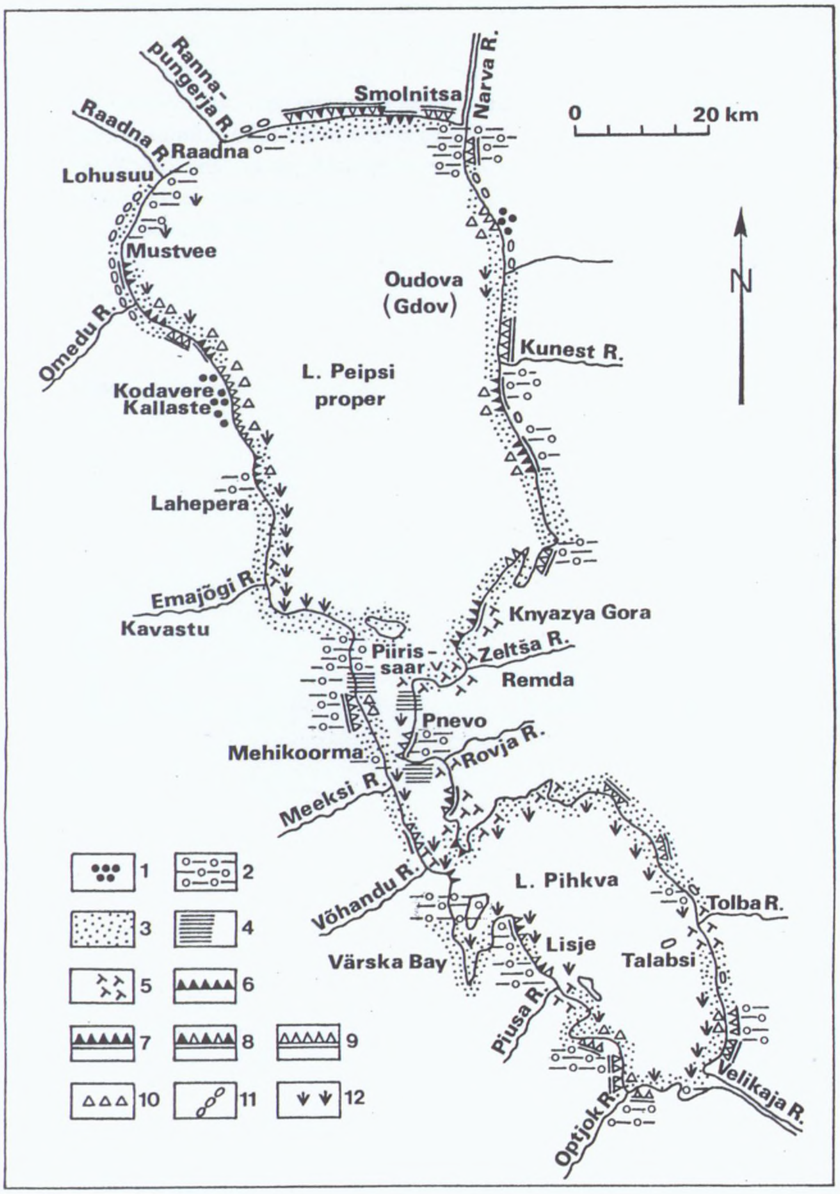


Fig. 178. Shore types of Lake Peipsi (after Raukas & Tavast 1989): 1 - Devonian sandstone; 2 - till; 3 - sand; 4 - clay; 5 - peat; 6 - cliff; 7-9 - bluffs in Quaternary loose deposits: 7 - active; 8 - passive, vegetated; 9 - dead; 10 - erratic boulders; 11 - beach ridge; 12 - bulrush and reed zone.

races of group C in the Ahja and Piusa valleys were formed when the glacier came to a halt at the line Kaiu - Gdov. Considering the altitude of river terraces in these valleys (54-50 and 51-48 m a.s.l.) and the uplift gradient, the water level at Kaiu and Gdov might have been about 51-56 m a.s.l. (Fig. 178).

During the Pandivere Stadial when the glacier readvanced again and came to a halt along the Männikvälja - Iisaku - Vaivara ice marginal formations (Fig. 177d), Peipsi Ice Lake (Pe_{III} after Раукас и Ряхни 1969) formed a single body of water with glacial lakes in the east (Luuga, Neeva a.o.). Kvasov (1979, Квасов 1975) termed it L. Ramsay which corresponded approximately to Great Peipsi Ice Lake by Hausen (1913b). According to Raukas and Rähni (Раукас и Ряхни 1969), the water level in the lake was 80 m a.s.l. at Saare, 70 m at Iisaku and 43 m a.s.l. at Kavastu (Figs. 177d, 178). But after Hausen (1913b), it was 53 m a.s.l. at Iisaku. Hang (1995) associates the highest level of the terrace group D in the Ahja and Piusa valleys (both 41 m a.s.l.) with the shoreline at a height of 47-45.5 m (Fig. 174) which was determined between Kallaste and Kavastu by Liblik (Либлик 1969). On this basis, the calculations have given 50 m a.s.l. for the height of the lake level at Iisaku. At the same time or a bit later, the Emajõgi Valley was formed. At Tartu, the water level was 42-43 m a.s.l. (Mieler 1926, 1927) and its lowering led to incision of meltwaters into Devonian rocks.

After the retreat of the ice from the northern slope of the Pandivere Upland in the vicinity of Männikvälja (Fig. 177a) and Uljaste, ice lakes west and east of the Pandivere Upland joined up (Fig. 167). The event is acknowledged as the beginning of the Baltic Ice Lake (Квасов и Раукас 1970).

After the retreat of the glacier into the Gulf of Finland, the water level dropped and the Peipsi Depression was isolated from the glacial lake, situated in the Gulf. It is supposed that the southern part of the Peipsi Depression dried up in the Younger Dryas or at the beginning of the Holocene (Орвику 1960в, Раукас и Ряхни 1969, Мийдел и Таваст 1981, a.o.). However, opinions have also been expressed that it happened considerably earlier - about 12,000 years ago (Kvasov 1979, Квасов 1975, Hang *et al.* 1995). In both cases, it is supposed that the northern part of the depression was occupied by L. Small Peipsi, into which the Velikaya River and its tributaries (Emajõgi, Võhandu a.o.) discharged (Fig. 177e).

At the beginning of the Pre-Boreal Chronozone, a shallow lake existed in the southern part of L. Peipsi (Пиррус и др. 1985, Давыдова и Киммел 1991, Miidel *et al.* 1995, Hang *et al.* 1995) in which either silts (in the mouth of the Optjok River, Värska Bay, at the Meeksi Brook) or fine-grained sands (in L. Lämmijärv) were deposited (Fig. 178). It is not exactly known what was the altitude of the water level, but in the mouth of the Optjok River it must have been considerably lower than today (Miidel *et al.* 1995). Pollen evidence (Пиррус и Тасса 1981, Пиррус и др. 1985, Давыдова и Киммел 1991, Miidel *et al.* 1995) suggests that at that time the lake was, in general, shallow and its surroundings were paludified. However, there were also deeper areas in the lake, the greatest depth being at least 7.5 m in L. Lämmijärv (Fig. 178).

At the end of the Pre-Boreal (mouth of the Optjok River, Fig. 178) or at the beginning of the Boreal (Värska Bay, Fig. 178), the silty deposits became overlain by organic rich silt or fen peat containing silt. Probably, this abrupt change in sedimentation was caused by a lowering of the lake level. When the formation of fen peat commenced at the mouth of the Optjok River, the water level must have been at least 10 m lower than at present (Miidel *et al.* 1995, Hang *et al.* 1995). The southern part of the lake basin was occupied by a swamp. The lowermost water level in the Holocene between 10,000-9,000 yr BP, was marked by a break in sedimentation (Давыдова и Киммел 1991, Miidel *et al.* 1995, Hang *et al.* 1995).

The slow rise of the water level, following its low stand in the Pre-Boreal, started in the mouth of the Optjok River in the first half (Miidel *et al.* 1995) and in Värska Bay in the second half of the Boreal (Пиррус и Тасса 1981). In the deepest part of the lake, silt sedimentation started anew, but the lake remained shallow (Давыдова и Киммел 1991). This period corresponds to Small Pihkva Lake by Rähni (1973).

During the Atlantic Chronozone, the accumulation of reed peat with shell fragments continued at the mouth of the Optjok River (Miidel *et al.* 1995). In the beginning of the period, paludification started in the mouth of the Kunest River (Fig. 178), in the second half - in the mouth of the Rovya River (Мийдел и др. 1975) and Emajõgi River (Thomson 1939b, Сарв и Ильвес 1975). At the end of the chronozone, paludification on the Island of Gorodets (south-east of Piirisaar Island) commenced (Мийдел и др. 1975). In the north, the lake submerged the area between Omedu and Rannapungerja, its waters extending as far as an ancient shoreline at Raadna (Fig. 178). The water level was 5-6 m higher there than at present. According to Rähni (1973), this lake was the Atlantic Peipsi. It is possible that the outflow from the lake via the Narva River was formed in the Atlantic.

In the second half of the Sub-Boreal, the water level rose rapidly (Miidel *et al.* 1995, Hang *et al.* 1995). The rate was so high that the development of a bog at Laane ceased and the bog was submerged (Пиррус и др. 1985). Swampy conditions spread around the mouth of the Samolva River (Мийдел и др. 1975).

The water-level rise continued in the Sub-Atlantic. According to Rähni (1973), the rapid water-level rise (the Atlantic Great Peipsi) was followed by an abrupt fall of 1.5-2 m. In the Sub-Atlantic, peat deposits were buried under lake deposits in some places.

The water-level rise in the southern part of the depression is associated with the intense glacioisostatic uplift which was faster in the north (Hausen 1913b, Ramsay 1929). The water-level rise was controlled by both tectonic and climatic factors (Miidel *et al.* 1995, Hang *et al.* 1995).

The curve of water-level changes, compiled for the southern part of L. Peipsi (Hang *et al.* 1995), demonstrates a very fast water-level lowering at the end of the Late Weichselian and the succeeding continuous rise since the beginning of the Boreal and fastening in the Sub-Boreal Chronozone (Fig. 179).

The data concerning the Holocene water-level changes in the southern part of L. Peipsi and the Emajõgi (Сарв и Ильвес 1975) and Väike-Emajõgi (Пиррус и др. 1993) rivers are in good agreement.

The lake continues to retreat southwards. According to Vallner with co-authors (Vallner *et al.* 1988), the northern part of the depression is rising at a rate of 0.2-0.4 mm, whereas the southern part is sinking at a rate of 0.8 mm per year. The tilting of the lake depression makes the water to flow from north to south. As a result, the banks of L. Pihkva are suffering from ever increasing erosion; wide stretches of lowlands around the lake have become paludified. In 1796, the area of Piirissaar Island in the southern part of L. Peipsi proper was 20.08 km² (Mieler 1926), to date it is only 7.39 km².

The water-level is slowly rising or more or less stable at the northern coast. Evidence is derived from coastal bluffs in dunes and buried peaty deposits. This is due to the relatively hard rocks, outcropping in the upper course of the Narva River, and long-shore drift, obstructing the outflow (Kajak 1964, Мийдел 1966б, Raukas & Tavast 1996).

The study of the evolution of Lake Peipsi has a long history, but many topical problems have remained unsolved. There are several gaps in the deglaciation history because of the lack of clear ice-marginal formations in the Peipsi Lowland and in its vicinity. Late-glacial river terraces have not yet been properly dated and, frequently, the water-level fluctua-

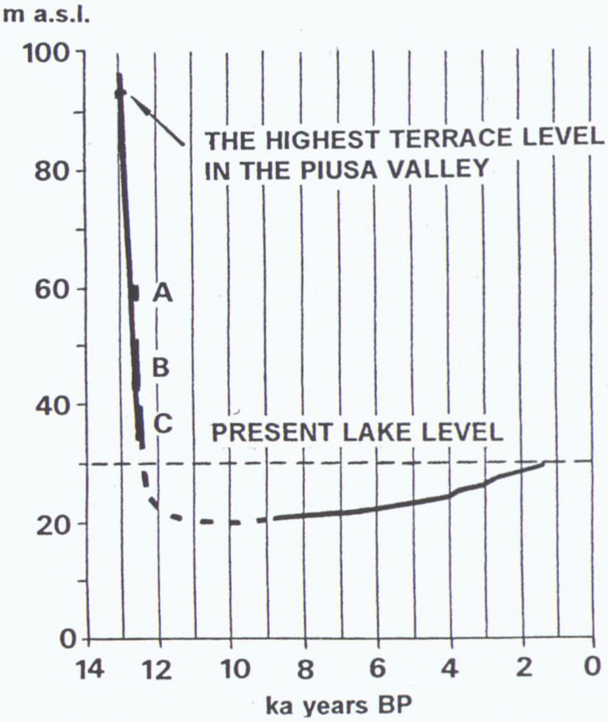


Fig. 179. Lake-level changes in the southern part of Lake Peipsi (Hang *et al.* 1995). A., B, C - altitudes of the groups of Late-glacial terraces in the Piusa Valley (see also Fig. 174).

tions are the best traceable indirect markers (the absolute height of the flat tops of eskers and kames, scarps, developed on the banks of the ice-dammed lakes, boulder fields, *etc.*). In the light of the present knowledge, it may be supposed that there were no long-lasting halts during the retreat of the ice cover promoting the formation of clear terrace surfaces. It is not excluded that the lake depression was filled with passive and dead ice. In this case, the evolution of the lake in the Lateglacial may be interpreted in a principally different way.

There are much more data available on the Holocene history of the lake which was controlled by the neotectonic movements, well dated by means of peat accumulation. Like the majority of lakes in the Northern Hemisphere, Peipsi has a more open eastern and a more swampy and overgrown western bank. Due to the prevailing south-westerly and westerly winds, the active erosion-accumulative or erosional shores are spread in the eastern and northern (Photo 54) parts of the lake, while the swampy coasts overgrown with bushes, bulrush and reed are characteristic of its western (Photo 55) and southern parts (Fig. 178).

Photo 54. Peipsi beach at Smolnitsa. *Photo by A. Miidel.*

Photo 55. The shores in the western part of the lake are often swampy and overgown with bushes, bulrush and reeds. The shore at Lohusuu. *Photo by A. Miidel.*

Lake Võrtsjärv

Võrtsjärv, the second largest lake in Estonia, has a surface area of 270.7 km². Its maximum length is 34.8 km and maximum width 14.8 km. The length of the weakly dissected shoreline measures 96 km, maximum depth is about 6 m, average depth 2.8 m, long-term water level stand 33.68 m, volume 756 million m^3 of water, catchment area 3380 km^2. The main tributaries number 18, the outflow is via the Emajõgi River (Mäemets & Raukas 1995).

The lake depression was formed in pre-Quaternary time, but during the course of thousands of years its shape has been radically altered by the glaciers. The orientation of the drumlins and clasts in the till of the last glaciation (Raukas & Tavast 1990) shows a more or less meridional direction of the ice movement between the Sakala and Ugandi plateaus.

The bedrock, mainly sand- and siltstones of the Middle Devonian Aruküla Stage, is exposed on the lake's steep east bank at Tamme (up to 8.5 m high), Trepimägi and Petseri (Fig. 180). Devonian rocks are rich in quartz (75-90 %) and micas (1-10 %). Heavy minerals are dominated by ilmenite, zircon, garnet and tourmaline (Kleesment 1994). The same minerals are widespread in the lake deposits. The carbonates (up to 14.1 % of fine sand fraction), amphiboles and pyroxenes (up to 30 % of heavy subfraction) occurring in beach deposits have been washed out from glacial sediments (Таваст 1990), rich in those minerals (Раукас 1978).

The depression around the lake is covered with till, glaciofluvial sand and gravel, glaciolacustrine silt and clay, gyttja, lake marl and peat (Орвику Л. Ф. 1958). In some places aeolian and alluvial sandy-silty sediments occur. The thickness of the deposits is mainly 5-10 m, seldom more (Раукас 1978).

Vast stretches of the low-lying shore are overgrown with reed and bulrush. In the west and east, reed forms an up-to-100-metre-wide belt. The small depth, high water temperature in summer and increased concentration of mineral nutrients promote overgrowing of the lake. The encroaching reed will hardly reduce the area of beaches available for recreational use (Tavast *et al.* 1983). The best sandy beaches are in the north at Vaibla and in the east in Vehendi Bay. Figure 180 shows different types of contemporary beaches and prevailing grain-size of shore deposits.

The bottom sediments consist mostly of fine sand and silt, sapropel (up to 9 m) and lacustrine lime (up to 8 m thick). In the northern part of the lake, sapropel and lacustrine lime either form a thin layer or are entirely absent (Veber 1973). About two thirds of the topmost part of the lake's sediments consist of sapropel (gyttja) and sandy sapropel (Fig. 181), with the total volume of about 200 million m^3, together with the lake marl ca 360 million m^3 (Veber 1973). Organic matter forms 87-92% of the sapropel. Silty clay, lacustrine lime and other types of sediments are less abundant. In places, especially in the northern part where the bottom sediments are absent, the lake depression exposes varved clay or till. In the southern part of the lake, the sediments are much thicker than in the northern part indicating a gradual rise of the water level in the southern portion of the basin (Pirrus & Raukas 1984).

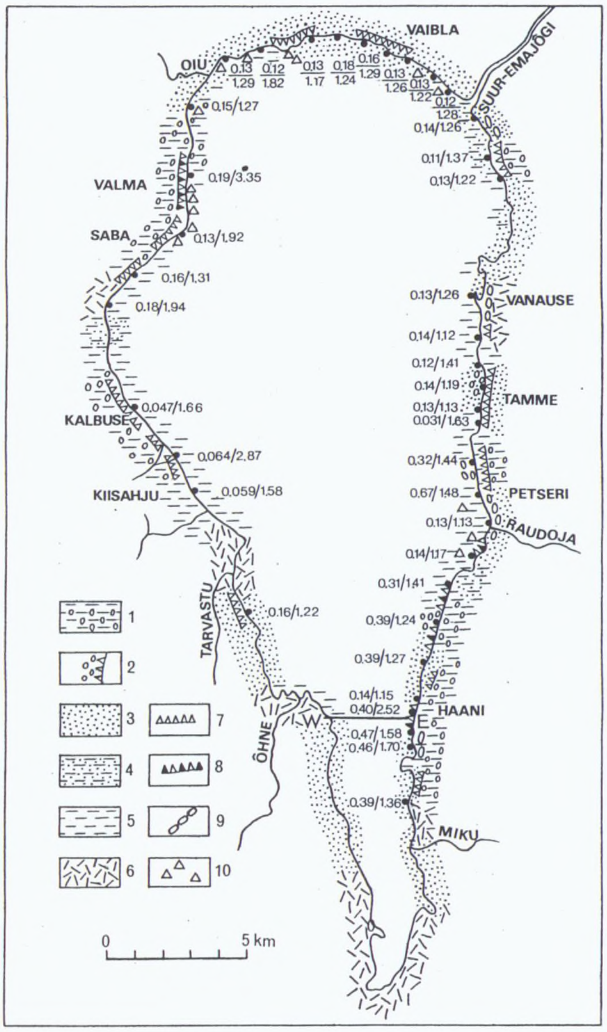


Fig. 180. Shore types and grain-size of beach deposits (numerator - medium diameter, denominator - coefficient of sorting) after Tavast (Таваст 1990) with complements: 1 - till; 2 - cliff in sandstones; 3 - medium size sand; 4 - silty sand; 5 - silt; 6 - peat; 7 - dead bluff; 8 - passive bluff; 9 - beach ridge; 10 - boulders; 11 - line of cross-section (Fig. 182).

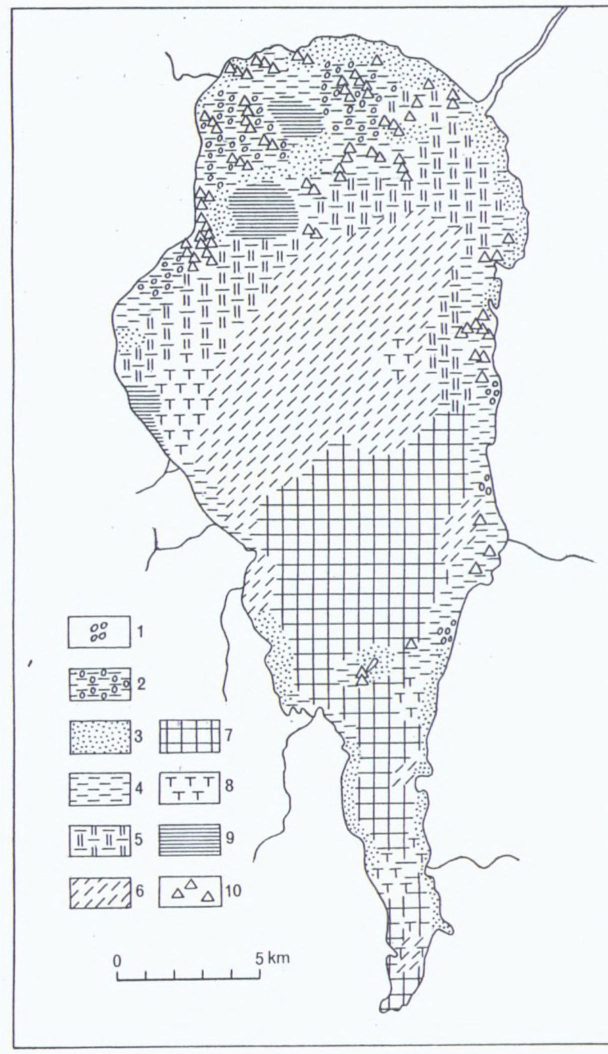


Fig. 181. Bottom sediments of Lake Võrtsjärv after Pirrus, Raukas and Tavast (Raukas & Tavast 1990, with complements): 1 - Devonian sandstone; 2 - till; 3 - sand, silty sand and gravelly sand; 4 - silt and sandy silt; 5 - silty clay; 6 - sandy sapropel (gyttja); 7 - sapropel (gyttja); 8 - lake marl; 9 - (varved) clay; 10 - boulder and cobble ridges (rewashed drumlins).

Figure 182 presents a characteristic cross-section of lake sediments.

The lake has a complicated history (Fig. 182). Glacial lakes of different shape and size were formed immediately in front of the retreating ice cover about 12,600 (Fig. 183A), 12,250 (Fig. 183B) and 12,050 (Fig. 183C) yr BP during the Otepää, Sakala and Palivere stadials, respectively (Raukas 1986). The outflow from those lakes was first to the south via the Väike-Emajõgi Valley to the basin of the Gauja River. Afterwards new outflows were formed to the west via the Viljandi Valley. In the Younger Dryas, at the end of the Late-glacial when meltwater of the glacier diminished, Lake Ancient Võrtsjärv (Orviku 1973) came into being in the Võrtsjärv Depression. Due to the neotectonic uplift, which was more intensive in the north-west, the outflow to the west gradually diminished and closed in the Early Holocene. In the depression, Lake Big Võrtsjärv (Орвику Л. Ф. 1958) was formed (Fig. 183D). The water level in the northern part of the basin was 4-5 m higher than today. At the beginning of the Middle Holocene, about

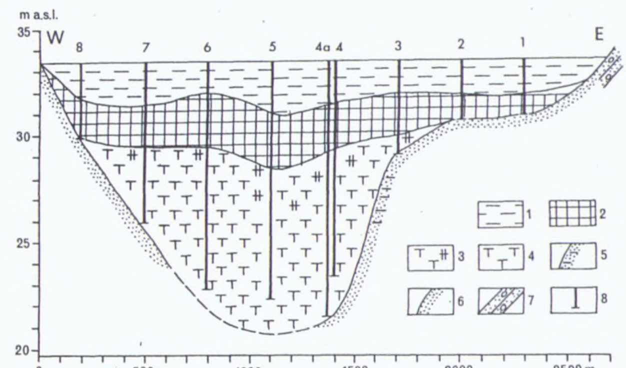


Fig. 182. Cross-section of lake sediments between Sula and Haani after Pirrus and Raukas (1984): 1 - water; 2 - sapropel (gyttja); 3 - sapropelic lake marl and limy sapropel; 4 - lake marl; 5 - silty sand; 6 - sand; 7 - till; 8 - boreholes.

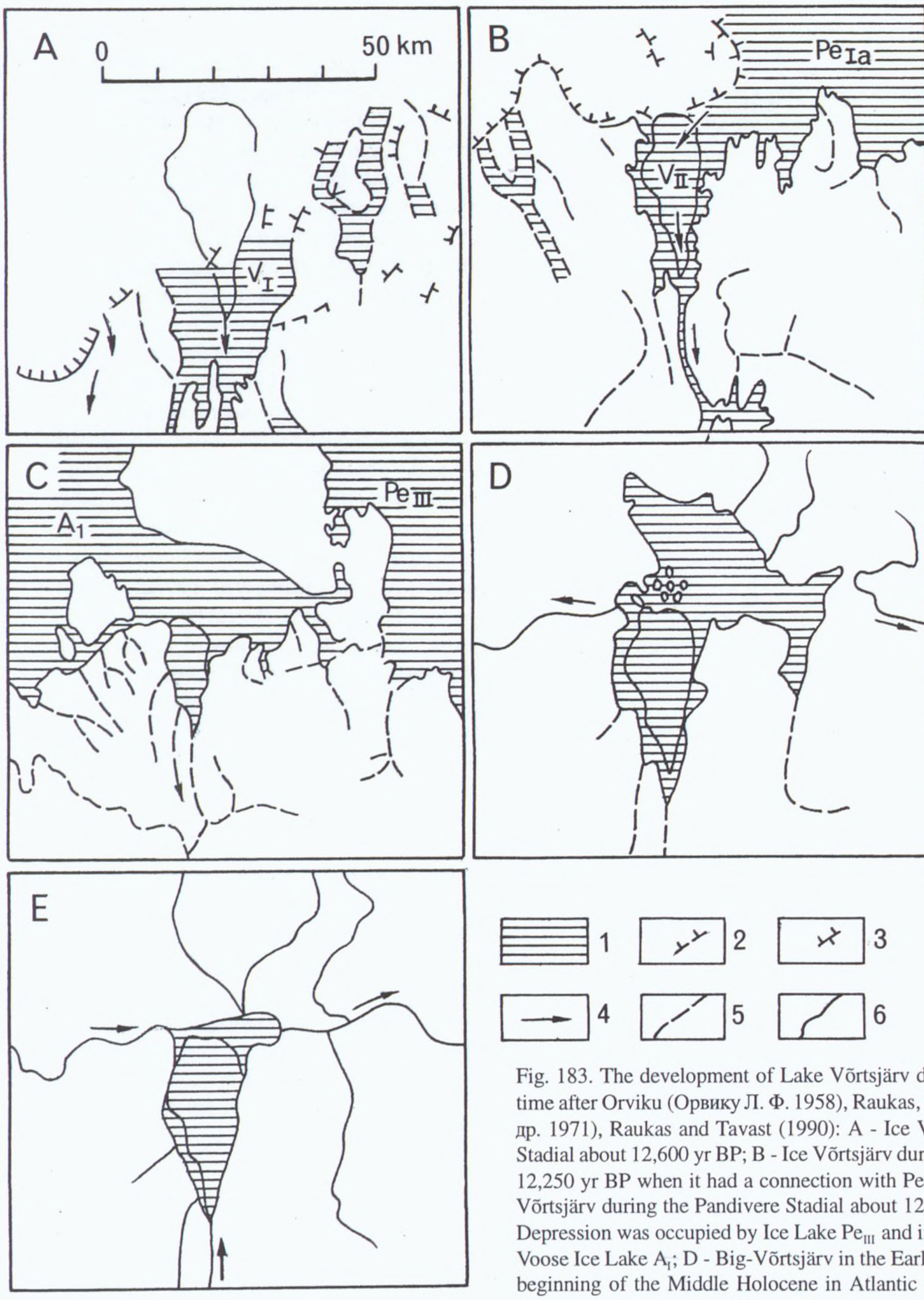


Fig. 183. The development of Lake Võrtsjärv during late- and post-glacial time after Orviku (Орвику Л. Ф. 1958), Raukas, Rähni and Miidel (Раукас и др. 1971), Raukas and Tavast (1990): A - Ice Võrtsjärv during the Otepää Stadial about 12,600 yr BP; B - Ice Võrtsjärv during the Sakala Stadial about 12,250 yr BP when it had a connection with Peipsi Ice Lake Pe_{Ia}; C - Ice Võrtsjärv during the Pandivere Stadial about 12,050 yr BP when the Peipsi Depression was occupied by Ice Lake Pe_{III} and in western Estonia there was Voose Ice Lake A_I; D - Big-Võrtsjärv in the Early Holocene; E - Lake in the beginning of the Middle Holocene in Atlantic time after the formation of outflow via the Emajõgi Valley to the east: 1 - water bodies; 2 - ice margin; 3 - dead ice; 4 - supposable direction of the water flow; 5 - valleys; 6 - rivers.

7500 yr BP, an outflow to the east developed via the Emajõgi Valley (Fig. 183E), and gradually the lake acquired its present contours (Orviku 1973).

At present, due to the uneven neotectonic uplift of the lake's depression, Lake Võrtsjärv is steadily retreating southwards inundating new areas (Jaani 1973). The water level in the lake is very unstable (in 1922 the maximum annual amplitude was 2.2 m, the maximum difference of the water table is up to 3 m, the rise during the spring flood up to 174 cm).

In the last century, the water level of the lake was about one metre higher than today. In the 1920s, the outlet of the Emajõgi River was thoroughly dredged, and the lake level dropped. Lake Võrtsjärv is rich in phytoplankton, 36 fish species inhabit the lake. Valuable commercial fish form about 70 per cent of the total catch (Mäemets & Raukas 1995). The regulation of water level would create more favourable conditions for the spawning of valuable fish and survival of their larvae. However, beforehand, careful study of potential environmental consequences is needed (Raukas & Tavast 1990). Without doubt, rising of the lake level will promote erosion processes both in Lake Võrtsjärv and on the banks of the Emajõgi River.

Small lakes

Introduction

There are about 1500 small lakes with a surface area less than 10 km^2 located irregularly in different landscape regions in Estonia (Mäemets & Saarse 1995). The Haanja and Otepää heights, the Saadjärve Drumlin Field and the Kurtna Kame Field are dotted with lakes; in the West-Estonian Lowland their number is small. Because of the infilling and overgrowing, the number and area of lakes is constantly decreasing, but due to the land uplift in the west and north-west (up to 3 mm per year) isolation of new waterbodies from the Baltic Sea is in progress. More than half of small lakes are glacial in origin and scattered in Upper Estonia. Basins of small lakes are filled with Late-glacial sand, silt and clay, Holocene organic and calcareous deposits which store information on the postglacial stratigraphy, vegetational history and climate change. The longest lake records start from the Older Dryas, most commonly from the Younger Dryas. Organogeneous deposition began at the beginning of the Holocene with some delay in most kame field lakes.

In the following, we shall deal with the evolution of the main types of small lakes (Saarse 1990) through the palaeoecological regions (Saarse & Raukas 1984).

Lakes of the Middle and Upper Devonian plateaus

The *Otepää and Haanja heights* are characterized by a highly disjointed topography containing some four hundred lakes with a small surface area, disjointed bottom relief and varying trophic conditions (Mäemets 1977). Glacial and residual lakes are the basic types in this region. It is almost impossible to distinguish between the lakes originating from the irregularities of the glacial drift and lakes of glaciokarst origin. Glaciolacustrine beds are commonly absent, lacustrine clayey-silty sediments are thin or non-existent (Mäetilga, Vaskna, Tuuljärv, Kurgjärv, Väikjärv, Päidla, Mähe, Ahvenjärv, Räbijärv, Juusa; Fig. 184) (Мяэметс 1983, Ilves & Mäemets 1987, Ilves 1980, Саарсе 1994). Steep slopes and jointed catchment topography promote abundant input of minerogenous matter into the lakes. In the littoral belt of some lakes lacustrine lime and calcareous gyttja have deposited. In L. Kurgjärv peat is buried under gyttja (Fig. 185). Buried peat occurs also on the steep slopes of the lakes Pangodi, Mäetilga and Tuuljärv, obviously drifted down during the solifluction. Pollen and ^{14}C records of bottom deposits indicate that organic sedimentation started at different times in the Pre-Boreal, but the lakes themselves, at least some of them, were evidently formed since the Haanja Stadial.

The basins of the residual lakes were first occupied by proglacial lakes. After the ice recession and disappearance of proglacial lakes, residual water bodies remained as separate lakes (Tamula, Pangodi, Kirikumäe, Pulli, Pühajärv). The mentioned evolutional changes are fixed in their bottom deposits: till or glaciofluvial sand and gravel covered with glaciolacustrine varved clays and laminated sands and silts, overlain by gyttja or calcareous gyttja. The thickness of organic lacustrine deposits varies from 2-3 m in Koobassaare, Saarjärv and Neitsijärv, up to 8 m in L. Pühajärv (Photo 56), 9 m in L. Vaskna and 11 m in L. Vagula (Ramst 1992). Biostratigraphical studies have been carried out from sequences of lakes Tamula (Пиррус 1969), Mäetilga, Kõverjärv (Мяэметс 1983), Tuuljärv, Vaskna (Ilves & Mäemets 1987), Pulli, Kirikumäe, Punso (Саарсе 1994), Pühajärv (unpublished), *etc*. Most of the sequences are supplemented by radiocarbon dates. From Pangodi the pollen diagram is absent and only radiocarbon dates are available (Ilves 1980).

Lakes in kettles left by melting ice blocks into pre-existing valleys and lakes formed in irregularities of hilly topography due to uneven deposition, are characteristic of the *Sakala Upland* and the *Ugandi Plateau*. Lake basins are elongated, they have uneven floors and a rather complicated sediment composition. In the Holocene part of the lacustrine deposits, calcareous and minerogenous gyttja dominates (Saarse *et al.* 1995b). Deep basins with steep underwater slopes and negligible assemblages of aquatic plants are poor in lacustrine deposits (Viljandi, Mäeküla, Parika, Pärsti; Ramst 1992), whereas in shallow (5-10 m) lakes organogenous deposits have accumulated (Õisu, Võistre, Kariste, Veisjärv). According to the pollen stratigraphy, minerogenous lacustrine sediments formed since the Allerød (Saarse *et al.* 1995b; Пиррус 1969), and were succeeded in the Pre-Boreal by organic and calcareous deposits. Varved clays are absent in the lakes which served as a drainage route for the glacial meltwater and were, thus, subject to erosion. The lakes Päidre (Пиррус 1969, Saarse *et al.* 1995c), Võistre (Саарсе 1994), Sinialliku and Viljandi (Lõokene 1979) have been palynologically studied, and only from L. Päidre about 20 radiocarbon dates are available (Saarse *et al.* 1995c).

Lakes of the Central Estonian Watershed

This region includes the Pandivere Upland and the Saadjärve and Türi drumlin fields. The drumlin lakes in the *Saadjärve Drumlin Field* are also glacial in origin, resulting from a complicated combination of deposition and erosion. These lakes with uneven longitudinal and transitional profiles are elongated towards the direction of the ice movement and contain a complete stratigraphic succession of Late-glacial and Holocene deposits (Pirrus & Rõuk 1979, Pirrus *et al.* 1987, Саарсе и Кярсон 1982). Older Dryas sediments are represented by a 1-3-m-thick complex of yellowish-brown varved clays. Allerød units are dark-grey and greenish-grey silts and silty clays with dispersed organic matter. Younger Dryas laminated silts are coarser, frequently with fine-grained sand and moss interbeds. The boundary between the Late-glacial and Holocene deposits is sharp: the minerogenous deposits are replaced by organic or calcareous ones. In shallow and medium depth lakes, during the Holocene two carbonate-rich units separated by an interlayer of organic gyttja were formed (Fig. 186). The total thickness of these units reaches 6 m in L. Elistvere, 7.5 m in L. Pikkjärv (Саарсе и Кярсон 1982) and 8.5 m in L. Raigastvere (Pirrus *et al.* 1987, Saarse *et al.* 1995b). These lakes were formed during the Otepää and Pandivere stadials of the ice recession. First they were submerged by a local proglacial lake, at the bottom of which varved clays deposited. When the threshold of the proglacial lake was freed of ice, the water level dropped and independent development of the inter-drumlin lakes started. Pollen diagrams are available from lakes Raigastvere (Pirrus *et al.* 1987), Pikkjärv (Pirrus & Rõuk 1988), Soitsjärv (Pirrus & Rõuk 1979), Elistvere (Pirrus, unpublished), Prossa (Kihno,

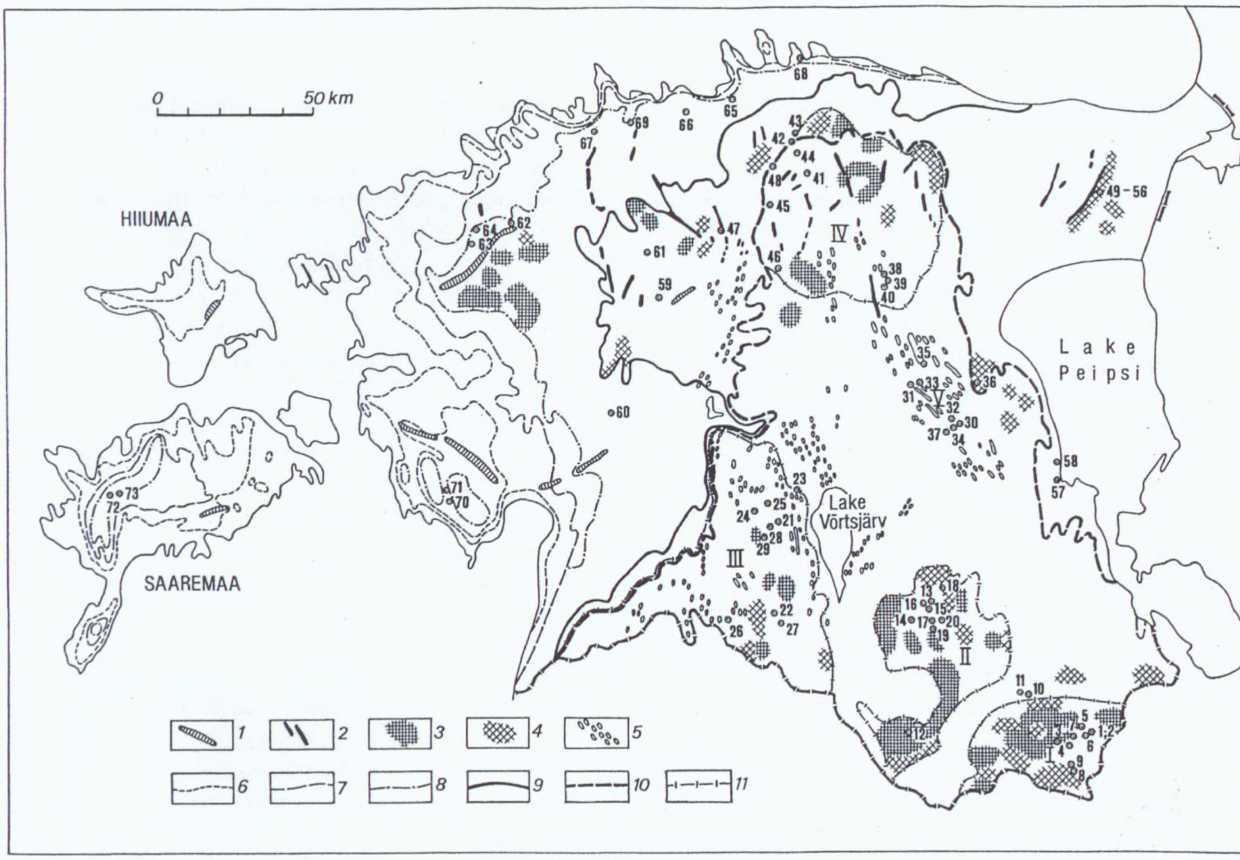


Fig. 184. Location of lakes mentioned in the text: 1 - end moraine and marginal esker; 2 - radial esker; 3 - morainic topography; 4 - kame field; 5 - drumlin field; 6 - limit of the Limnea Sea; 7 - limit of the Litorina Sea; 8 - limit of the Ancylus Lake; 9 - limit of the Baltic Ice Lake B_{III}; 10 - limit of the Baltic Ice Lake B_I; 11 - boundary of heights and uplands. Location of lakes: 1 - Kirikumäe; 2 - Punso; 3 - Kurg-; 4 - Väik-; 5 - Mäetilga; 6 - Vaskna; 7 - Tuul-; 8 - Pulli; 9 - Saar-; 10 - Tamula; 11 - Vagula; 12 - Koobassaare; 13 - Päidla Suur; 14 - Mähe; 15 - Ahven-; 16 - Räbi-; 17 - Juusa-; 18 - Pangodi; 19 - Püha-; 20 - Neitsi-; 21 - Viljandi; 22 - Mäeküla; 23 - Parika; 24 - Pärsti; 25 - Võistre; 26 - Kariste; 27 - Veis-; 28 - Sinialliku; 29 - Päidre; 30 - Elistvere; 31 - Pikk-; 32 - Raigastvere; 33 - Prossa; 34 - Soits-; 35 - Kuremaa; 36 - Kaiu; 37 - Saad-; 38 - Äntu Sini-; 39 - Äntu Valge-; 40 - Linaleo; 41 - Neeruti Ees-; 42 - Viitna Pikk-; 43 - Lina-; 44 - Vohnja Kõver-; 45 Niker-; 46 - Matsimäe Püha-; 47 - Kiruvere; 48 - Udriku Suur-; 49 - Räätsma; 50 - Haug-; 51 - Räätsma; 52 - Martiska; 53 - Konsu; 54 - Liiv-; 55 - Lina-; 56 - Ümar-; 57 - Kalli; 58 - Lahepera; 59 - Imsi; 60 - Kaisma; 61 - Järlepa; 62 - Järveotsa; 63 - Must-; 64 - Valge-; 65 - Kahala; 66 - Rummu; 67 - Ülemiste; 68 - Käsmu; 69 - Maardu; 70 - Ermistu; 71 - Tõhela; 72 - Karu-; and 73 - Mustjärv.

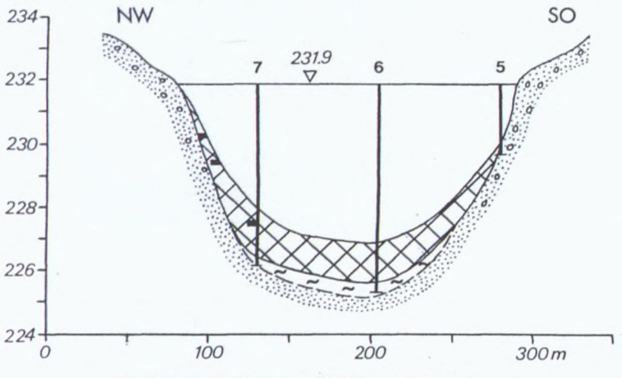


Fig. 185. Cross-section of L. Kurgjärv: 1- gyttja; 2 - peat; 3 - gravel; 4 - macroremains.

Photo 56. Lake Pühajärv is rich in organic deposits. *Photo by A. Raukas.*

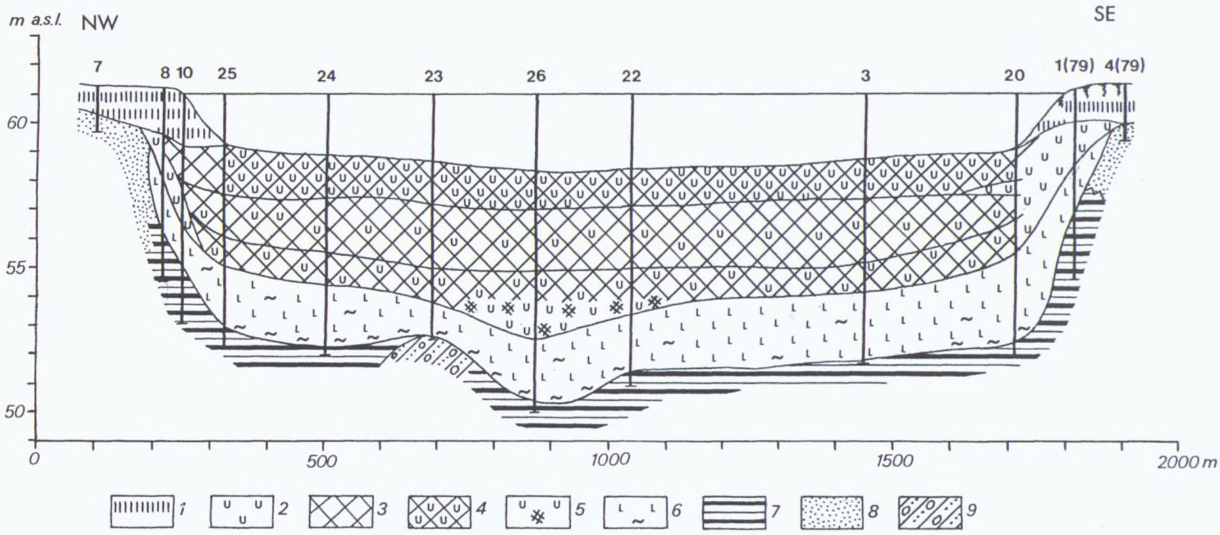


Fig. 186. Cross-section of L. Prossa with two carbonate-rich gyttja layers: 1 - fen peat; 2 - lacustrine lime; 3 - gyttja; 4 - calcareous gyttja; 5 - organic rich lime; 6 - silt with moss remains; 7 - varved clay; 8 - sand; 9 - till.

unpublished), Saadjärv, Kaiu (Zirna & Pirrus 1961). The sequences from lakes Raigastvere, Elistvere and Prossa have been dated by the radiocarbon method (Pirrus *et al.* 1987, Ilves 1980).

On the *Pandivere Upland* the Ordovician and Silurian limestones crop out or lie under a thin Quaternary mantle causing the calcite dissolution in cracks, favoured by the zones of tectonic dislocations, now occupied by karst lakes (Raukas 1993). The karst lakes in sink-holes are temporary bodies of water, devoid of lacustrine sediments. The glaciokarst lakes of esker ridges (Photo 57) and kame fields are most widespread near Nõmmküla (Joonuks 1967).

The Äntu group of lakes has been studied in particular detail (Saarse & Liiva 1995). L. Äntu Sinijärv is rare with its highly transparent water and thick lacustrine lime unit which accumulated throughout the Holocene. The thickness of lacustrine calcareous deposits in other lakes is moderate, commonly 2-4 m, in L. Äntu Valgejärv it reaches 5 m; gyttja in L. Neeruti Eesjärv is up to 6 m thick (Caapce 1994).

A special group of lakes are the alkaline dammed valley lakes which existed during the Early Holocene, some of them until the end of the Atlantic. Their sediments consist of Lateglacial clayey units and Holocene lacustrine lime with basal peat in several basins (Kärsa, Lehtse, Tapa, Vatku, Kadrina; Männil 1961), indicating the low lake level status in the Pre-Boreal and at the beginning of the Boreal. The average thickness of lacustrine deposits is 2-3 m, in L. Vatku – up to 7 m.

Because of the prevalence of highly calcareous deposits, few lacustrine sequences of this region are radiocarbon dated (Äntu Sinijärv, Valgjärv, Linaleo; Saarse & Liiva 1995a).

Intermediate Estonia (*Kõrvemaa*) the transitional region between the West-Estonian Lowland and the Pandivere and Sakala uplands, abounds in glaciokarst kettle-hole lakes (Viitna Pikkjärv, Linajärv, Vohja Kõverjärv, *etc.*). Their formation is closely related to the formation of esker ridges and kame fields (Saarse 1992). Kettle lakes in pitted outwash are rich in organic gyttja (Nikerjärv, Matsimäe, Viitna Pikkjärv, Linajärv). The lake sediments in morainic topography are mostly characterized by calcareous gyttja (Kiruvere; Fig. 187). The lacustrine sedimentation started at least in the Younger Dryas with

Photo 57. L. Neeruti Tagajärv was formed in a glaciokarst hollow near the esker ridge. *Photo by Ago Aaloe.*

the formation of sandy-silty beds. In some lakes minerogenous deposition continued in the Pre-Boreal and even in the Boreal (Udriku Suurjärv, Vohnja Kõverjärv; Caapce 1994). In most of the lakes studied in this region, gyttja accumulated since the Pre-Boreal onwards. The thickness of lacustrine deposits varies between 2.5-3 m, reaching 5 m in L. Vohnja

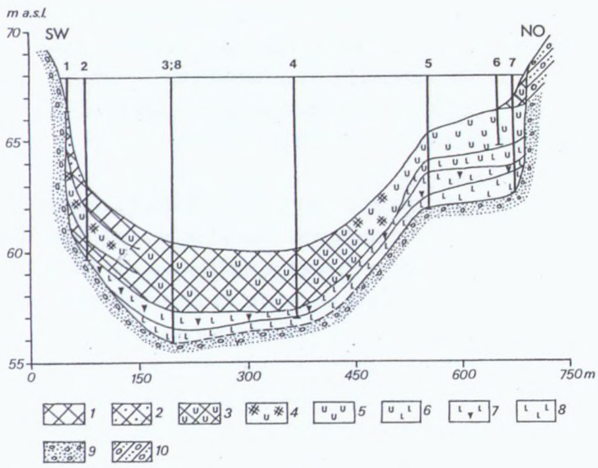


Fig. 187. Cross-section of L. Kiruvere: 1 - gyttja; 2 - sandy gyttja; 3 - calcareous gyttja; 4 - organic rich lacustrine lime; 5 - lacustrine lime; 6 - silty lacustrine lime; 7 - silt with plant remains; 8 - silt; 9 - sandy gravel; 10 - till.

Kõverjärv, 8 m in L. Nikerjärv, and 8.7 m in L. Viitna Linajärv. Palynologically studied lakes are Viitna Pikkjärv and Linajärv (Pirrus, unpublished), Kõverjärv, Udriku Suurjärv and Kiruvere (Саарсе 1994). Radiocarbon dated deposits come from lakes Viitna Pikkjärv and Linajärv.

The **Peipsi and Võrtsjärv lowlands and the Alutaguse area** are rich in mires, residual and telmatogenic lakes and kame field lakes. As elsewhere, the kame field lakes are rich in gyttja with a small amount of minerogenous and calcareous compounds. The basal glaciolacustrine and lacustrine clay and silt in kame field lakes are commonly absent. In L. Valgejärv (Illuka Kame Field), basal peat is buried under gyttja (Саарсе и др.1985). The depth and bedding condition of the lacustrine deposits are variable due to the differences in the bottom topography, water chemistry, trophic stage and biological productivity. In the studied lakes, the average thickness of gyttja is 2-4 m (Саарсе и др. 1985), maximum 8 m in L. Räätsma. The sequences of lakes Haugjärv, Räätsma, Martsika, Konsu, Liiv-, Lina- and Ümarjärv have been biostratigraphically studied and except the two former, supplemented by radiocarbon dates (Саарсе и др. 1985, Koff 1994).

The deposition of gyttja started at the beginning of the Pre-Boreal in L. Konsu, and at the end Pre-Boreal in the other studied lakes.

The lakes situated east of L. Peipsi often have bottom peat (Kalli, Lahepera) (Паап и др. 1981, Orru 1995), obviously Pre-Boreal or Boreal in age, unfortunately not yet bio- and chronostratigraphically studied.

Residual coastal lakes in **northwestern and western Estonia** are few in number on the terraces of the Baltic Ice Lake (Imsi, Kaisma, Järlepa, Järveotsa, Karujärv, Mustjärv), but rather numerous in the modern coastal area. Lakes, left on the terraces of the Baltic Ice Lake, occupy depressions in bedrock hollows or in the gently rolling glacial relief, later modified by the Baltic Ice Lake. These lakes are shallow, but rather large in area. Their independent development began at the end of the Late-glacial, after the area had emerged from the Baltic Ice Lake. The basal glacial and glaciolacustrine deposits are commonly covered by clayey lacustrine units, lacustrine lime and gyttja. Pollen records are available from lakes Järlepa, Järveotsa, Karujärv and Mustjärv (Саарсе 1994, Poska 1994).

The number of residual coastal lakes is high in northern and western parts of mainland Estonia and on the islands of the West-Estonian Archipelago. Their formation started after the regression of the Yoldia Sea (Kahala, Rummu, Ülemiste). Those lakes occupy former depressions, but some of them are dammed up by spits, bay-mouth bars and beach ridges, mostly formed during the Ancylus transgression (Ülemiste, Pitkasoo, *etc.*). The basins of coastal lakes are filled with organic or clayey gyttja. In some lakes (Ülemiste, Käsmu, Ermistu), the lithostratigraphy of the bottom beds is more complicated, comprising also calcareous or minerogenic deposits (Fig. 188). The mineral bottom of lakes Ülemiste, Ermistu, Valge- and Mustjärv is covered by Pre-Boreal peat or peaty gyttja, overlain by lacustrine bed.

Of about 80 lakes on Saaremaa Island, most are young polyhalobous bodies of water with thin or non-existent lacustrine deposits, which isolated from the sea recently due to the land uplift. The oldest lakes on Saaremaa are located on the Western Saaremaa Elevation. According to pollen and ^{14}C records, the accumulation of organic deposits started very soon after the emergence of this area from the sea, on the transition

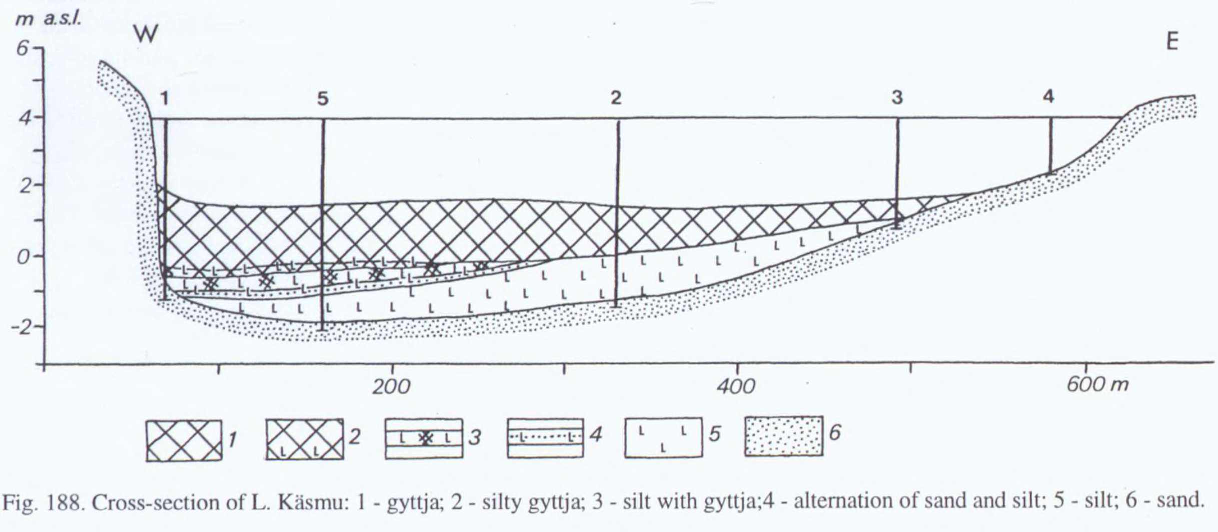


Fig. 188. Cross-section of L. Käsmu: 1 - gyttja; 2 - silty gyttja; 3 - silt with gyttja;4 - alternation of sand and silt; 5 - silt; 6 - sand.

from the Pre-Boreal to the Boreal (Pelisoo, Pitkasoo, Käesla; Saarse & Königsson 1992, Кессел и Раукас 1967). During the Ancylus regression, these basins became shallow and paludified.

The biostratigraphy of lacustrine sequences on the terraces of the Baltic Sea was studied by Kessel (1961, Кессел и Раукас 1967). New high-resolution pollen diagrams are available from lakes Kahala (Poska, unpublished), Maardu (Veski 1992, 1996b), Ermistu, Tõhela, Kiilaspere, Mustjärv (Veski, unpublished), Pitkasoo, Surusoo and Vedruka (Saarse & Königsson 1992, Veski 1996a, Poska, unpublished) which, besides the Baltic Sea history, provide information on human activities (Poska 1994, Veski & Lang 1996).

Palaeogeographic conclusions

About 10% of Estonia's small lakes have been geologically studied. This is far from being sufficient for drawing any fully acceptable conclusion on the development of these lakes. Lakes in southern Estonia with almost complete Late-glacial lithostratigraphic record since the Older Dryas, have not been subject to detailed studies. Lakes in central Estonia, particularly in the Vooremaa area, contain almost full Late-glacial sequences which have been studied well. The lakes situated in northern and western Estonia (Ülemiste, Maardu, Kahala, Ermistu, Mustjärv), have been studied in detail, but they have less complete Late-glacial sequences.

Late-glacial biostratigraphic records are rather representative in Raigastvere, Kirikumäe, Päidre and Kahala lake deposits. Complete Holocene records are available from most of the small lakes in Upper Estonia, but also from some lakes in Lower Estonia, *e.g.* Ermistu, Pitkasoo, Kahala. Several lakes, including Kirikumäe, Tuuljärv on the Haanja Heights, Päidre on the Sakala Upland, Raigastvere in the Saadjärve Drumlin Field, Äntu on the Pandivere Upland, Kahala and Maardu on the North-Estonian Plateau, Ermistu on the West-Estonian Lowland and Karujärv on Saaremaa Island, can serve as Holocene reference sites.

The sediment composition in small lakes is variable and differs with regions. Organic, rather homogenous sediments are characteristic to the kettle lakes in pitted outwash (Kirikumäe, Viitna, Kurtna, Aegviidu), calcareous sediments prevail in the lakes fed by groundwater (Valgjärv, Äntu Sinijärv, Väinjärv), and mixed sediments are typical of lakes with calcareous tilly catchment (Päidre, Võistre, Järveotsa, Ülemiste).

Lacustrine sequences store information on the evolution of lakes, on the vegetation and the Baltic Sea history (in western Estonia), but also on human activities and climatic changes.

Lake level fluctuations serve as an useful tool for the moisture balance control. Lakes respond to changes in the local hydrological balance by changing in depth and area (Saarse & Harrison 1992). The lake level reconstruction of the 13 Estonian lakes (Saarse *et al.* 1995a) shows that the most pronounced lake level lowering occurred about 9000-8000 and 4000-3000 yr BP. The lakes in Estonia were at the highest level about 9500, 7000, 3000 yr BP and at the present. This indicates drier conditions during the second half of the Pre-Boreal and Boreal and in the middle of the Sub-Boreal, and wetter conditions at the beginning of the Pre-Boreal, in the Early Atlantic, and at the end of it. The long-term trend in climatic change has been explained as a result of changes in the Earth´s orbital parameters (Berger 1978, Berger *et al.* 1995). In the Pre-Boreal and Boreal, the Northern Hemisphere insolation was about 8% greater than present in summer and 8% less in winter, creating very severe continental conditions and higher evapotransipation during the summer season (COHMAP Members 1988). Latest model simulations have shown that main forcing for the lake level fluctuations is precipitation and evaporation, as well as cloudeness (Harrison *et al.* 1993).

In western Estonia, the lacustrine sequences have been used to reconstruct the Baltic Sea history (Kessel 1961) and shore displacement curves (Kessel & Raukas 1979). The Latest studies on coastal lakes (Maardu, Kahala, Ülemiste, Ermistu) suggest that the Ancylus transgression started about 9500-9300 yr BP (Саарсе и др. 1990, Раукас и Хюваринен 1992, Saarse *et al.* 1995c). Prior to the transgression, there was a low water stand during which peat was formed in several depressions.

Aeolian activity

In Estonia, where both inland and coastal dunes are encountered, the aeolian redistribution of fine aqueoglacial and beach material was highly controlled by land uplift, palaeoclimatic parameters, including wind direction and activity, soil moisture, but also by the grain-size of initial sediments. Due to the limited supply of sand, concentration of heavy storms to autumn and winter periods and high precipitation rate, dunes are relatively low (mainly 5-15 m) and rendered stationary by vegetation. At present, there is practically no dune sand movement in Estonia.

On several occasions in times past, the aeolian sands posed a great threat to inhabitants and the environment. At the end of the 19th century, a mobile dune on the Island of Saaremaa endangered the Kärla church, pastor's mansion and farmsteads in the vicinity. The advance of the dune was stopped by a pine stand (Tiismann 1924). The same author describes the movement of sand in the surroundings of the Ristna lighthouse where the strong southwesterly winds picked up a mass of sand and deposited it behind the doors. Every time the islanders had to work days to cart the sand off and clear the access to the houses.

The intensification of agriculture and large-scale land improvement in the 1950s-60s brought about huge fields with a steppe-like appearance and caused deflation on sandy and peat soils. For instance, during April 29 - May 11, 1974, some 16.2 tonnes of dry soil per hectare was carried away by wind action from the Apometsa fields of the former Ranna State Farm in the present-day Harju County. In the same year, during May 1-7, the cloud of dust, blown up from the fields of the Paluküla and Tubala villages on Hiiumaa Island, reached a height of a few tens of metres and deteriorated traffic conditions on the road. A layer of sand, up to 30 cm in thickness, deposited on the road; on the roadside its thickness reached 75 cm (Kees 1992).

Particles of silty soil and dry peat start moving already when the wind speed is as low as 3-4 m/sec. With the wind speed of about 15 m/sec, the amount of soil set into motion generates already a surface dust storm (Photo 58). In Estonia, the wind speed may reach 40 m/sec at a height of 10-20 m above the ground. If the wind blows with a speed of 35 m/sec during several minutes, it may cause great damage to agriculture.

There are some 200,000 hectares of land endangered by deflation in Estonia. On Hiiumaa Island, such fields make up two thirds of the arable lands. When the danger was understood, fields larger than 50-60 hectares on the lowlands and 20-30 hectares on the elevations were prohibited. Belts of trees were planted to shelter fields, while gentle peat soils were planned for a long-term use as grasslands. As a result of the liquidation of state large-scale agriculture in Estonia, the area of fields endangered by deflation has essentially decreased.

In Estonia, continental dunes occur in the Iisaku-Illuka area (Rähni 1959). These parabolic and transverse formations, 0.8-2.7 km long and up to 15-20 m high, indicate a westerly-northwesterly palaeowind direction (Zeeberg 1993). The west and northwest, windward slopes of the dunes are gently sloping (3-18°), while the opposite, leeward slopes are much steeper (18-24°). Small coversand hillocks occur on top and on the slopes of dunes.

Most probably, these coversands, dunes and drift sands originate from the Younger Dryas and the beginning of the Pre-Boreal when a significant regression of Lake Peipsi took place in the Alutaguse Lowland, and the so-called Small Peipsi was formed (Раукас и Ряхни 1969). The sand and silt material in the area was formed locally from glaciofluvial and glaciolacustrine deposits reworked by the waves of ancient Lake Peipsi. Dunes were commenced here immediately after source deposits became available, and the process stopped when they became overgrown with vegetation.

The inland dune areas in northeastern Estonia are covered with pine forests and surrounded by bogs. The maximum ages of the dunes may be inferred from the sediments or soils from which they started to develop. These, mostly glaciolacustrine plains and kame fields, formed some 12,200 years BP (Раукас и др. 1971). However, the sand became available for redeposition not until it had drained and dried, possibly after the retreat of the glacier of the Palivere Stadial some 11,000 yr BP.

Unfortunately, this cannot be checked by palynological sampling of dune sands or dating of organic remains. In 1988, attempts were made to solve the problem by means of the TL-method (Раукас и др. 1988), however, all the dates obtained suggested much younger ages, between 4000 - 7100 yr BP.

Coastal dunes occur in Lower Estonia where sandy-silty sediments were available for aeolian processes. The largest dunes, up to 20-25 m in height, formed during the transgressive phases of the Baltic Ice Lake, Ancylus Lake and Litorina Sea (Eltermann & Raukas 1966). Rising sea levels brought about shoreline erosion, destruction of fore-dune and beach ridge vegetation, and initiation of transgressive dunes (Cooper 1958). The most prominent dunes associated with the Baltic Ice Lake transgressive shoreline are located in the Lahemaa National Park and on the Kõpu Peninsula on Hiiumaa Island. The dunes of the Ancylus Lake occur on the West-Saaremaa Elevation, on the Tõstamaa and Kõpu peninsulas and near Häädemeeste. The dunes related to the Litorina Sea transgressive shoreline are found at Sininõmme, Rannametsa and Tõstamaa. The dunes are most numerous on west-facing shores, where the prevailing winds are westerlies and south-westerlies (Fig. 189).

As a result of the uplift of the Earth's crust, coastal dunes are nowadays situated at some distance from the contemporary shore and at different heights above sea level (Eltermann & Raukas 1966, Martin 1988). At the present seashore only low, a few metres high fore-dune ridges occur, *e.g.* at Kloogarand, Narva-Jõesuu and Valgerand. The biggest contemporary dunes with specific morphology, termed "basket-trap" dunes by Orviku (1933b), are located on the north coast of Lake Peipsi (Fig. 189). Some small dunes occur also around Lake Võrtsjärv (Tavast *et al.* 1983).

Estonian ridge-like coastal dunes, the length of which ranges from 50 to 150m and the width from 20 to 50 m, are elongated in the direction of the prevailing winds (Fig. 190). They usually parallel each other in echelon-like series (Orviku 1933b). The length/width ratio is mostly 2:1 - 4:1 (Fig. 190). The windward slope (5-20°) is often scattered with small, secondary coversand hillocks. The leeward slope is steep (25-40°) and even. Occasionally, *e.g.* in the Litorina Sea dunes at Rannametsa (Photo 59), both the wind- and leeward slopes are almost equally abrupt (25-45° and 25-30°, respectively).

Photo 58. Sand storm on the Pirita beach, Tallinn. *Photo by Karl Orviku.*

Photo 59. The largest dunes in Estonia were formed during the transgressive phases of the Baltic Sea. Dune ridge at Rannametsa (SW Estonia) with equally steep wind- and leeward slopes. *Photo by A. Raukas.*

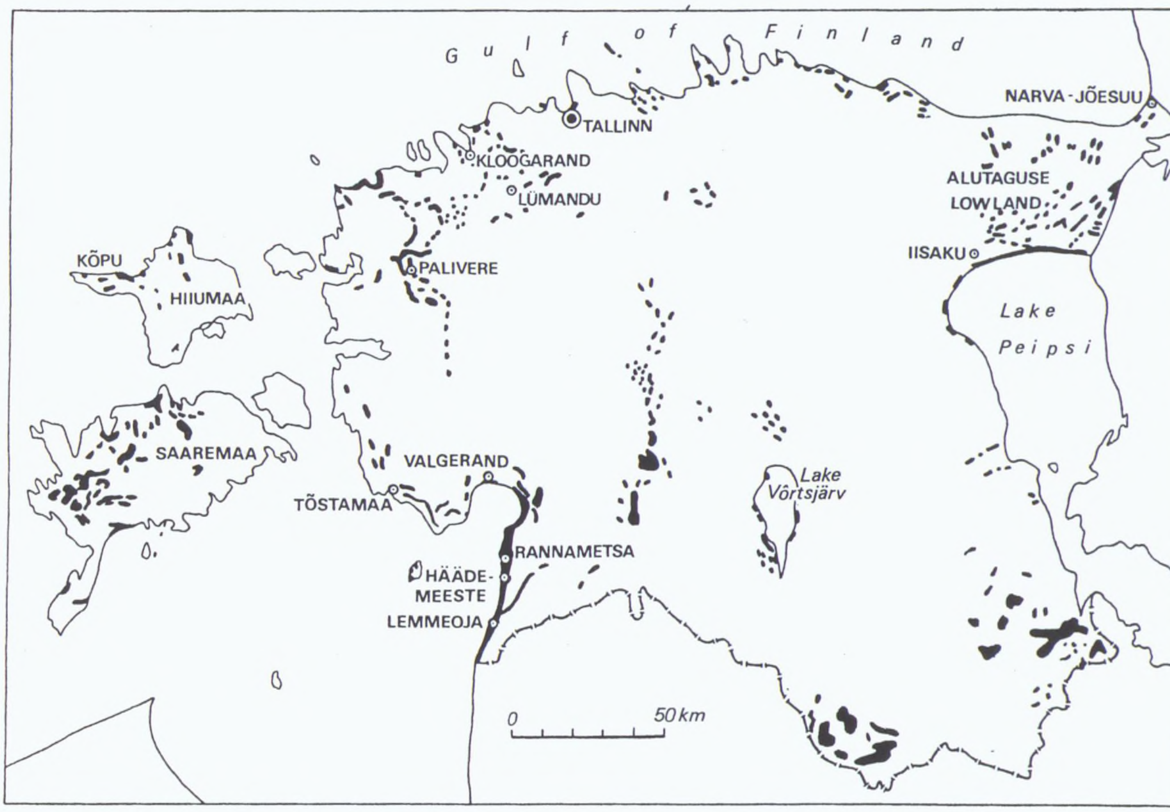


Fig. 189. The biggest dunes and main areas of aeolian sand.

This may be due to the esker buried under aeolian deposits (Raukas 1988) or the later marine erosion of the windward slope during the stages of high water level (Eltermann & Raukas 1966, Martin 1988). In the areas with variable winds, as for example on the Kõpu Peninsula, a clear orientation of dunes is absent or they are oriented in two or three different directions (Eltermann & Raukas 1966).

As a result of uneven migration of dunes, during which the central ridge blowouts moved downwind and the low-lying arms fixed with vegetation lag behind, parabolic dunes came into being. In such dunes the windward slope has an inclination of 10-20° and the leeward slope 20-30°. Both, the ridge of parabolic dunes and the fields made up of such dunes, contain multiple reaping-hook segments. Erosional forms in dune fields are only some metres deep.

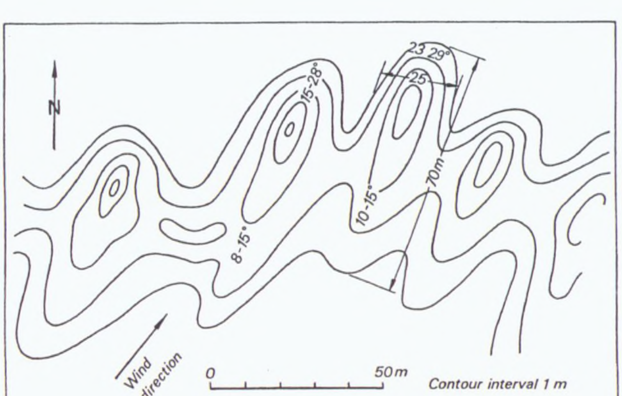


Fig. 190. Ridge-like dunes usually parallel one another in echelon-like series. The Aavasoo Dune Field of the Ancylus transgressive shoreline (after Eltermann & Raukas 1966).

During the invasion of the sea, the migration of the deposits up to the coastal slope resulted in their mechanical and mineralogical differentiation which, in its turn, brought about accumulation of well-sorted sand on the shore (Раукас 1966). In Estonian dunes, fine-grained (0.1-0.25 mm) sand prevails. In 25.4 per cent of analyses (Raukas 1968) it formed 80%, and in 9 per cent of analyses even 90 % of the dune material. Medium-grained (0.25-0.5 mm) sand predominates in 18 per cent of the analyses, whereas in about 10 per cent of cases the coastal dunes contain more or less equal amounts of both fine-grained and medium-grained sand, or fine-grained sand and coarse-grained (0.05-0.1 mm) silt. The size of sand grains in dunes depends mainly on the composition of the initial deposits. Compared to the initial deposits, the dune-sands are much better sorted and lower in both coarse and fine fractions (Raukas 1968).

The aeolian sands of Estonia are mostly bimineral and less frequently oligomineral, consisting mainly of quartz and feldspars (Raukas 1968). In northern Estonia and on the islands of the West-Estonian Archipelago, where the initial deposits are rich in carbonates, in some places the dune-sands contain, next to quartz and feldspars, also great amounts of carbonates (at Palivere 11.2 - 26.2 %, in the Püha Andreas Dune Field on the Kõpu Peninsula 2.2 - 13.1 %, at Lümanda 2.0 - 6.8 %). In comparison with the initial deposits, the aeolian sands are somewhat richer in heavy and more or less isometric and weathering-resistant minerals (quartz, zircon, *etc.*). The content of lamellate and tabular minerals (especially micas and chlorites), and the minerals liable to weathering and wear (carbonates, feldspars, *etc.*) is lower. The aeolian sands contain more SiO_2, and less Al_2O_3 and K_2O than the initial deposits. This fact agrees with the results of mineralogical analyses (Raukas 1968).

According to the interior structure, Estonian dunes may be divided into several types (Raukas 1968). The structure of dunes differs in longitudinal and cross-sections. In cross-sections the sets of lamina are usually more or less symmetrical in relation to both slopes. According to the classification of Botvinkina (Ботвинкина 1965), a wedge-like inclined stratification predominates. Concavo-convex wave-like sets of lamina are rarer and they mostly occur in longitudinal sections (Fig. 191).

Depending on their closeness to the ancient shorelines, the Estonian coastal dunes are traditionally classified as the Ancylus transgressive shoreline dunes, Litorina dunes, *etc.*, however, they may have been repeatedly reblown and rather young in age. Forest cuttings and fires, military actions (Kroodi, Värska) and other kinds of human activities may trigger the movement of surficial sand. The soil profile, occasionally several metres thick (*e.g.* Lemmeoja), shows that at least some dunes became overgrown with vegetation immediately after the deposition and never moved again. Wet environmental conditions, sparse population and rapid spread of vegetation prevented extensive redistribution of loose sandy sediments by wind in Estonia.

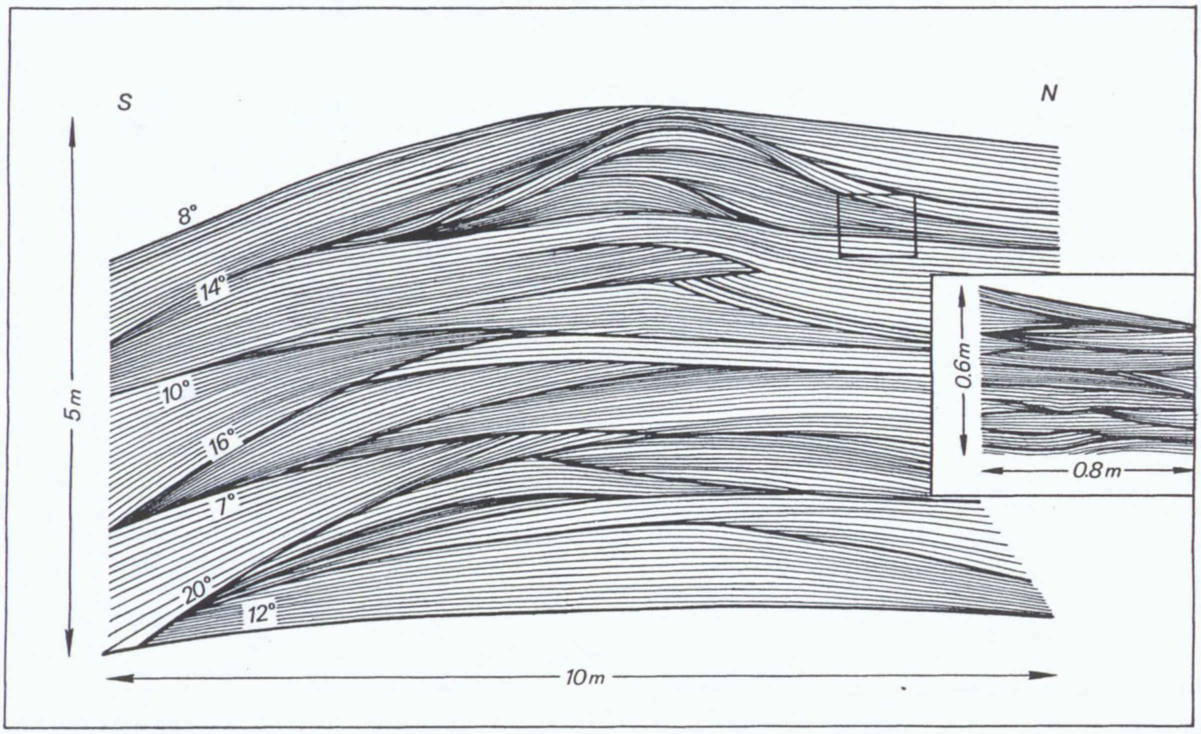


Fig. 191. Litorina dune at Sininõmme displays a typical lamination in the longitudinal direction (after Raukas 1968).

Genesis and development of mires

Introduction

Like in other lowland countries, the surface structure and climate in Estonia favoured the formation and expansion of mires. Estonia has a rather dense network of rivers and a high proportion of lakes. Per 100 km^2, there are about 22.8 km of rivers, about 50 km of streams and channels and 3 lakes with a mean surface area of 1.1 km^2. The climate varies from submaritime in the western coastal region to subcontinental in the easternmost region. Average annual precipitation is in the range of 500-700 mm. Mean temperature in July is 16.5° to 17.5° C, in February -4.0° to -7.5° C.

Peatland is a term applied to any peat covered area ameliorated or untouched, mire is a general term for any peat covered area in virgin state. Mires are divided into two basic types: ombrotrophic (raised) bogs which are totally fed by atmospheric precipitation, and minerotrophic fens with additional feeding by ground and surface waters. Peatlands cover about 1,009,101 ha, *i.e.* 22.3 % of Estonia's land surface (Orru 1992).

Estonia belongs to the hummock-ridge bog region. The boundary between the Baltic Coast Bog Province and the East Baltic Bog Province divides its territory into two parts (Botch & Masing 1983). The main phytogeographical boundary crossing Estonia in a SSW-NNE direction (Laasimer 1965) corresponds well with the geomorphological differences that follow the maximum transgression limits of the Baltic Sea. The western part, the so-called Lower Estonia, was covered by the Baltic Ice Lake and Holocene stages of the Baltic Sea, while in the eastern part or Upper Estonia local periglacial lakes lingered briefly (Fig. 192).

Following the classification by Masing (1975, 1988a), **minerotrophic mires** are subdivided into soligeneous, topogeneous, limnogenous and transitional. The water in soligeneous or spring fens is commonly calcium-rich, sometimes with a very high Ca content. *Schoenus nigricans* and *Juncus subnodulosus* communities are found in such calcium-rich spring fens in the Island of Saaremaa.

Extremely rich fens are predominantly distributed on the

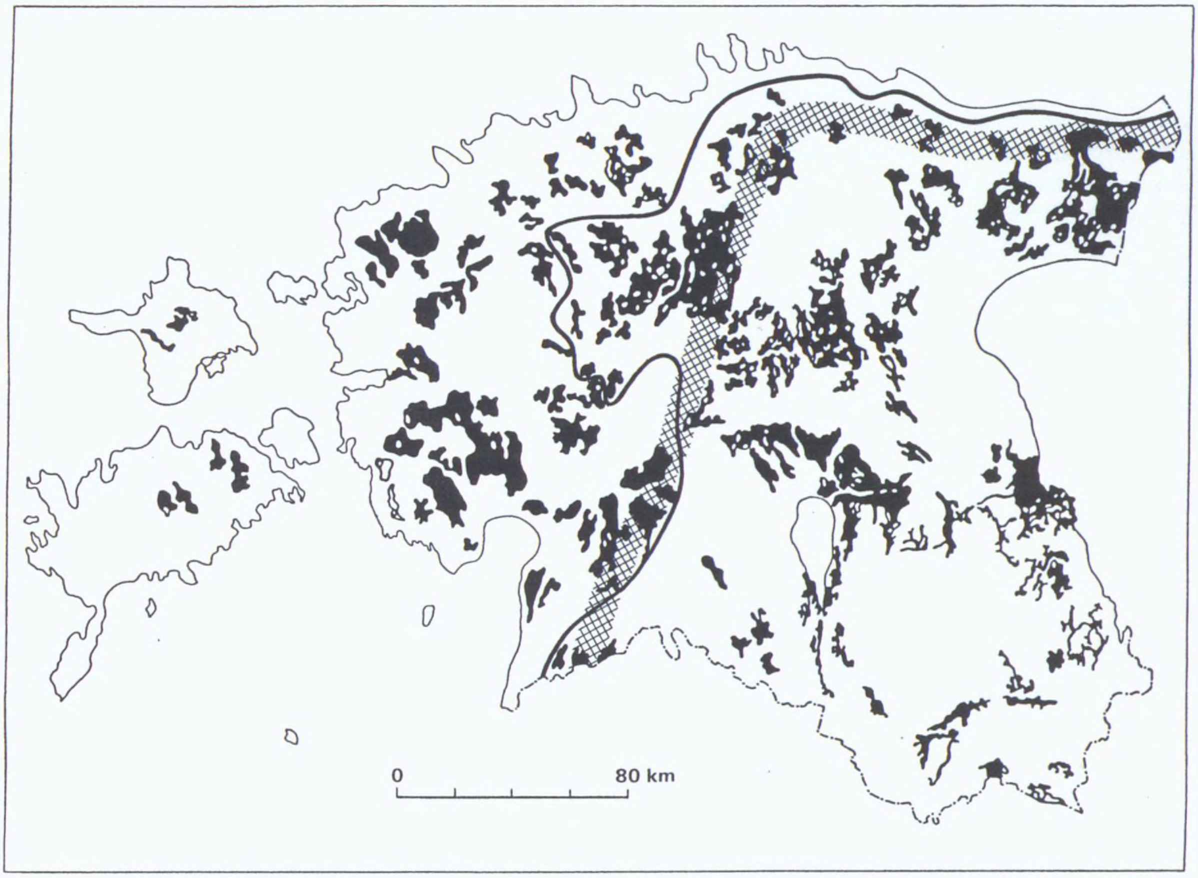


Fig. 192. Distribution pattern of mires larger than 1000 ha in Estonia (after Allikvee & Ilomets 1995, with complements). Solid line indicates the maximum distribution limit of the Baltic Ice Lake B_3 and divides Estonia's territory into Lower and Upper Estonia. The cross-striped belt follows the approximate boundary between the East- and West-Estonian geobotanical provinces (after Laasimer 1965) and corresponds well with the distributional pattern of the West-Estonian type of plateau bogs and the East-Estonian type of convex bogs.

carbonate-rich substrates of western Estonia. *Myrica gale, Schoenus ferrugineus* and *Cladium mariscus* fens occur in the western part of Saaremaa Island and in some places on the west coast of mainland Estonia. Rich, particularly tall *(Carex acuta - C. elata* ass.) and low sedge fens *(Carex nigra - C. panicea* ass.) are more common in the eastern part of the country.

Flood-plain fens are related to the South-Estonian rivers where ground water plays an important role.

Poor fens (transitional or mixotrophic mires) are divided into transitional fens and wooded transitional bogs. Different categories of transitional fens are distinguished. Sedge-moss fens *(Carex lasiocarpa* and *C. lasiocarpa - C. rostrata - C. limosa* ass.) occur mostly on floodplains around lakes in western and central Estonia; elsewhere they are rather rare. *Myrica-Schoenus* moss fens with *Sphagnum* patches on calcium-rich substrate are spread in western Estonia. Wooded transitional bogs often form a belt around large ombrotrophic bogs, particularly in northern Estonia.

Ombrotrophic mires are divided into moors and bogs. Moor heaths on thin peat and underlying pure sand occur in depressions between dunes in the western coastal area of the mainland and on western Estonian islands (particularly on Hiiumaa). On the basis of the density of tree canopy, two types of bogs are commonly distinguished: wooded bogs and open bogs. The bog margins are usually covered with bog forests. A bog in its early stage of development may be entirely covered with pine forest. Unpatterned, open or wooded bogs dominate in western Estonia, while patterned bogs with strings, hummocks and pools are common in the eastern part of the country.

Regionally, on the basis of differences in the vegetation and mire-complex types, Estonian bogs are divided between East-Estonian convex and West-Estonian plateau bogs. According to Allikvee & Masing (1988), there are 8 mire districts and several subdistricts in Estonia.

Mire initiation

The terrestrialization of water basins, mainly lakes, as a result of infilling (Photo 60) and the paludification of mineral soils are two alternative ways in mire initialization. In all likelihood, the mire formation is controlled by hydrothermal conditions. So, the terrestrialization should be dominating there where the water table is dropping as a result of decreasing humidity or increasing evapotranspiration. In the case of increasing precipitation and cooling climate the paludification processes ought to have some advantages. Naturally, there are several other factors, such as lithology, geology, neotectonics, giving also an impulse to the mire formation. However, on such a rather uniform and compact territory like Estonia, the factors other than climate and, to a certain extent, also the neotectonics, can be eliminated as having been stable throughout the Holocene. Here we have to consider that the larger the data set in use, the higher the objectiveness of the results.

With this in mind, for the purpose of mire formation stud-

Photo 60. About 25 per cent of mires in Estonia are of limnogenous origin and infilling of lakes is in progress. *Photo by R. Karukäpp.*

ies, stratigraphical sections of the deepest points of the mires, larger than 300 ha, were selected from the peat resources inventarization data of the Peat Group of the Estonian Geological Survey. Of the 467 mires under consideration, 345 are ombrotrophic bogs (Table 48). The distribution of mires between Lower and Upper Estonia is uneven - 35 and 65%, respectively. In both cases, some 25% of mires are limnic in origin.

Peat samples from 35 Estonian mires were dated in the radiocarbon laboratories of the Institute of Zoology and Botany (Tartu) and Institute of Geology (Tallinn). Of the total of 500 datings, about 240, which underlie the present calculation of peat increment values (Ilomets 1995), were obtained on the stratigraphically well-studied cores. As the data are rather unevenly distributed between different peat types, *e.g.* 65% of samples are characterizing the ombrotrophic *Sphagnum* dominated peats, the average peat increment values may be fitted for bog peats rather than for minerotrophic fen peats. It is supposed that the calculated mean values may characterize the growth rate of the peat in all 467 mires studied. Following the arrangement and thickness of peat types in the stratigraphical sections, the likely age of every peat deposit was calculated. The calculations yielded an age of *ca.* 10,000 yr BP for the peat-mineral subsurface contact for several mires in Upper Estonia. The oldest peat sample, collected from the Vaskna fen, gave a radiocarbon age of 9930 ± 70 BP; TA-1600.

The first marked increase in **paludification** (Fig. 193) in Estonia took place about 8500 yr BP. During the next paludification interval around 7100 - 6100 yr BP, about 13% of mires were formed. This interval of intensive paludification can well be correlated with the transgression phase of the Litorina Sea (7000 - 6800 BP, Kessel & Punning 1984). In Upper Estonia, about 24% of topogeneous mires started to develop about 5100 - 4100 yr BP and ca. 13% between 3500 and 2700 yr BP. In Lower Estonia, there were several periods with not a very clearly defined intensification of paludification processes: 6800 - 6400, 4500 - 3700 and 2700 - 2000 yr BP.

The data indicate that the first increase in **terrestrialization** (Fig. 194) started some 2000 years later than the first paludification rise up (Fig. 193). Most of limnogeneous mires (67%) in Upper Estonia were formed within a rather limited interval during the second half of the Atlantic Chronozone between 6500 and 4500 yr BP. About one third of topogeneous mires came into being 5400 - 3800 yr BP. In Upper Estonia, this was the most powerful paludification interval during the Holocene. In this part of Estonia, intensive mire formation started evidently some 700 years earlier and completed ca. 500 years before it commenced in Lower Estonia. Although the glacioisostatic uplift in Estonia reached its zero level about 5000 yr BP (Punning 1985), the terrestrialization frequency increased in Lower Estonia at about 5500-5000 and 4500 - 4000 yr BP. It should be pointed out that in the Holocene, there was only one period - around 4500 yr BP, when the processes of paludification and terrestrialization increased simultaneously in Lower and Upper Estonia (Fig. 193). The plant species composition (mostly *Bryales* and *Carex* species) of initial communities, as reflected in the lowermost peat layers, indicates that the paludification of forests might have been of no importance, but in most cases the peat formation started in the lagoons during the regressive phases of the Litorina Sea development. Under the conditions of slow water-level decrease, the shallow fresh-water

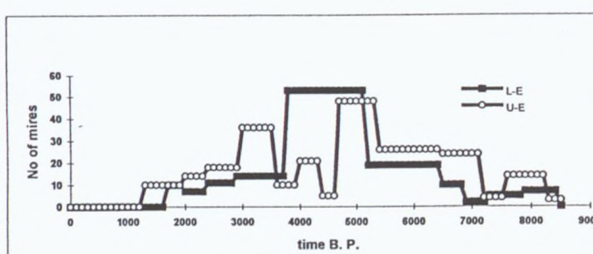


Fig. 193. Temporal changes in mire initialisation by paludification. Symbols: L-E - Lower Estonia, U-E - Upper Estonia.

Table 48. Distribution pattern of Estonian mires

	Number of mires		Limnogeneous		Topogeneous	
	Total	Bogs	Total	Bogs	Total	Bogs
By amount						
Lower Estonia	164	151	38	36	126	115
Upper Estonia	303	194	69	41	234	153
Total	467	345	107	77	360	268
In percentage						
Lower Estonia	35	44	23	24	77	76
Upper Estonia	65	56	23	21	77	79
Total	100	100	23	22	77	78
Bogs (% of total mires)						
Lower Estonia	85		79		87	
Upper Estonia	60		52		62	

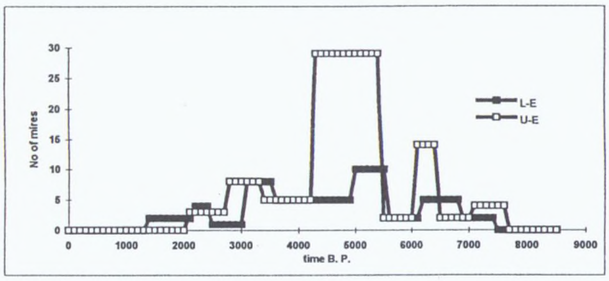


Fig. 194. Temporal changes in mire initialisation by terrestrialisation. Symbols as in Fig. 193.

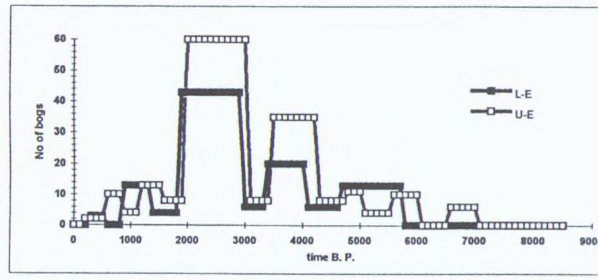


Fig. 195. Temporal dynamics of raised bog initialisation. Symbols as in Fig. 193.

lagoons, which had isolated from the sea, were filled up with peat. However, in some cases, a thin (ca 2 - 5 cm) layer of gyttja accumulated there. Most probably, there was no paludification increase in Lower Estonia and all the mires, which formed during that interval, are limnic in origin.

More than 30% of mires in Upper Estonia started to develop during the time span from 2500 to 6500-4000 yr BP. Of those mires, about 2/3 are of limnic origin. In all likelihood, this is a result of the integrated effect of several factors, favouring mire formation. One of the reasons of continuous infilling of lakes may be related to the stagnation or even decrease of the water level as a result of deterioration of climatic conditions. Digerfeldt (1988) has shown that in southern Sweden the major decrease of water level in lakes occurred 6500-4600 yr BP.

Most of Upper Estonia was freed from the ice in the Pandivere Stadial about 12,000 yr BP (see Ch.VIII.8). It must have taken at least some 5000 years, before a lake became infilled with mud and the process of terrestrialization could start. The longevity of the interval also indicates that if the infilling had already reached a certain level, then even the water-level rise after the stop of glacioisostaic uplift some 5200 yr BP, was incapable of stopping the terrestrialization processes.

Mires reached the ombrotrophic stage of development at well distinguished intervals and synchronously in Upper and Lower Estonia (Fig. 195). The oldest bogs in Estonia date from about 8000 yr BP. However, there are three intervals of intensive mire formation which started 7000, 6000 and 5000 yr BP and lasted for about 500 years each. Quite a small part of our raised bogs was formed during those intervals. It is worth of mentioning that the two first intervals indicate the beginning of bog stage mostly in Upper Estonia, while in Lower Estonia the bog formation intensified only 5000 yr BP. The most intensive bog formation started ca 4000 yr BP and during the succeeding 2000 years (up to 2000 BP) more than 40% of the bogs under consideration came into being. This time span can be divided into two periods: 4000 - 3400 and 3000 - 2000 yr BP. During the first period, bogs developed at equal rates in the two parts of Estonia, while during the second period the process was more intensive in Lower Estonia. The last intensive bog formation interval when more than 30 mires, today larger than 300 ha, reached the raised bog stage showed marked differences between Lower and Upper Estonia. Anyway, the calculations indicate that in Upper Estonia some ten bogs were formed during rather restricted time spans 1500-1200 and 700-500 yr BP, but in Lower Estonia it happened only between 1200 and 800 yr BP.

Although mires may reach the raised bog stage at different times, the temporal dynamics of bog stage initialisation shows the importance of external factors giving impact to these processes. These factors ought to be different from those influencing the temporal dynamics of mire initialisation as the latter process cannot be well correlated with the temporal dynamics of mire origin and as mires are developing simultaneously all over Estonia.

As mentioned above, 40% of the bogs under consideration developed during the course of *ca.* 2000 years (4000 - 2000 BP). In all likelihood, this phenomenon results, at least partly, from autogenic succession. A mire may reach the ombrotrophic stage after a certain time span, most importantly determined by the combination of geomorphological, water quality (supply of mineral components) and peat accumulation rate peculiarities. The research data indicates that the time span needed is at least 1000 years, commonly 2000 - 2500 years. Therefore, by autogenic reasons the expansion of raised bogs and, consequently *Sphagnum* could not start earlier than 4000 yr BP, as most of mires developed between 6500 and 4000 yr BP.

Development

In view of the different origin, the development of limnogeneous and topogeneous mires is analysed separately (Ilomets 1992).

The most probable sequences of **limnogeneous mires** are:

1. In the case of deeper lakes and thicker lake sediments: *Phragmites* stand → *Betula-Phragmites* fen → *Betula-Picea-Carex* forested fen → *Carex* open fen → *Carex-Sphagnum* poor fen → hollow-ridge complex with *Eriophorum vaginatum* and *Sphagnum* species → pool-ridge complex *with S. fuscum* on hummocks

2. In the case of shallow water bodies with a thick layer of lake sediments the two most probable pathways are:

2.1. *Phragmites* stand → *Betula-Phragmites* fen → *Betula-Picea* fen forest → *Betula-Picea-Pinus* fen forest → *Pinus* bog forest → hollow-ridge complex with *Calluna vulgaris* and *Sphagnum* → pool-ridge complex with *S. fuscum* on hummocks.

2.2. Bryales quaqing mat → *Carex-Bryales* quaqing fen → *Carex* open fen → *Carex-Sphagnum* poor fen → *Sphagnum* poor fen → hollow-ridge complex with *Eriophorum vaginatum* and *Sphagnum* species → pool-ridge complex with *S. fuscum* on hummocks.

There is a high probability that in the course of the limnogeneous development all the communities reach the *Carex* dominated fen stage (Fig. 196) and thereafter the ombrotrophic stage over the mixotrophic *Carex/Sphagnum* community.

So, it may be concluded that the type of origin may play a rather important role in the developmental pattern of a mire. Most surprisingly, in very many cases this phenomenon may affect the developmental peculiarities in the ombrotrophic bog stage. Thus, for topogeneous mires *Sphagnum magellanicum* dominated lawn communities are not holding a very important position between the pine forest state and the *S. fuscum* dominated hummock-ridge state. In the case of limnogeneous mires the same *S. magellanicum* lawn communities keep a central position between *Eriophorum* dominated lawn and *S. fuscum* dominated hummock-ridge communities.

The formation of **topogeneous mires** as a result of paludification may have started with the formation of at least 15 different peat types. The most probable sequences of peat types are:

1. In the case of paludification of forested areas:

Betula-Pinus herb rich fen forest → *Betula-Phragmites* forested rich fen → *Betula-Picea* wooded rich fen → *Picea-Betula-Pinus* poor fen forest → *Pinus* bog forest → hummock-ridge complex with *Sphagnum magellanicum* → hollow (pool)-ridge complex with *S. fuscum* on hummocks.

2. In the case of open (treeless) depressions, temporarily overflooded:

Bryales mat or *Sphagnum* carpet → *Carex* open rich fen → *Carex* poor fen → *Carex-Sphagnum* poor fen → *Sphagnum* poor fen → hollow-ridge complex with *Eriophorum vaginatum* and *Sphagnum* species → pool-ridge complex with *S. fuscum* on hummocks (Fig. 197).

The two sequences are converging into the same point - a patterned *Sphagnum* dominated bog. The first case emphasizes the paludification of forests and indicates that forested mires can develop up to the ombrotrophic bog stage as mire forests. The second case illustrates the development of open areas and leads to minerotrophic *Carex* dominated open fen sites. Further development can be characterized by a gradual rise of the role of *Sphagnum* species.

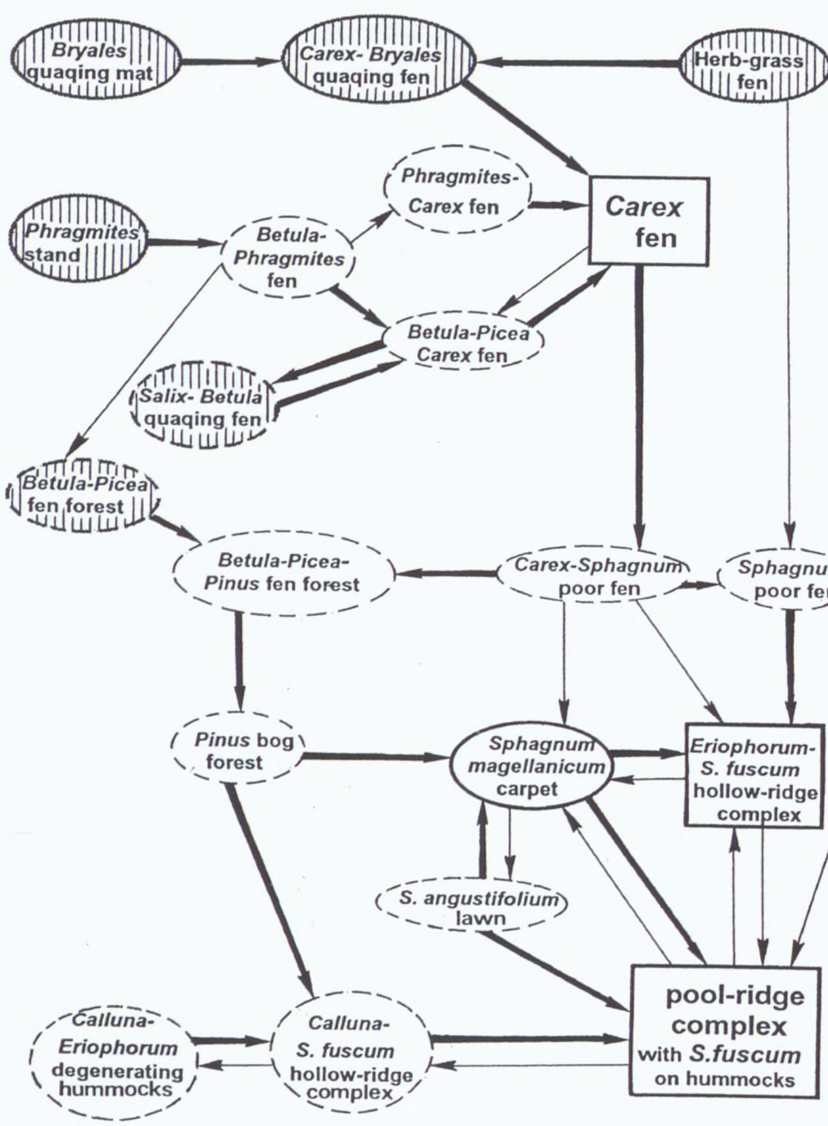


Fig. 196. Developmental pattern of limnogeneous mires in Estonia. Arrows: thick line - more common sequences, thin line - less common sequences. Ellipsoids: striped - initial states, white - serial states, continuous line - most common states, broken line - less common states. Rectangulas - climax states.

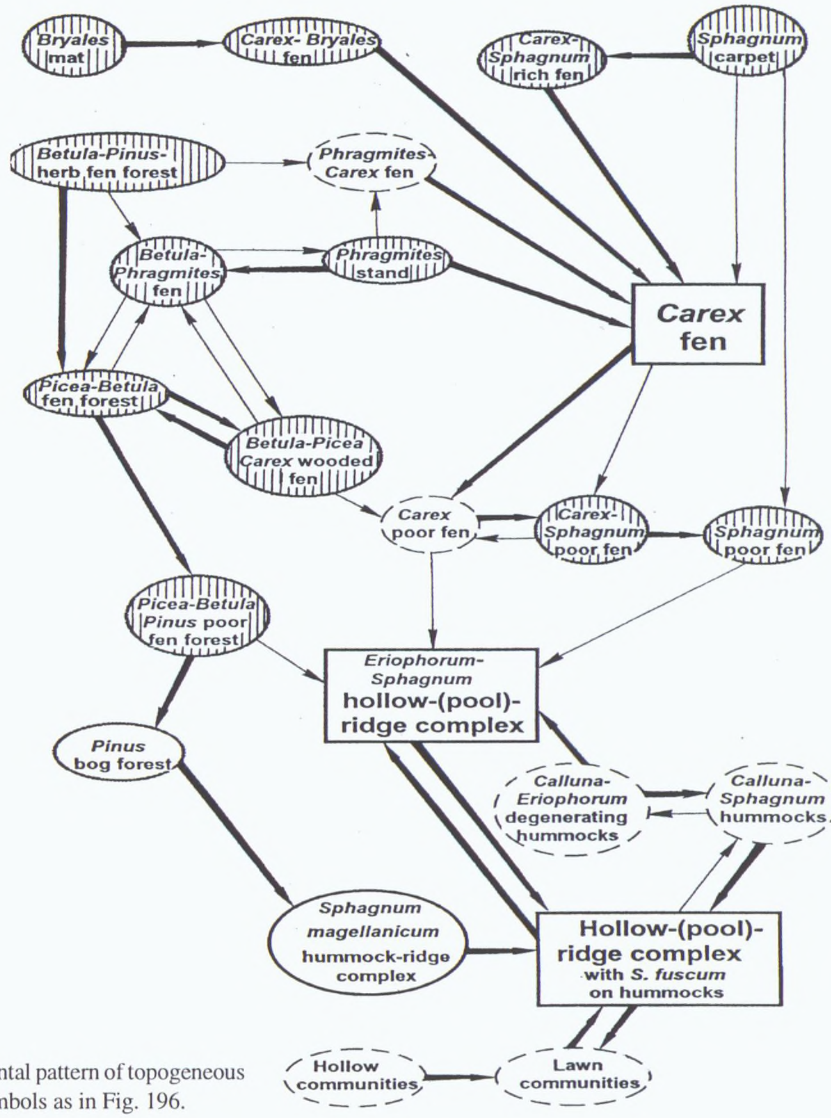


Fig. 197. Developmental pattern of topogeneous mires in Estonia. Symbols as in Fig. 196.

SOIL FORMATION

Pre-Pleistocene pedogenesis

Although the territory of Estonia became dry land already at the end of the Devonian, there is no direct evidence concerning the formation of soils in the Pleistocene, to say nothing of the more ancient Neogene, Palaeogene and entire Mesozoic era. The soils formed during these times have either been completely destroyed by later geological events or transformed beyond recognition by glacial erosion. Therefore, the reconstruction of soil history must be based on the contemporary analogies, proceeding from the theoretical assumption that the plant, faunal and microbal organic matter as well as its derivates are always and everywhere recognized as the formative power of pedogenesis, and soils are considered as the result of and a pre-requisite for the production processes (Reintam 1978).

In the Late Neogene, coniferous and deciduous forests were spread in Central and North Europe, the animal life was represented by the species of herbivores and carnivores, much the same as today (Raukas 1987). This means that the herbaceous ground vegetation in forests, and perhaps even grasslands in localities of thin forests and/or treeless plains had to vouch for the existence of pasture food chain as well as of energy fluxes and turnover of substances there. There is no information as to the thickness of weathering crust on Ordovician and Silurian calcareous deposits and Devonian sandstones with clayey interlayers. Presumably, it was extremely thin or entirely lacking. It is possible that biological weathering and pedogenesis on limestones and dolomites under the herbaceous forest or grassland vegetation were developed by the way of argillization and humus accumulation analogously to those of Holocene origin.

The contemporary calcareous tills in the North-Estonian Plateau and in central Estonia contain 8.8-11.4% and 11.5-13.1% of clay-size (<0.001 mm) particles, respectively. Fine earth in unchanged tills could not result from the mechanical crushing by the moving glacier. It was probably formed as a result of Pre-Pleistocene weathering and pedogenesis. The same conclusion is possible for the organic carbon, the small amount of which (0.1–0.2%) is always characteristic of unchanged tills and cannot represent the phenomenon of modern pedogenesis. The Pre-Pleistocene soils were destroyed by glaciers. However, the organic matter, clay or any secondary product of chemical and biological weathering, accumulated in these soils, did not disapear but, mixed with pure mineral matter, participated in the formation of tills.

Under the herbaceous sites, on the Ordovician and Silurian limestones as well as on their desintegrated weathering products, there were excellent preconditions for the intensive humus accumulation and rendzina formation. Most likely, the Pre-Pleistocenic Rendzic Leptosols* were characterized by the variable thickness, stony debricity of active section and different productive capacities. It is possible that the progress of argillization of dolomitized limestones and overlying terrigenous sediments resulted in the development of cambic properties and the formation of Pre-Pleistocenic Cambisols. The latter could have occurred besides Rendzic Leptosols, on limestones only there where the weathering rate was as high as in the Holocene and exceeded two grams clay per sq.m per year (Reintam 1975). An analogy to that may be also assumed on Upper-Devonian limestones, though on Ordovician and Silurian limestones the pedogenesis is some millennia older.

Quite a different scenario applies to the Middle-Devonian (*Old Red*) sandstones during the Pre-Pleistocene. Presumably, the bulk density of their upper strata was smaller and water permeability greater than in the topmost part of the contemporary sandstones underneath the thick cover of Quaternary deposits. Base unsaturation and slight coatings of sesquioxides on the grains of quartz are typical of Devonian sandstones. These chemical relationships, accompanied by the relatively high water permeability and low moisture capacity, had to result in poorer phytocoenoses than those on calcareous deposits. The eluviation of solum, podzolization and formation of Podzols were possibly analogous to some contemporary podzolic (spodic) formations on sands under the stands of *Eucalyptus* or *Agathis*, or both, in humid regions of Africa, South-Eastern Asia and Australia (Dudal 1990, Eswaran *et al.* 1995). Bearing in mind the contemporary analogies, podzolization and formation of humus-illuvial (carbic) podzolic sections in Pre-Pleistocene time could have been sporadic. Sandy Ferralic Cambisols and Arenosols could have occurred under more favourable mineralogical or phytocoenotical conditions.

It is difficult to suppose that the lessivage and formation of Luvisols took place on the Devonian sands, but their possible occurrence on clays and different clayey products of the weathering of calcareous strata cannot be excluded. Undoubtedly, the phenomena of surface, perched and ground gleyization could have also been involved in the Pre-Pleistocene pedogenesis. In spite of eventual high level of ground water table in limestones and dolomites, their extremely poor capillarity hardly favoured the realization of ground gleyization there. The progress of the latter could have been plausible on Devonian sands which in places contain different ferrous relics up today.

* Soil identification here and further by FAO-UNESCO (1985, 1994)

Holocene pedogenesis

In the Late Pleistocene and Holocene, the pedogenesis is related to climatic changes, alteration of forests and the evolution of the Baltic Sea. Although direct data on the temporal and spatial changes in soils during that time have been received in the course of the last decades or centuries, the final product in the kind of ped sections is still assumed to reflect soil's past and present, memory and moment everywhere (Соколов и Таргульян 1976). World experience and information on the regularities in any interaction within different plant-soil systems enable the reconstruction of trends and nature of soil formation and evolution in dependence on climatic, lithologic and phytocoenotic situation (Fig.198). Without the highly informative palaeobotanic evidence, reconstructions of this kind would be impossible.

According to pollen evidence, after the retreat of the last glacier a tundra-like vegetation appeared in Upper (Watershed) Estonia (Thomson 1929, Lippmaa 1935, Laasimer 1965). Proglacial lakes and permafrost tilly landscapes occurred in front of the retreating glacier. Without doubt, the presence of water-saturated fine earth in tills was an immediate prerequisite for the permafrost formation there. It is hardly possible that permafrost could have formed in limestones and in local till in which coarse fraction formed more than 75%. That is why simultaneously with the tundra situation at the time of ice lakes, arctic steppe (meadow) associations expanded on unfrozen tills, poor in coarse fractions, in central and southern Estonia (Гричук и Гричук 1950). Textural peculiarities of Late Pleistocene deposits and the contemporary situation in Russian and Canadian permafrost areas (Tedrow 1977, Морозова 1981) suggest that cryoxerophilic meadows on (Gelic?) Leptosols and even on Kastanozem-like formations could have occurred on limestone outcrops (alvars), North-Estonian stony till, drumlins and end moraines with an extremely coarse texture, at least in the core.

The permafrost weathering in combination with surface gleyization became, probably, prevalent under the tundra vegetation, especially in the upper section of the Quaternary stratum which had seasonally been subject to the alternating frozening and thawing processes. Besides the slight textural differentiation, an eluviation of fine-dispersed particles tends to be also characteristic of contemporary Gelic Gleysols in tundra. If to presume that soils on tills in the Late-Pleistocene tundra were much more waterlogged than their contemporary analogies then, in addition to the surface runoff, an intensive internal stream and migration of substances ought to have been developed in Gelic Gleysols there. This had to result in the progress of decalcification and decolmatation (Gerasimov 1960) of solum above the frozen till and translocation of products from Upper to Lower Estonia. The highly similar mineral composition of Upper Estonian tills and varved clays (Пиррус 1968) allows to conclude the formation of the latter from the silt and clay decolmatized from the section of Gelic Gleysols of the Pandivere and Sakala uplands. This, in its turn, gives rise to the presumption that pedogenesis in Late-

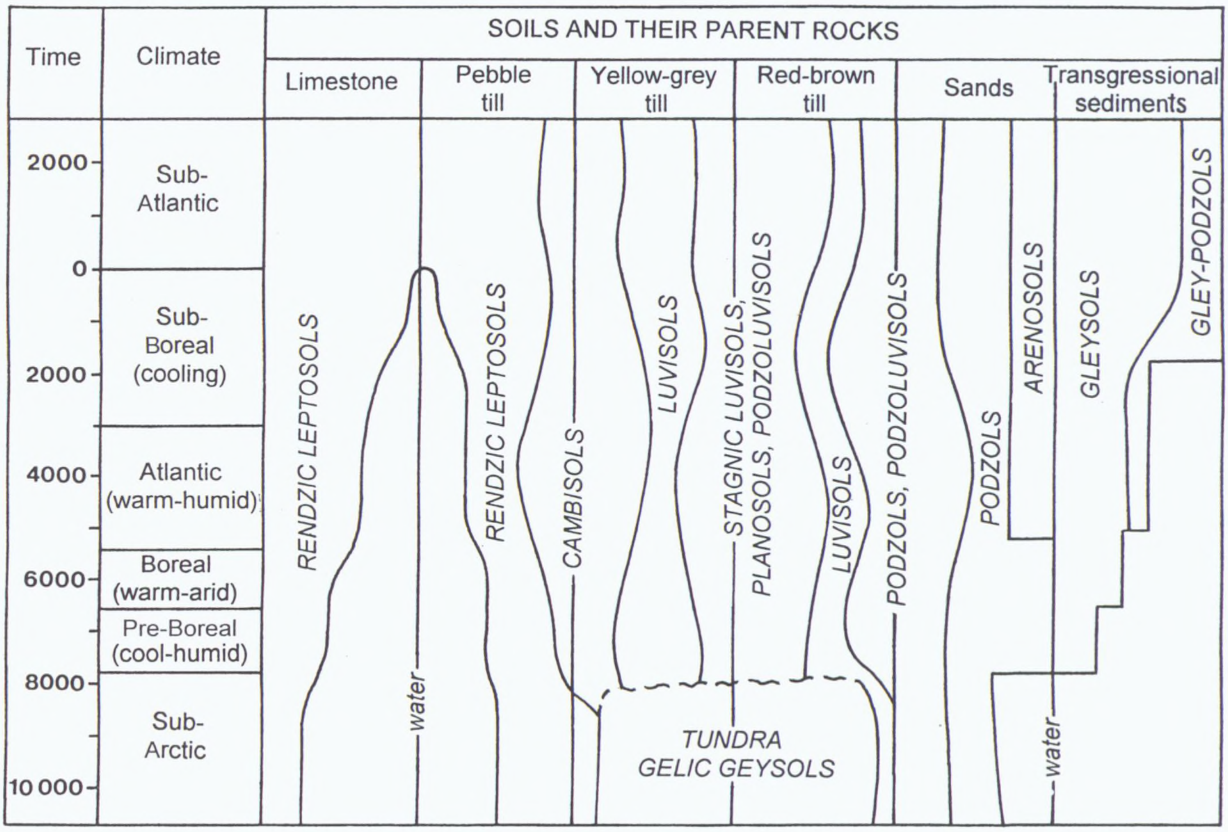


Fig. 198. Progress of soils in the postglacial period.

glacial tundras was connected with the formation and accumulation of fulvic humus, characteristic of eluvial Gelic Gleysols, surface gleyization, weathering of primary minerals and migration of the latter's ferrous products. As a result of these processes, a gley horizon, light in texture and improverished with bases and sesquioxides, was formed above the permafrost in Gelic Gleysols of the Late Pleistocene.

In all likelihood, such a gley horizon in the forest conditions ensuing the tundra situation, was transformed, in dependence of the calcareousness of tills and further development of moisture relationships, vegetation and pedogenetic details, into the eluvial horizon of contemporary Luvisols, Planosols, Podzoluvisols, or even some Podzolic formations. The bisequal or even multisequal texture of geological-pedogenetic origin, characteristic of a lot of contemporary luvisolic, planosolic and podzoluvisolic sections, has probably also formed as a result of permafrost weathering and lessivage in the Late Pleistocene.

Like in contemporary dry site types, in the conditions of cryoarid (subarctic) steppes, the underground herbaceous biomass had highly to exceed the above-ground biomass. On this bases, a progress of the humus-accumulative process can be expected. Without doubt, the arctic rendzina formation took place there. Thus, Rendzic Leptosols in the Pandivere Upland and on the drumlins, end moraines and Upper-Devonian limestones, have a remarkable absolute age beginning from the time of ice lakes. Like the contemporary arctic pedogenesis, the rendzina formation in the Late Pleistocene took probably place at an extremely slow rate. Therefore, rendzinas in Upper Estonia are thin and skeletal, and do not differ from their younger varieties in Lower Estonia. At that, the formation of Gelic Leptosols and their further evolution over Calcari-Gelic Cambisols up to Calcaric Cambisols cannot be excluded, except for the arctic Rendzic Leptosols. Dwarf Ferric and Ferri-Gelic Podzols with Al-Fe-humus ochric epipedon could have developed on sands of different genesis. Their further evolution into deep Ferri-Carbic and Carbic Podzols tends to be quite natural during the Holocene.

The clear differentiation of pedogenetic processes apparently began under the boreal pine-birch stands depending on the calcareousness, layeral texture, chemical, mineral and moisture relationship of parent deposits. Herbaceous ground vegetation, established by pollen analyses, did not apparently form on sands inhabited already in the Late Pleistocene by green mosses and shrubs. In other sites, the humus-accumulative process, induced by herbaceous plants, was common everywhere. Without doubt, beginning from under the boreal stands the pedogenetic transformation and translocation of mineral particles was differentiated in dependence of the structure, texture and composition of parent strata (Fig. 199) into the sections of clay-accumulative (*A–Bm–C*), clay-translocative (*A–EL–Bt* or *Bmt–C*), stagnic ferric-accumulative (*A–(Baf)–ELg–Bt–C*), and hydrolytic-eluvial (*A–E–B–C* and/or *A–E–Bh–C*) origin in both automorphic and hydromorphic conditions.

The lasting retreat of the glacier and the drainage of proglacial lakes decreased soil overmoistening in general. However, ground gleyization in micro- and mesohollows could have intensified due to the continuously high level of the water table in these hollows. The disappearance of permafrost had led to eluviation of the top of former Gelic Gleysols, and their gleyic formation changed into podzolic and luvic on noncalcareous and calcareous tills, respectively. In several other places, the underlying till, frozen prior to the regression of ice lakes, served after the thawing as a stratum for the stagnation of perched water favouring the progress of stagnic, surface-gleyic and luvic properties. Contemporary occurrence of stagnic, stagni-gleyic and surface-gleyic soils as well as Histosols of atmospheric nutrition tends to demonstrate an intensification of seasonal surface waterlogging against the background of a general decrease in overmoistening regime in Upper Estonia at that time.

Surface gleyization stopped and argillization *in situ*, described by Bystriakov (Быстряков 1988), initiated on well-drained tills, rich in carbonates and alumosilicates, on the Pandivere and Sakala uplands and on drumlins. Therefore, besides Rendzic Leptosols developed already in cryoarctic conditions, Calcaric Cambisols and Luvisols differentiated under the Pre-Boreal pine and birch stands are associated with calcareous yellowish-grey till in the Pandivere Upland and on drumlins. Calcaric and Stagnic Luvisols, Planosols and Podzoluvisols differentiated on two-layered glacial deposits and/or reddish-brown tills of medium and heavy texture. Some Podzoluvisols and Podzols tend to be associated with coarse permeable sandy and/or loamy sandy noncalcareous tills and glaciofluvial deposits. Figure 198 shows the formation and progress of Holocene pedogenesis through time. The gradual formation of soil sections and genetic-evolutional connections as functions of the moisture conditions as well as the chemism of parent deposits are shown in Figures 199 and 200, respectively.

During the Ancylus Lake, climate became more arid and warm (Raukas 1995c). With the increase in pine in the composition of Ancylian stands the aggressivity of humus substances rose and hydrolytical processes intensified in soil formation. It was probably impossible on calcareous materials, because the supplies of alkaline earths in biological turnover were scarcely less than those in modern cyclings (Kõlli & Reintam 1970). Podzolization intensified on sands and noncalcareous reddish-brown till of light texture. It is not excluded that also seasonal decrease in stagnic and surface gleyic phenomena developed prior to Ancylus time on bisequal deposits of pedogenetic-geological origin. This could result in the local progress of cheluviation there. In the conditions of arid summer characteristic of boreal forests, the hydrolysis of minerals and cheluviation of products was irregular, because at the same time accumulation of weathering and pedogenetic products formed within the warm season *in situ* favoured similar to contemporary situation (Фирсова 1977).

It may be supposed that as a result of the Boreal regression of the Baltic Sea, the importance of ground gleyization decreased, and reoxidation of reduced ferrous, manganese, *etc.* compounds became evident in deep subsections. Remnant neoformations and concretions in the Bt- and BC-horizons of Stagnic Luvisols, Planosols and Podzoluvisols deeper than one metre as well as ferric-ferrous hydrocarbonates in yellowish-grey calcareous till serve as an indication of the existence of these relics in the conditions under which the contemporary seasonal hydromorphism is lacking.

The warming of climate and consecutive increase in humidity which had begun at the end of the Ancylus period reached the maximum during the Litorina Sea stage. The so-called Atlantic climatic conditions gave rise to the progressive development of broad-leaved herbaceous forests favouring intensive production phenomena, turnover of substances, humus-accumulative process and biological weathering. The development of rendzinas on limestone and stony till was going on simultaneously with the accumulation of secondary weathering products of alumosilicates. Naturally, amorphous sesquioxides prevailed during the first stages. Their step-by-step transformation into crystalline hydroxides was probably similar to the processes characteristic of the contemporary slightly continental subtropics and marine-continental temperate climate (Зонн 1982, Zonn 1986). Biogenic argillization, accompanied by the breakdown and leaching of carbonates, and accumulation of nonsiliceous sesquioxides intensified under the herbaceous broad-leaved stands, rich in hazel underwood. The presence of calcium and magnesium in the interlayer structure of clay minerals and in clays, as a whole, can be hardly qualified as a result of modern pedogenesis (Reintam 1971). This seems to be characteristic of pedogenesis on calcareous-alumosilicate tills from the very beginning.

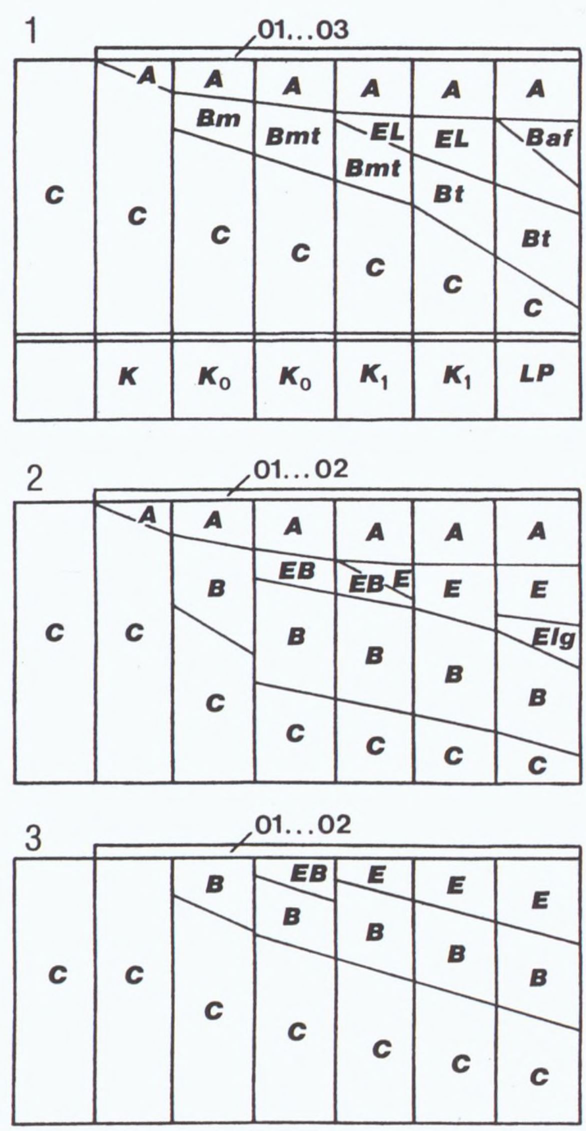


Fig. 199. Formation of soil sections on different parent materials: 1 - mineralogically and chemically rich till; 2 - mineralogically and chemically poor till; 3 - sands. K - Rendzic Leptosols, Ko - Cambisols, Kl - Luvisols, LP - Stagnic Luvisols and/or Planosols. 01-03, A,E, EL, Bm, *etc.* - soil horizons (FAO 1985, 1994; Reintam 1995).

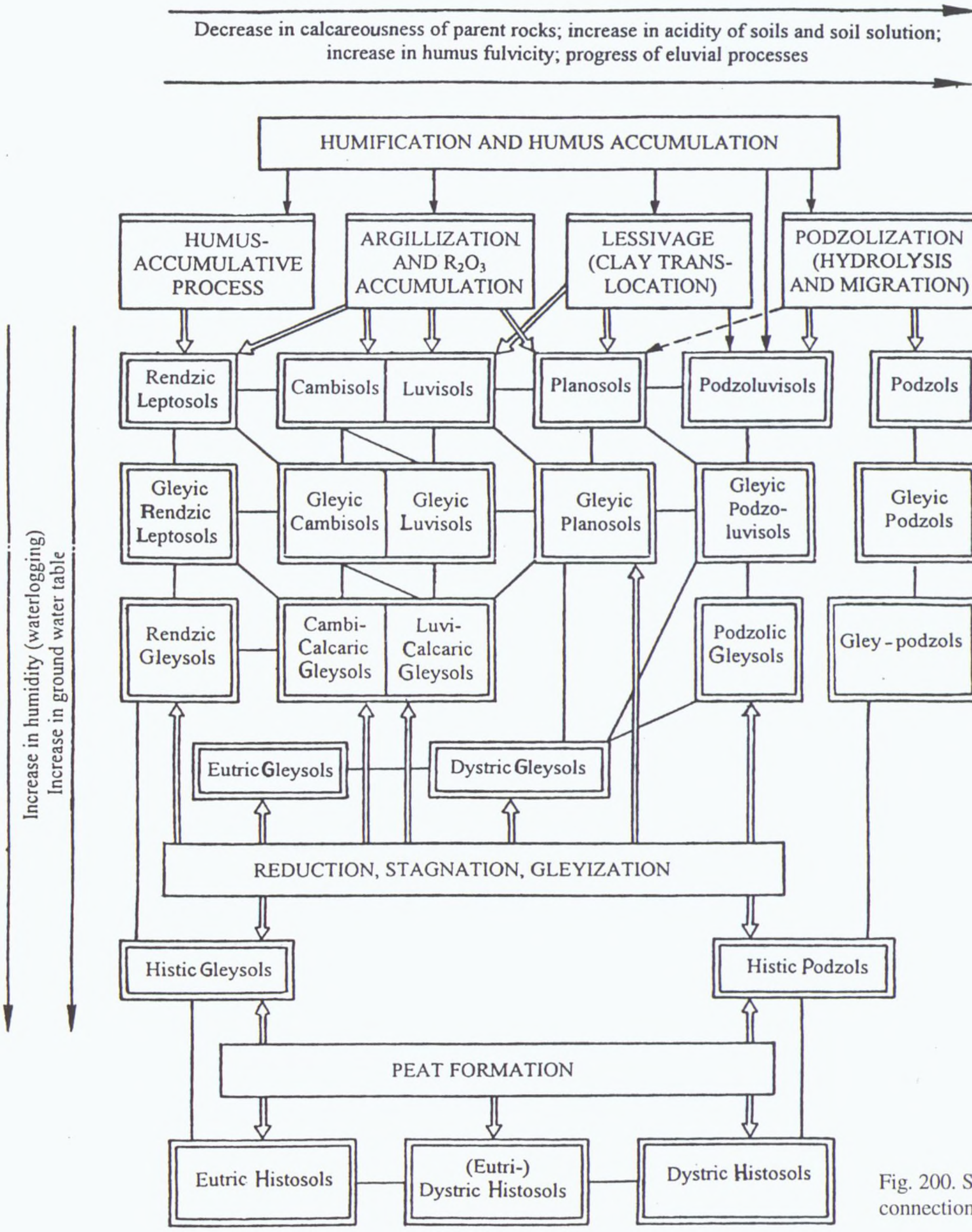


Fig. 200. Soil progress and genetic connections.

So, it must be emphasized that the culmination of argillization, the progress of ferrolysis simultaneously with the accumulation of nonsiliceous sesquioxides, lessivage and the formation of stagnic properties (pseudopodzolization) became widespread under the broad-leaved forests of the Atlantic Chronozone and are still ongoing. The preconditions for the podzolization were (and are) absent on texturally medium and heavy, and chemically rich deposits. That is why Holocene podzolization is mainly restricted to light and chemically poor deposits under pine stands.

The synchronous progress of Cambisols and Luvisols on calcareous tills was probably induced by local textural peculiarities, formed prior to the Atlantic climate and broad-leaved forest vegetation. The influence of lessivage and ferrolysis on the differentiation of luvic sections cannot be excluded, whereas perhaps due to the multisequal development of ferrolysis described by Brinkman (1979) layeral distribution of amorphous iron took place against the background of its pedogenetic profile accumulation. This had been more considerable on glacial deposits, the bisequal texture of which was formed as a result of Pre-Atlantic pedogenetic-geological processes. Seasonal stagnation of perched water at the juncture of the subsections of different texture was accompanied by the lessivage, ferrolysis, deferritisation and autoaluminisation (Pedro *et al.* 1974), and resulted in the formation of Stagnic Luvisols and Planosols in the same way as in contemporary humid forests with a high productive capacity, intensive turnover of substances and pedogenetic activity. Yellowish-brown Baf-horizon below the humus one and above the whitish horizon with stagnic and surface gleyic properties was probably formed as early as in the time of the Litorina Sea. It represents the biogenic accumulation of weathering products above the seasonal perched water table in the conditions of variable hydrothermic relationships. Appearantly, part of amorphous iron accumulated has a subsoil (tilly) origin and has been translocated into the topsoil by the mediation of roots and falling litter. Iron hydroxides are quite stable against the reduction and leaching, because modern stagnic and gleyic properties develop under the seasonal perched water with depth, and biogenic accumulation prevents from topsoil degrading.

Surely, pine stands and Podzols did not disappear on different sands within warm and humid Litorina time. Nevertheless, the development of podzolization in depth could have been inhibited by the deficiency of basic constituents able to neutralize the acid products of ground litter humification. That is why the acidoic chelates precipitated immediately in the top of soil section, very close to the bottom of forest floor. The formation of humus-illuvial (carbic) horizons of different morphology took place. They are often present as very thin stripes in modern Podzol sections and barely have contemporary genesis during some last millennia. The presence of oak in Litorina pine stands is acceptable, whereas the mixed oak-pine falling litter was a significant temporary abation against podzolization, especially in deep sandy massifs on valley terraces. Relics of bygonal humus-accumulativity and/or -illuviation can be found there.

Taking also into consideration different stagnic and gleyic expressions of surface- and ground gleyization in lowlands and depressions of glacial and postglacial topography, it is possible to conclude that the contemporary diverse soil cover (Fig. 200) developed in the Atlantic period. At that, against the background of humus-accumulative phenomena, the ecological situation was more favourable for the progress of argillization, lessivage, ferrolysis and surface gleyization simultaneously with the properties of ferritization and/or ferrallitization above the seasonal perched water, but also for the ground-gleyization in the conditions of high ground water table in depressions and, perhaps, for the podzolization in coarse, permable and poor deposits. Plausibly, accumulative and eluvio-accumulative processes developed more and eluvial ones less intensively than prior to and after the Atlantic Chronozone. Evidence is derived from the comparison of any soil type in the transgression area of the Ancylus Lake and in Upper Estonia where the absolute age of the soil cover is some 4,000–5,000 years greater.

As a result of the Baltic Sea regressions, new territories were subjected to pedogenetic phenomena. Although Rendzic Leptosols, Rendzic and Calcaric Gleysols of Ancylus, Litorina and Limnea age are thinner than their older analogies, the similar substantional-regime characteristics as well as profile structure are typical of all the soils of those types. This testifies to the considerable rate of pedogenesis in its early stages and to its decelerating stabilization with time. The presence of crystalline constituents in calcareous till made the formation of Cambisols and their further evolution into Luvisols possible even in the territory of the Ancylus transgression where they are thinner than in the Pandivere Upland or in the drumlin area. It is difficult to suppose that gleyed and gleyic formations could have occurred prior to the modern automorphic Rendzic Leptosols and Cambisols. Nowhere, all of a sudden and total appearance of vegetation accompanied by synchronous soil formation was incredible, because water regression from morainic hillocks had been temporarily distanced from the process in concave forms of undulating morainic topography.

Alternation of different Gleysols in morainic hollows covered by Ancylus sediments, gleyic formations in transitional areas, and Rendzic Leptosols, Cambisols and/or Luvisols on micro- and mezohillocks is characteristic of current natural situation and tends to demonstrate the phytocoenotic and pedogenetic mosaicity already within the Ancylus and Litorina stages. At that, the progress of argillization *in situ* can be partly connected with thin Holocene sediments on tilly underground and with the weathering phenomena of the latter in the alternating hydrothermic conditions. As stony till, drying analogeously to the modern situation deeply up, did not favour the argillization on local watersheds, the total occurrence of littoral Fluvisols in the entire area of the Ancylus transgression tends to be scarcely supposable. Fluvisols of limnic nature could have occurred only on the concave forms of relief near the regressive and sometimes once again offensive lake.

In connection with the climatic re-cooling and penetration of spruce into the composition of stands about 4,000–5,000 years ago, spruce-decidous forests became prevalent on tills. Pine stands preserved on deep sands and in droughty limestone and stony till areas. Although the acidity of forest humus could have increased under the impact of spruce falling litter, intensification of hydrolytic-eluvial processes was hindered by calcareousness and richness in bases of parent materials and former soil strata as well as by the quite high ashness, characteristic of spruce litter up to now. Naturally as usual, decidous, especially hazel underwood and widespread herbaceous ground vegetation was of a great importance in the prevention of any destructural-eluvial pedogenetic phenomenon. On the basis of these regularities, it is possible to conclude that in the Sub-Boreal and Sub-Atlantic pedogenesis continued following the same trends and directions as before, whereas the mineral and chemical composition (potential) of parent tills and Holocene sediments, exerted as usual, an impact on the phytocoenotic processes and affected interactions between organic matter and mineral strata. Probably, due to these interdependences, the stabilization of plant-soil systems on different parent deposits was reached which, in its turn, enabled the vicinity existence of various soil combinations (Rendzic Leptosols, Cambisols and Luvisols in central Estonia on yellowish-grey calcareous till; Luvisols, Planosols, Podzoluvisols and even Podzols in southern Estonia on reddish-brown till) within a short distance.

As a result of climatic cooling, the mobility of humus substances could have grown and an increase in the migration of organic-mineral complexes as well as in the compaction of illuvial horizons became feasable there where these phenomena had not previously taken place. Of course, the stagnation of perched water and the progress of stagnic and surface-gleyic properties intensified against the background of former lessivage and/or podzolization. Therefore, Planosols and Stagnic Luvisols of bisequal texture have a great territorial importance, whereas many-featured morphology and regional variable properties are characteristic of them (Kokk *et al.* 1985).

During the last two or three millennia, the pedogenesis and soil properties have been considerably affected by human activities (Photo 61). The development of burnt-over tillage brought about the same phenomena found in Karelian and Vepsian materials (Рейнтам и Моора 1983): a decrease in acidity and humus stability, an increase in humus fulvicity, exchangeable bases and base saturation. Clear-cutting of forests, disappearance of forest floor and the opening of soil surface to the weather conditions resulted in the intensification of soil leaching and lessivage from the beginning of human agricultural impacts. Simultaneous use of harvesting residues, manure, composts, crop rotation, *etc*., gave a rise to the man-generated accumulation of complementary organic-mineral complexes into the humus horizon, accompanied by the gradual growth of the latter in depth. Although the amounts of organic matter participating in soil processes have decreased and the turnover of substances in the plant–soil system has become ever opened in arable soils, there is no reason to suppose principal changes in pedogenetic trends and directions as a result of human agricultural activity. In principle, all the Holocene pedogenetic processes and formations are characteristic of both forest and arable lands. Highly different Anthrosols are special and rare.

As a successive drying up of the thin-rooted and weather-opened topsoil is always accompanied by the formation of water-permeable cracks and cleveages in fields, the intensification of lessivage tends to be typical of arable soils from the beginning of permanent agriculture. These phenomena are prevented by the ground litter in the forest ecosystems. In its turn, lessivage leads to the formation of textural argillic horizon instead of cambic one, and stagnation of seasonal perched water in subsoil. That is why also a certain anthropogenic intensification of ferrolysis (pseudopodzolization), and progress of Stagnic Luvisols and Planosols have been found as a result of tillage and cultivation. Advantageous translocation of clay in arable soil conditions enables to explain the light texture of the humus horizon of some Rendzic Leptosols and Cambisols without the eluvial differentiation of their entire sections.

The differentiation of current gleyic soil sections and the character of pedogenetic processes inducing gleyization tend to be universal in different moisture relationships. Without doubt, not only ferrolysis and lessivage, but also any eluvial phenomenon, in dependence on parent material, were favoured by the gleyization within the entire Holocene. As these conditions were less conductive to the argillization and ferrallitization, cambic, ferralic and/or argic properties are weak in the sections of some Gleyic Cambisols, Gleyic Luvisols and Cambi-Calcaric Gleysols. A great number of Eutric Histosols have been evolutionated from Gleysols through their Histic subtypes.

Current soils

Automorphic Leptosols, Cambisols, Luvisols, Planosols, Podzoluvisols, Podzols and Arenosols with their gleyic subtypes and eroded kinds cover 42.0%, different Gleysols 32.5%, and Histosols (mires and bogs) 23.2% of Estonia's territory (Table 49). Fluvisols (alluvial and saline litoral soils) form 2.1% and technogenic rendzic formations on reclaimed land 0.2%. By texture, sandy soils make up 26.7%, peaty soils - 23.7%; loamy sands, loams and clays occupy 17.0, 27.8 and 4.8%, respectively. Of the latter, 50.1, 38.2 and 11.8% are arable, correspondingly (Reintam 1995, Kokk 1995). Nearly 50% of the underground consists of Ordovician and Silurian calcareous sediments, 75% of soil parent material is calcareous.

Rendzic Leptosols (*Rendzinas*) are formed on limestones and dolomitized limestones, on strongly calcareous stony till and on coarse glaciofluvial materials. They represent an original soil formation having been developed on stony calcareous deposits within any historical stage of pedogenesis. As Rendzic Leptosols occurred already in the Pre-Pleistocene, their (although temporarily interrupted) dissemination is steady and typical of northern and central Estonia. Calcareousness higher than 10% already in the humus horizon, short skeletal non- or slightly differentiated profile, calciphilous vegetation and relative richness in humus have always been characteristic of Rendzic Leptosols.

Cambisols on calcareous yellowish-grey or reddish-brown tills are characterized by the argillization *in situ* of the entire solum or Bm-horizon under the humus one, relative accumulation of nonsiliceous hydrates of sesquioxides, favourable hydrothermal, redox and biological relationships. Their independent progress is related to the Holocene, because eventual Pre-Pleistocenic Cambisols were destroyed by glacier.

Luvisols on calcareous tills have much the same early genesis as Cambisols, but they are characterized by morphologically-distinguished lessivage (translocation of clay). A–EL–Bt(Bmt) section (solum) is also rich in nonsiliceous sesquioxides, biologically active and like that of Cambisols

Photo 61. Eroded soils cover extensive areas in the hilly topography of southeastern Estonia and in these regions special soil-protective agrotechnology and crop rotation should be used. *Photo by A. Raukas.*

rich both in humus and available moisture. In spite of territorial closeness, the pedogenetic differentiation of Luvisols from Cambisols commenced under the boreal herbaceous forests is still ongoing.

Stagnic Luvisols (*brown pseudopodzolic soils*) are formed on bisequal glacial deposits of pedogenetic-geological origin under the seasonal stagnation of perched water at the juncture of layers of different texture. These soils are characterized by an accumulative Baf-horizon, rich in amorphous iron, between humus (A) and stagnic eluvial (Elg) horizons. The progress of Stagnic Luvisols started as early as the Boreal against the background of Gelic Gleysols after the regression of tundra and permafrost. It culminated under the broad-leaved forest vegetation of Litorina age.

Planosols (*light pseudopodzolic soils*) on tills of loamy and clayey texture are characterized by the presence of a white-coloured stagnic horizon, rich in amorphous iron and ferric concretions, immediately below the humus horizon. Significant textural argillization is characteristic of these soils, however, some surface gleyization can be also found. Genetically, they are similar to Stagnic Luvisols with which they have progressed synchronously during the Holocene.

Podzoluvisols on bisequal glacial, glaciofluvial, aeolian, *etc.* sediments above glacial ones have both podzic and stagnic properties in their multisequal section. Textural contrasts between layers make possible the synchronous development of a great number of pedogenetic processes. Prior to the distinguishing of Stagnic Luvisols in FAO-ISRIC system, the latter were interpreted as Podzoluvisols, because there are some similarities in their genesis and profile structure.

Podzols on noncalcareous reddish-brown loamy sandy till have a humus horizon in their top. This is usually lacking or rarely present on sands of different nature. Flowing regime, good filtration, low-intensive ash-deficient turnover of substances and agressive mobile humus are characteristic of Podzols. Humus-illuvial (Carbic, Carbi-Ferric) subsection is widespread in sandy Podzols. Podzolization and Podzols have always been connected with sands not only within the Holocene, but also in the Pre-Pleistocene. Therefore, Podzols (as Rendzinas) have occurred in any stage of climate and vegetation.

Arenosols on coarse glaciofluvial sands have a slightly ferritized cambic and weakly differentiated profile. Their formation tends to be connected with moisture deficiency and arid herbaceous forest vegetation since the ice-lake stage. Since they resemble both weak Cambic Podzols and Ferralic

Table 49. The occurrence of soils, % (Reintam 1995)

Soils	Total	Forest area	Arable area
1. Rendzic Leptosols on limestone	1.2	0.8	0.8
2. Cambi-Rendzic Leptosols on stony till	4.7	1.9	9.0
3. Calcaric Cambisols and Calcaric Luvisols in complex	6.6	2.6	16.0
4. Stagnic Luvisols and Planosols	5.9	1.6	15.1
5. Podzoluvisols	3.0	2.7	3.3
6. Podzols	1.6	3.8	–
A. Total dry and normal-moistened soils	**23.0**	**13.4**	**44.2**
7. Gleyic-Rendzic Leptosols	1.6	0.6	2.1
8. Calcari-Gleyic Cambisols and Calcari-Gleyic Luvisols	7.3	3.1	12.0
9. Stagni-Gleyic Luvisols and Planosols	3.6	2.0	6.2
10. Gleyic Podzoluvisols	2.0	1.6	1.9
11. Gleyic Podzols	0.9	2.2	–
B. Total *Gleyic* soils	**15.4**	**9.5**	**22.2**
A + B. TOTAL AUTOMORPHIC SOILS	**38.4**	**22.9**	**66.4**
12. Calcari-Rendzic Gleysols	1.4	0.7	0.9
13. Eutri-Calcaric and Dystri-Calcaric Gleysols on till	10.0	10.0	7.0
14. Eutric and Dystric Gleysols on aqueous deposits	11.1	10.1	8.2
15. Podzolic Gleysols	2.9	4.1	0.8
16. Carbi-Gleyic Podzols	2.2	5.1	–
C. Total *Gleysols*	**27.7**	**30.0**	**16.9**
17. Histic Gleysols	4.7	5.3	2.5
18. Carbi-Histic Podzols	1.6	3.1	–
D. Total *Histic* soils	**6.3**	**8.4**	**2.5**
19. Eutric Histosols	13.8	16.1	7.8
20. Eutri-Dystric Histosols	3.7	6.9	0.1
21. Dystric Histosols	5.7	13.7	–
E. Total *Histosols*	**23.2**	**36.7**	**7.9**
C +D +E. TOTAL HYDROMORPHIC SOILS	**57.2**	**75.1**	**27.3**
22. Eutric and Mollic Fluvisols	1.4	1.0	0.8
23. Salic Fluvisols	0.7	0.3	–
F. TOTAL FLUVISOLS	**2.1**	**1.3**	**0.8**
24. Eroded and deluvial soils	2.1	0.1	5.5
25. Others (Anthrosols, *etc.*)	0.2	1.0	–

Cambisols, they are not yet distinguished as a distinct entity in soil maps.

Gleyic subtypes of any automorphic soil type are characterized by some raw humus in A–horizon, stagnic properties in topsoil and gleyic properties in subsoil. As an exception serve Stagnic Luvisols, Planosols, Podzoluvisols and loamy sandy Podzols, in which ground-gleyization is lacking, and surface gleyization in the kind of bluish marble and iron neoformations occurs in ELg–Btg (Bg, BCg) subsection. The importance of gleyic phenomena has been changing in time and space during the Holocene up to now.

Gley-Podzols (*Podzolic Gleysols*) are formed on noncalcareous sandy deposits. Carbic and/or Carbi-Ferric nature is mostly characteristic of them. Their Holocene history and genesis tend to begin from the time of ice lakes. Later they developed and differentiated under pine stands.

Gleysols have been the prevailing soil formation since the permafrost tundra stage and they include different hydromorphic types of ground and/or combined surface-perched-ground aquic nutrition. Rendzic Gleysols, Cambi-Calcaric and Luvi-Calcaric Gleysols develop on calcareous till, but most of Eutric and Dystric Gleysols on graded deposits of the Baltic Sea transgressions. On different water-laid sediments, the base-saturated Gleysols on flat territories or on microlowlands often alternate within a short distance with the acid Podzol-Gley formations on microheights. The properties of Gleysols highly depend on the water nutrition as well as on the chemism of water and parent material. Together with Rendzic Leptosols and sandy Podzols, Gleysols represent the third formation which is spread and developed within any stage of phytocoenotic and pedogenetic activities. At that, Histic Gleysols have always occupied the transitional territory from Gleysols to Histosols.

Histosols (*mires* and *bogs*) have a peat (*Histic*) horizon with a depth of more than 30 cm. They have developed from Gleysols within the Holocene, but also as a result of the eutrophication of waterbodies. Eutric Histosols (lowland mires) are characterized by a ground-water or flooding regime, Dystric Histosols (transitional mires and raised bogs) – by atmospheric nutrition.

Fluvisols are the current formation of seasonal inundation and accumulation of alluvial, lacustrine, *etc.* suspensions. They have been steadily formed within the Holocene and even Pre-Pleistocene, prior to any following geological event arose a rechange in pedogenetic phenomena.

IX HUMAN IMPACT

Prehistoric times

Interpretation of the standard pollen diagrams in terms of human impact during Mesolithic and Early Neolithic times when man's economy was mainly based on hunting, fishing and gathering, is a complicated task. The development of woodlands during that time was controlled by climatic, edaphic and other ecological factors rather than by human influence (Poska & Königsson 1996). Nevertheless, the Stone Age people also utilized the environment around habitation sites, causing disturbances of local importance and introducing in this way more favourable conditions for the flourishing and spread of light-demanding and nitrophilous herb species (Behre 1988, Poska 1994, Veski 1992, 1996a, b).

The first traces of primitive farming in Estonia date from Neolithic time. Cattle rearing and crop cultivation were properly introduced during the Bronze Age (Lõugas 1992). With more advanced agriculture, man´s influence upon the environment greatly increased.

The systematic research on human impact in Estonia was undertaken by palaeoecologists and archaeologists in the 1990s when coastal areas of Saaremaa Island (Jõhvikasoo), northern (Maardu, Kahala, Tondi) and northeastern Estonia (Kunda, Narva) were studied in the frame of the PACT Project (Hackens *et al.* 1996). As the coastal area is supposed to be a cradle of Estonian farming, the above-mentioned regions were chosen to describe the development of the early agriculture. During the mentioned project, fossil field remains were discovered and dated at Maardu and Saha-Loo (Lang 1996).

The following is an attempt to give an overview of the data available on the history of the colonization and land-use based on the biostratigraphical evidence in Estonia. The discussed material covers almost the whole Estonian territory with different landscape regions and a time span from the beginning of the Mesolithic until historical times.

More than 30 pollen diagrams were examined to find indications of human impact on the environment; 23 diagrams were selected to treat this problem (Fig. 201). The chosen localities, except some earlier-mentioned sites, were studied mainly for stratigraphical purposes. Most of the used sequences were dated. The time scale of five undated sequences (Kunda, Vedruka, Tõhela, Surusoo and Pitkasoo) is based on the correlation with the sites in the closest surroundings. All radiocarbon dates are uncalibrated.

Biostratigraphical data has been summarized in terms of human impact using the method suggested by Behre (1981). Based mainly on the works of Behre (1988) and Hicks (1990), the following plant groups were separated:

1. Cultivated land (Cerealia, *Secale, Triticum, Triticum spelta* t., *Hordeum, Avena,* Cannabaceae, *Cannabis* t., *Centaurea cyanus, Fagopyrum, Polygonum convolvulus*)

2. Fresh meadows (Ranunculaceae, *Ranunculus acris* t., *Plantago sp., Plantago lanceolata, P. m/m* t., *P. maritima, Potentilla, Cerastium, Achillea* t., *Saussurea* t., *Solidago* t., *Cirsium* t., Caryophyllaceae, *Trifolium, Linum catharticum, Rhinanthus* t., *Centaurea scabiosa, C. jacea,* Scrophulariaceae, *Helianthemum, Succisa* t., *Alcemilla* t., *Hypericum* t., *Gypsophila muralis, Vicia cracca, Valeriana*)

3. Ruderal communities (*Artemisia,* Chenopodiaceae, *Chenopodium album,* Brassicaceae, *Urtica, Plantago m/m* t., Onagraceae, *Chamaenerion,* Polygonaceae, *Polygonum aviculare, P. persicaria, Rumex ac/ac* t., *R. acetosella, Centaurea sp., C. nigra,* Malvaceae)

4. Grazed forest (*Pteridium, Melampyrum*)

5. Dry pastures (*Juniperus, Cerastium* t., *Campanula*).

Conclusions considering forest disturbances are based on the abrupt changes in tree pollen curves, finds of light-demanding NAP taxa, increase in charcoal dust amount and a general rise in anthropogenic indicator graphs. A fluctuating character of these indications is considered to show different phases of settlements: establishment, flourishing and abandonment. Appearance of Cerealia pollen together with other evidence is supposed to characterize the cultivation practice.

The first traces of human habitation in Estonia come from the Mesolithic (9500-6000 yr BP, Kriiska 1996a). Settlements of this period were commonly situated along the water-edges, in ecothons where the plant communities are sensitive to disturbances and even a relatively weak stress may yield pronounced and long-lasting consequences. The nomadic character of the Mesolithic people, the successive utilization of the different biotopes (coastal and inland sites) after seasonal food availability may have enlarged the influenced area considerably. In Estonia, the oldest known settlement site, dated to 9600±120 (TA-245) and 9575±115 (TA-176; Кессел и Пуннинг 1969) was situated near the present-day Pulli Village on the Pärnu River bank (southwestern Estonia). Detailed palynological studies in the surroundings of the settlement site have not yet been performed. The Kunda Settlement (northern Estonia), which appeared somewhat later and gave a name to the Mesolithic Kunda Culture of Estonia, has been better studied (Photo 62).

The finds of Kunda Culture are known from different parts of Estonia. Based on biostratigraphical material, the net remains found by Indreko (1932) from the former sea bottom at Siiversti in the Narva area are referred to the Boreal/Atlantic limit (Jaanits *et al.* 1982), being the oldest of this kind in northern Europe (Kriiska 1996c).

Particularly abundant Kunda Culture finds come from the Võrtsjärv Lowland. The Siimusaare, Jälevere, Tamme,

Photo 62. Lammasmägi Hill near Kunda. About 8500 years ago its surroundings were inhabited by hunters and fishers who lived in small clan communities and gave a name to the well-known Mesolithic Kunda Culture. *Photo by A. Miidel.*

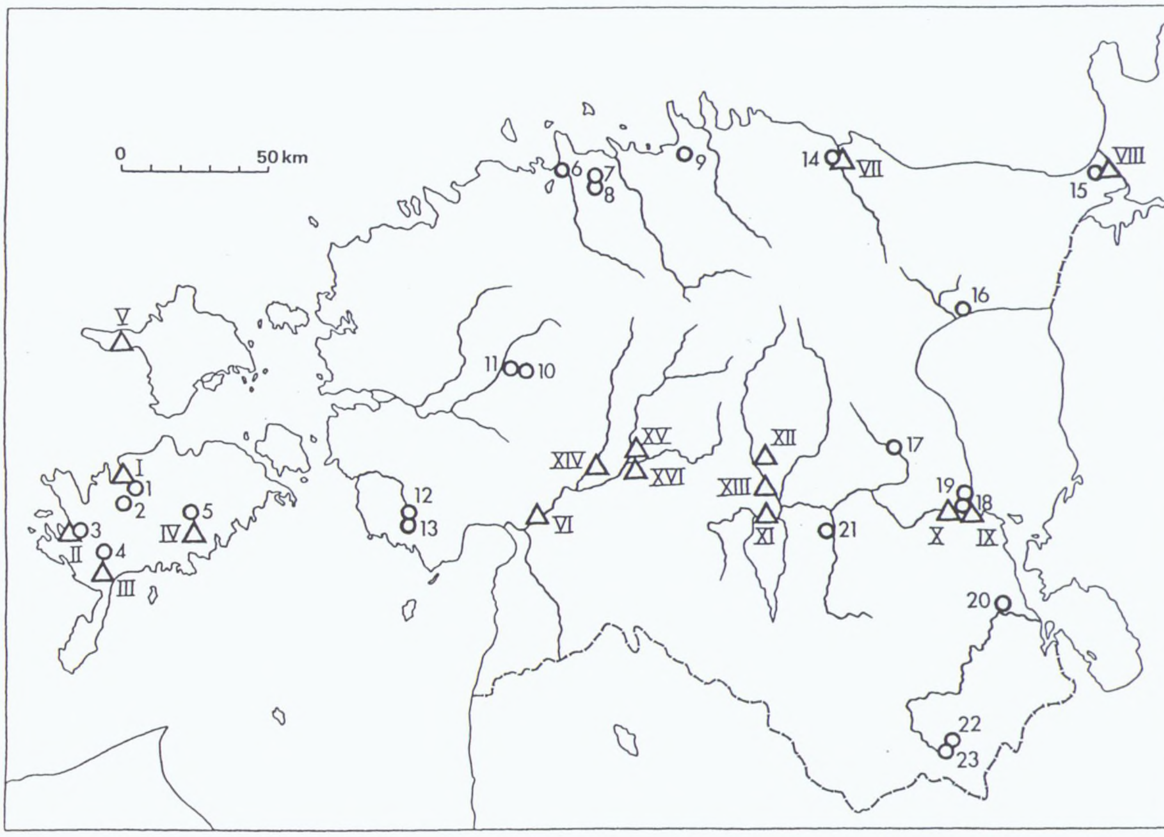


Fig. 201. Location of biostratigraphical (dots) and archaeological sites (triangles), mentioned in the text. Biostratigraphical sites: 1 - Surusoo (Veski 1996b), 2 - Tuiu (Jõhvikasoo, Hansson *et al.* 1996), 3 - Vedruka, 4 - Pitkasoo, 5 - Kaali, 6 - Tondi (Kimmel *et al.* 1996), 7 - Maardu (Veski 1992, 1996a), 8 - Saha (Veski, unpublished), 9 - Kahala, 10 - Velise (Veski, unpublished), 11 - Kiilaspere (Veski, unpublished), 12 - Tõhela (Veski, unpublished), 13 - Ermistu (Veski, unpublished), 14 - Kunda Arusoo, 15 - Leekovo (Lepland *et al.* 1996), 16 - Võhma (Kimmel 1994), 17 - Raigastvere (Pirrus *et al.* 1987), 18 - Akali (Moora *et al.* 1988); 19 - Saviku (Сарв и Ильвес 1975), 20 - Kalsa (Kimmel 1994), 21 - Ulila (Ильвес и Сарв 1970), 22 - Vaskna (Ilves & Mäemets 1987), 23 - Tuuljärv (Ilves & Mäemets 1987). Archaeological sites (Jaanits *et al.* 1982): I - Võhma, II - Loona, III - Naakamäe, IV - Kõnnu, V - Kõpu, VI - Pulli, VII - Kunda, VIII - Narva, IX - Akali, X - Kullamäe, XI - Moksi, XII - Umbusi, XIII - Siimusaare, XIV - Lepakose, XV - Tamme, XVI - Jälevere.

Lepakose, Umbusi and Moksi dwelling-places have been examined by archaeologists (Jaanits *et al.* 1982, Jaanits 1992). The biostratigraphical material is scattered or absent in the direct vicinity of the above-mentioned settlement sites. Still, some evidence of landscape disturbance in the area during Mesolithic time may be found on the available pollen diagrams (Ильвес и Сарв 1970).

In eastern and southeastern Estonia, the Neolithic dwelling-places at Akali and Kullamäe are well known at the mouth of the Emajõgi River (Янитс 1959, Moora *et al.* 1988). The bio- and chronostratigraphical data refer to the existence of the habitation in the Akali area earlier than recorded by archaeologists.

During the last decades, new Late Mesolithic settlement sites have been excavated on the islands of Saaremaa (Võhma, Kõnnu; Lõugas 1982) and Hiiumaa (Kõpu; Лыугас 1988, Kriiska 1996b). The last biostratigraphic studies in the vicinity of the ancient settlement sites on Saaremaa: Pelisoo, Surusoo, Jõhvikasoo (near Võhma); Pitkasoo (Naakamäe) and Vedruka (Loona, Kurevere) have yielded supplementary information concerning human activities during Late Mesolithic and Neolithic times (Saarse & Königsson 1992, Poska 1994, Hansson *et al.* 1996, Veski 1996a). New palynological evidence from the region points at the presence of Mesolithic settlers not only in the Võhma, but also in the Käesla (Naakamäe?) and Kihelkonna (Loona?, Kurevere?) areas. Pollen evidence from Kõivasoo Mire (Hiiumaa) shows forest disturbance about 6000 yr BP which is consistent with the latest data from archaeological excavations in the Kõpu area (Kriiska 1996b).

Up to the end of the Stone Age, the economy of man was mainly based on hunting, fishing and gathering. In the Neolithic, indications of human impact traceable with biostratigraphic methods, were much the same as during Mesolithic time. Nevertheless, primitive crop farming techniques, such as slash-and-burn agriculture, were introduced to Estonia already during the Neolithic, and settlements started to spread all over the country (Jaanits 1992). The expansion of the farming activities led to the introduction of some new plant species (both cultivated and accompanying ones), and the need for cleared land increased.

During the Neolithic, the first more extended indications of the forest clearance appear on the pollen diagrams between 5000-4500 yr BP. The earliest finds of cereals in the Maardu area are dated to about 3800 yr BP (Veski 1992), in the Saadjärve Drumlin Field to 4495±35 yr BP (Kõrenduse; Pirrus & Rõuk 1988) and *ca.* 4100 yr BP (Velise; Veski, unpublished data), being older than in northern Estonia. Such distribution of the first cereal findings indicate that the tillage could have been introduced to Estonia from the south and southeast direction. The crop farming spread then all over the country, but gained importance in man's economy only in areas with favourable natural conditions (soils, water, *etc.*).

In the Bronze Age, farming gained importance, overruled fishing and hunting, and became basis of man's economy. The Late Bronze farming revolution (Veski & Lang 1996) led to the drastic changes in land use and the spread of cultivation from easily tilled alvar soils to the other land. Due to the evolution of agrarian techniques, the habitation moved from coast to inland. The main plant species cultivated during the Bronze Age were: *Triticum dicoccum, T. monococcum, Hordeum vulgare, Linum usitatissium, Pisum sativum, Lens esculenta* (Behre 1988). In contrast to the Stone Age forest clearances, after which a regeneration of forest normally occurred, the clearances in the Bronze Age often brought about permanent changes in ecosystems. The formation of the majority of Estonian *alvars* probably started during the Bronze Age.

The Iron Age, as a whole, is characterized by well established quickly spreading and developing farming. New agrarian techniques enabled man to extend the size and amount of the fields and pastures and explore less favourable parts of Estonia. The final revolution in land occupation started in the Late Iron Age about 1100 yr BP.

Regardless of relatively high density of sites investigated in terms of pollen analysis on Estonian territory during the last decades, the quality of the determinations and the availability of datings only in a few sequences corresponds to the level required for investigations of prehistoric human impact. Still, on the basis of the discussed data some outstanding regions and periods in the prehistoric land use are proposed:

1. The biostratigraphical records show the same Stone Age habitation centres in Estonia as proposed by archaeologists (Jaanits *et al.* 1982, Kriiska 1996a,b):

a) North-Estonian Coast,

b) Võrtsjärv Lowland,

c) Eastern part of the Peipsi Lowland (lower course of the Emajõgi River),

d) Islands of Saaremaa and Hiiumaa.

Palynological indications of the human influence upon the vegetation in southeastern Estonia do not coincide with the archaeological records. As the data available are rather episodic, further investigations are needed.

2. Based on the available biostratigraphical evidence, the following regional phases of the human impact may be distinguished:

a) At 9000 yr BP, the start of a period with indications of woodland disturbances in the coastal area of the Baltic Sea, on lake shores and in river valleys of northern and northeastern Estonia.

b) At 7000-6500 yr BP, the second period of forest disturbances occurred in northern Estonia and on the Võrtsjärv Lowland, at ca. 6500-6000 yr BP on the islands of Saaremaa and Hiiumaa, and in eastern Estonia.

c) At 5000-4500 yr BP, the relatively well defined period of increasing human impact is traceable almost all over Estonia. The first Cerealia pollen finds are recorded in eastern Estonia (Kõrenduse).

d) At 3800 yr BP, pollen spectra of several diagrams indicate increasing forest clearance, and in some cases, also a primitive crop farming (Cerealia and *Plantago lanceolata* pollen in L. Maardu; Veski 1996b).

5) At 3300-3200 yr BP, indications of extensive forest clearances and crop farming in northern Estonia, on the Võrtsjärv Lowland and on the Island of Saaremaa.

6) The first finds of the *Secale cereale* pollen are known already from 2500 yr BP (Ilves & Mäemets 1987), but obviously it was then growing as a weed in crop fields.

The real introduction of *Secale cereale* is likely to have happened at about 1400-1500 yr BP.

Historic times

Population. At the beginning of the Middle Ages the population increased, favoured by climate and social relations. The inhabitants settled all over Estonia, even the Alutaguse and Kõrvemaa areas were sparsely populated. A historical document dating from 1228, names Hiiumaa a "desert" island, but it seems not to have been the truth (Tarvel 1992). Estimates place the population of Estonia in the 13th century at about 100,000-150,000 people (Tarvel 1992). The main settlement type was village, in southeastern Estonia scattered habitation prevailed. The *Liber Census Daniae* (Eisen 1920, Johansen 1933) contains data about Estonian villages, showing that the 13th century Estonia was a typical agrarian country with villages and fenced strongholds, into which manors started to integrate.

During the course of the next three (14th-16th) centuries, the population of Estonia doubled. According to Vasar, some 250,000-280,000 people lived in rural districts. Together with citizens, the population of Estonia could not have been more than 300,000 at the end of the 16th century (Tarvel 1992). However, since there is very little demographic data about the ancient inhabitants of Estonia, this number is very approximate. The main fields of economy were crop farming and cattle rearing, about one fourth of the yield was exported to Finland, Sweden, the Netherlands, Germany and even to Portugal.

Population dynamics and economy were fundamentally affected by wars, epidemics and famines, causing the death of thousands of people. After the Russian-Livonian War (1558-83), 50-70% of farms were deserted or abandoned. After the Polish-Swedish War (1600-29), about 87% of farms in the present-day Tartu County stayed desolated (Tarvel 1992). The fighting and outbreaks of epidemics and famine ravaged the country and reduced the Estonian population nearly by two-thirds (from 250,000-280,000 inhabitants before the Livonian War to about 70,000-100,000 in 1625). The period following those hard times witnessed a rapid economic development; deserted farms were taken into use and the yield export expanded once again. The number of population increased and before the Great Famine there were about 375,00-400,000 inhabitants in Estonia (Palli 1992).

The regional mapping in 1684 shows that the distribution of farms and rural settlements was much the same as in recent times (Troska 1992). After the Great Famine (1695-97), the Northern War (1700-21) and the plague that ravaged Estonia in 1710, the survivors numbered 150,000-170,000 in 1712. The consequences of the epidemic were most horrible in western Estonia.

In the 18th century the rural population increased. In the middle of the century, Estonia had 330,000 - 340,000 inhabitants, of those only 5% lived in towns (Palli 1992). In 1797, there were 17,212 farms in the Estonian Province (North Estonia). In the beginning of the 19th century, the number of farms in Estonia was at its highest - 18,879 (Troska 1992); thereafter it started to decrease. In the middle of the 19th century, the size of the population had doubled, reaching about 750,000 inhabitants, 14% of whom lived in towns. The population was concentrated to villages and, thus, Estonia still stayed a typical agricultural country. Since the 1950s, the urban population rapidly increased, forming 71.4% of the total population in 1990 and showing a drastic reorientation in the Estonian economy (National....1992). The present-day Estonia is a highly urbanised industrial country.

Landuse. Feudal relations, foreign conquest and division of Estonia into several parts between different powers were the main reasons, why agriculture flourishing at the end of the 12th century, made a slow progress in the Middle Ages. Unfortunately, the written sources do not give the full clarity of the medieval tools and tillage. For this reason, it is not known when exactly the manuring of fields started. It was first mentioned in historical documents in the 15th century, which does not mean that the manure had not been used earlier. The main tools were wooden plough, harrow, hoe, sickle and scythe. As to the cereals, most important was *Secale cereale* (rye) cultivation, followed by *Hordeum* (barley) and *Avena* (oat). The share of *Triticum* (wheat) was low, being more widely grown in western and southern Estonia. In Livonia (South Estonia), *Linum* (flax) was also cultivated and exported (Kivimäe 1992). From horticultures cabbage, carrot, onion, parsley and hop were grown. Besides crop farming, animal husbandry was also a very important branch of economy and a booty for the conquerors (Henriku...1993).

After the Polish-Swedish War (1600-29), the economic development in Estonia gained in momentum following the same trend as in the succeeding centuries. This was promoted by improved tillage and manuring the fields. Three-field-crop-rotation system which dominated up to the 17th-18th centuries was accomplished, slash-and-burn agriculture lost its importance in northern and western Estonia, but remained in use in southern Estonia (Livonia). In northern Estonia, the clearing of stony fields was widely practiced. The maps of the 17th century show stone edges and cairns.

The first data on draining of land by ditches in Estonia comes from the 14th century in the Maardu area, but these were only single attempts. It was not until the 17th century that large-scale amelioration was undertaken in the estate fields of northern Estonia and on the Island of Saaremaa.

In the late 19th century, a large number of tenants began to buy farms in perpetuity. This brought about a new rapid development of agriculture, particularly crop farming and animal husbandry. The production of timber made a rapid progress, the area under arable land was expanded and the forested area as a whole, decreased to *ca*. 20% by the beginning of the 20th century. The situation turned critical and called for afforestation. Forest management has a long tradition in Estonia. The first orders prohibiting felling of forest were put into force on the islands in 1297, the status of forest keepers was introduced in 1696, regulations for forest management in 1782 and forest plantations have been cultivated since the last century (National... 1992). Currently, about 48% of Estonia's land area is taken up by forests, the main forest-forming species being pine, birch and spruce. This means that during the last 50 years the forested area has increased 2.2 times, mainly on the account of the overgrowing of abandoned fields and meadows with brushwood. Land drainage, air pollution, intensive felling, *etc*., have affected the spread of swampy black alder and boreo-nemoral hardwood-spruce forests which are currently rare or on the verge of extintiction. In recent years, the first signs of damage caused by acid rains and new species of forest pests have begun to manifest themselves (National... 1992).

In this century, the Estonian agriculture has passed three revolutionary changes: (1) 1918-19, implementation of the land reform which touched mostly manors and dispossessed big landowners of land, (2) large-scale collectivization in 1949, and (3) privatization and establishment of small private farms in the 1990s. The agrarian structure, which has a long historical background, being developed in the frame of private farmsteads and estates, was completely destroyed during the Soviet period. Small patches of arable land, tracking landscape peculiarities, were united into huge fields. One third of the former fields were left uncultivated and almost as large area of natural grassland was cultivated into fields. In the Haanja and Otepää heights and on the Sõrve Peninsula about 50% of previous fields were abandoned because, being too small and scattered, they were not profitable for cultivation with heavy machinery. The use of heavy machines resulted in the compaction and degradation of arable soils, affecting their porosity and moisture content. Large fields suffer from wind erosion, the fields on the hummocky terrain are subject to surface water erosion. But there were also positive aspects of the collective farming: large fields were cleared up of stones, fields were regularily limed and fertilized, and almost 66% of fields were subject to amelioration. However, today the out-of-date drainage system needs reconstruction to avoid paludification of drained fields.

During the Soviet power, the grasslands, meadows and natural or seminatural pastures which in 1939 formed 24.5% of Estonia's area (National... 1992) were partly cultivated into grassland, partly afforested and partly became overgrown with brushwood.

Ecological problems. The development of the landscape and environment has been mainly controlled by two factors: natural (climate, geological-geomorphological-hydrological conditions, soil texture, *etc.*) and social (human activities). We have reached the state where special expenditures are needed for finding out major agents affecting soil fertility and production level, for controlling the exploitation of national resources along rational complex line and for ensuring that man´s living conditions will not deteriorate any more.

In the 1950s, Estonia became a region of very intensive agriculture and industry. This caused damage to soils, inland and sea water, air and to the environment, as a whole. The soils and water were subject to pollution by the wastes of urban and rural settlements, industry, big farms and military camps. From some rather big urban settlements, like Tartu, where sewage treatment plants are lacking, the unpurified waste water is discharged directly into the rivers. The main pollution load comes from northeastern Estonia which is the region of mines, electric power stations and chemical plants. Self-ignition in oil-shale waste hills, harmful chemicals and minerals, phenols and oil have contaminated soils and rivers and reached the sea. The water of the Gulf of Finland contains nutrients, oil products, phenols, heavy metals, toxic elements, the concentration of which is rather high near Tallinn, Sillamäe and Narva.The residual of petrol and rocket fuel from military airfields, camps and depositories has polluted the soil and entered the groundwater (at Tapa, Kärdla, Aruküla). Excessive use of pesticides and herbicides in agriculture and set up of big rural settlement without proper utilization of domestic wastes, has enlarged the pollution danger to the country.

The situation is especially critical in the areas with a high concentration of population, mines and industrial enterprises (Tallinn, Kunda, Kohtla-Järve, Sillamäe, Narva). The area spoilt by mining, buildings and industrial activities amounts to 30,648 ha (Ranniku 1993). Since the beginning of the use of oil shale deposit, 858 million tons of oil shale have been mined; the production culminated in 1980 (Paalme 1995). The total area under the oil-shale mines is about 11,000 ha. Of that some 9000 ha damaged by surface mining has been reclaimed (Paalme 1995). About 6000 ha of land is under milled peat fields. The mining of phosphorites was terminated in 1991, but 59 ha of the mined-out area still remain to be recultivated. Most of the exhausted areas are usually reforested, some turned into fields, some into settlements.

The re-establishment of the independent Estonian Republic in 1991 brought about several changes in the landuse and environmental policy. Big farms were liquidated and foundation of small farmsteads is under way. In 1883, the land structure was the following: arable land - 14%, meadows - 21%, heath (pastures) - 14%, mires - 21% (Kahk 1992). After one hundred years, arable land made up 25%, meadows 6.04%, forests 44.05%, water basins 2.00%, and others (including mires and brushwoods) 22.91% of Estonia's total area (Järv 1994). In the historical retrospective, during the last 100 years the arable area has increased by 78%, and forested area 68%, while the area under meadows and pastures has drastically decreased - 5.8 times. It leads to the conclusion that the number of cattle in the 19th century could have been even bigger than nowadys and the manuring of fields very extensive. At the beginning of 1994, there were 10,179 farms with a total area of 252, 200 ha. The average size of a farm was 24.8 ha, of that 43.2% were fields, 33.8% woodlands and 16.1% grasslands and shrubbery (Sein 1994). Compared with the beginning of the 19th century, the total number of farms has decreased 1.8 times. The centres of arable farming have also shifted. At the beginning of Medieval times, the arable farming was flourishing in northern Estonia (Henriku... 1993). Now the percentage of arable land is highest in the Tartu County (34.7%) and lowest in the Ida-Viru County (10.9%). The area of forest is largest on the Island of Hiiumaa (58.4%) and smallest in the Tartu County (38.8%).

The recent restructuring of economy has brought about several positive trends in the environmental management. The decrease in waste water and pollution load with efficiency of the new sewage treatment plants have improved the water quality in rivers (Pachel & Loigu 1995). But still some rivers are polluted with nutrients (Keila, Vääna, Selja, *etc.*), some of them even with heavy metals, phenols and oil products (Purtse, Kohtla). Especially large pollution load comes from the ash hills of the joint-stock companies Kiviter, Silmet, and Eesti Põlevkivi. The Kroodi Brook is a real waste water drain, receving untreated or poorly treated domestic and industrial waste water from the Maardu Settlement and from the joint-stock company Eesti Fosforiit.

As to air pollution, the situation is worst in the Kunda, Kohtla-Järve and Narva areas. The total emisson of pollutants into the atmosphere was highest in 1980. In 1994, it was only half as high, owing to the several international conventions Estonia has recently jointed and the better technology introduced. Still, in 1994 the volume of pollutants, emitted into the atmosphere, was 354,000 tonnes, including 161,500 tonnes of solid, 141,100 tonnes of SO_2 and 14,600 tonnes of NO_x (Saar 1995). Of the total air pollution load, about 50% comes from transport and about 50% from the other sources.

In conclusion, it should be pointed out that during half a century material-and-pollution-extensive large-scale industry and agriculture have caused serious environmental problems in northern Estonia. Exploitation of mineral recources and utilization of waste have deteriorated the groundwater quality, contaminated soils and air. Luckily, several problems which were topical during the last decades of Soviet power in agriculture, like those related to large farms and point pollution sources, huge fields and soil erosion, intensive use of agrochemicals, *etc.*, are gradually loosing their actuality.

The above shows that during the course of the historical times human impact has considerably increased and its character changed. Only one century ago Estonia was a typical agrarian country with well developed rural settlements and farms orientated to cattle rearing and crop farming. Roughly 20% of the territory was cultivated. To date, arable land forms *ca.* 25% of Estonia's area; nearly 75-80% of the territory has undergone land improvement. It means that anthropogenic stress on our environment has been tremendous. Hopefully, the implementation of the private-farm-system in agriculture will promote restoration of traditional nature management habits in Estonia.

X MINERAL RESOURCES

The Mineral Resources Classification System, worked out by the Estonian Commission on Mineral Resources, is built up on internationally accepted principles and is in good accordance with the last (1979) Classification of the UN Committee on Natural Resources. Category T corresponds to category R1, category R to R2, category P to R3 in the UN classification. The term "active" expresses the economic (E) status of resources; while the term "passive" marks their subeconomic (S) status.

In this chapter, the term "output" means the amount of mined out resources, the mining losses are not included.

KUKERSITE OIL SHALE

Introduction

The Baltic Oil Shale Basin (*ca* 50,000 km^2) is situated prevailingly in northeastern Estonia with a part of it extending eastward into Russia. Three well-explored oil shale deposits – Estonia, Leningrad and Tapa – occur within the oil shale basin. The boundary between the Estonia and Leningrad deposits is purely geopolitical and it runs along Estonia's border with Russia. Both deposits are currently mineable, while the Tapa oil shale deposit in central Estonia is considered to be a prospective one.

At present, the Estonia deposit is the largest commercially exploited oil shale deposit in the world; its total resources exceed 7 x 10^9 tonnes of oil shale. The resources of the prospective Tapa deposit are in order of 2.6 x 10^9 tonnes. The Estonia deposit has been mined continuously since 1919, with the maximum annual output of 29.7 million tonnes in 1980. Currently, kukersite oil shale is mined in six underground mines and in three open-cast pits. The total annual output is about 13 million tonnes.

The main oil shale (kukersite) sequence is of Middle Ordovician (Llandeilo - Early Caradoc) age. It contains up to 50 laterally continuous kukersite seams which can be traced at a length of 250 km in the eastwest direction. Basinward, however, they lose organic matter rather abruptly and are no longer distinguishable from the host limestone beyond 40-50 km.

Geological setting of the kukersite oil shale

Minor kukersite-type organic matter (OM) accumulations have been recorded throughout the whole Middle Ordovician time in the Baltic Oil Shale Basin (Мянниль 1966, Рыымусокс 1970, Пылма 1982). The major kukersite OM accumulation, however, took place during early Middle Ordovician time (Llandeilo - Early Caradoc, Männil 1990) in an area presently known as the Baltic Oil Shale Basin which is located on the southern slope of the Fennoscandian Shield. The kukersite OM accumulated in an area larger than 50,000 km^2 extending

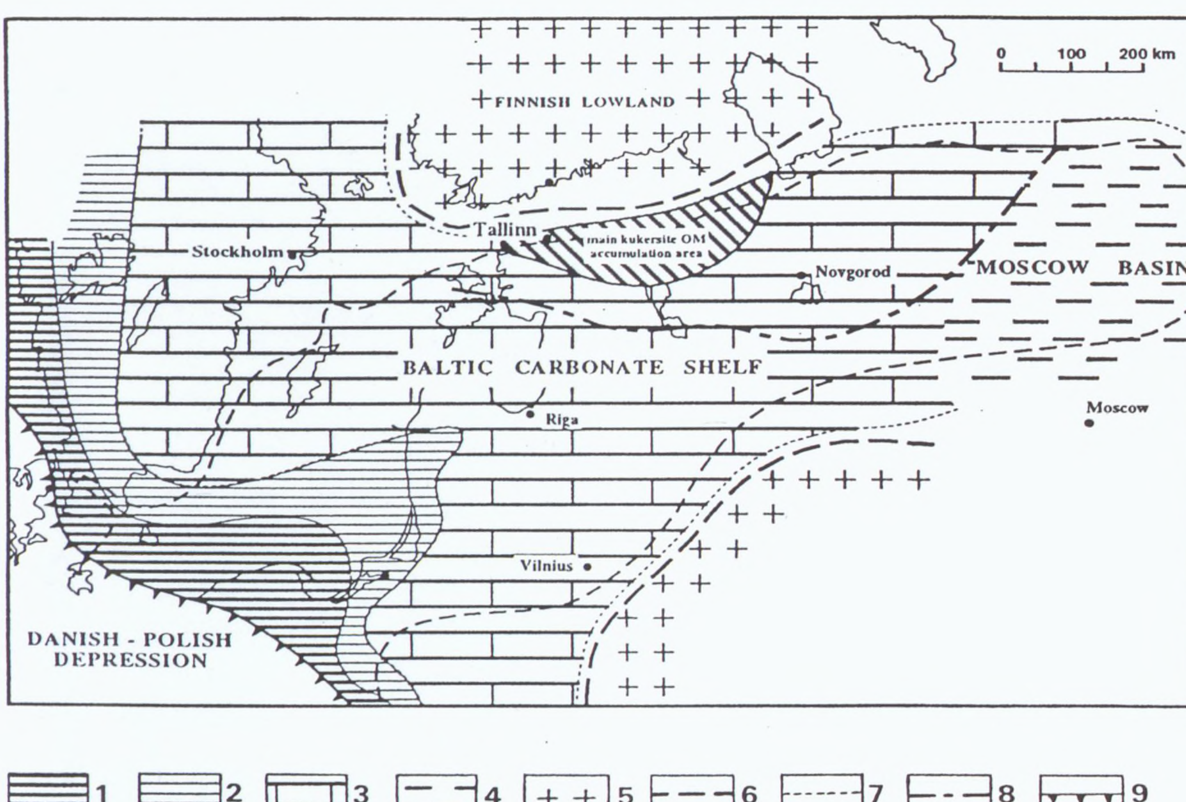


Fig. 202. Palaeogeographic reconstruction of Baltoscandia for *Nemagraptus gracilis* time (Middle Ordovician). Modified from Männil *et al.* (Мяннил и др.1986): 1 - black graptolitic shale; 2 - calcareous marl; 3 - biomicritic limestone, variably argillaceous; 4 - clayey marl; 5 - emerged areas; 6 - supposed margin of sedimentation area; 7 - present-day margin of *Nemagraptus gracilis* age rocks; 8 - margin of kukersite OM accumulation area; 9 - Tornquist lineament.

from western Estonia to St. Petersburg in Russia. In its present outline, the main kukersite accumulation area resembles an elongated crescent (Fig. 202) and apparently represents the southern and central parts of the original deposit. The actual shape and the northward extent of the original depositional basin remains unknown because of the post-Devonian erosion and the formation of the Gulf of Finland (Fig. 203). However, palaeogeographical reconstructions (Мянниль 1966, Мянниль и др.1986) and lithological studies (Пылма 1982, Мянниль и Бауерт 1986, Bauert 1989) indicate that the kukersite OM accumulated in a shallow subtidal area of the Palaeobaltic Sea bordered by the Finnish Lowland in the north.

The Baltic Oil Shale Basin is administratively divided into the Estonian andLeningrad regions by the boundary coinciding with the border between Estonia and Russia (Fig. 203). The northern boundary of the Estonia deposit is erosional, while the western and southern boundaries are drawn by the farthest appearance of discrete kukersite seams in the oil shale sequence. The deposit boundary is arbitrarily drawn within the limits of the following parameters: thickness of the mineable bed > 1.4 m and calorific value > 6.1 MJ/kg.

Estonia oil shale deposit

The Estonia kukersite oil shale deposit is located in northeastern Estonia. In terms of the areal extent, it is the largest deposit in the Baltic Oil Shale Basin. Besides, it contains the highest quality oil shale and the highest volume of mineable oil shale.

The wedge-shape deposit has an east-westward length of about 130-140 km. South of the northern erosional boundary, the extent of the deposit increases eastward; reaching about 40-45 km between Lake Peipsi in the south and the erosional boundary in the north (Fig. 203). The area of the Estonia oil shale deposit is about 3000 km^2.

Stratigraphy

The kukersite oil shale beds within the Estonia deposit are lithostratigraphically confined to the Kõrgekallas and Viivikonna formations (Figs. 204, 205). Both formations have been divided into several members, of which the Pärtliorg, Erra, Kiviõli, Maidla and Peetri members contain kukersite beds. Isopach maps of the Kiviõli, Maidla and Peetri formations are given in Fig. 206.

The currently accepted stratigraphic nomenclature of the kukersite beds found in the Kõrgekallas and Viivikonna formations follows the historical tradition of more than 60-year-long study by various authors and is not uniform throughout the sequence. The general rule here is that the kukersite beds in the Pärtliorg and Erra members have been indexed with lowercase letters (Fig. 204). Capital letters have been used to designate kukersite beds in the Kiviõli Member and in the lower part of the Maidla Member, whereas for the rest of the Viivikonna Formation, Roman numerals have been used.

The current commercial (mineable at the time of the determination) kukersite oil shale bed is a composite of individual kukersite seams A - F$_1$, comprising also some limestone interbeds (Photo 63). Stratigraphically, it forms the lowermost part of the Kiviõli Formation.

Historical background of oil shale exploration and mining

The first notes on kukersite oil shale were already released by A. W. Hupel in 1777. The first scientific paper on this specific fossil fuel was published by J. Georgi in 1791 (for review see Aaloe 1989).

The stratigraphic position for the kukersite oil shale formation was first established by Schmidt (1858, 1881) who carried out his field observations in the vicinity of the former manor at Kukruse, northeastern Estonia. The name **kukersite** is derived from the name of this locality.

The need for locating new sources of fossil fuel was responsible for launching a wide exploration for oil shale deposits in an area along railroad tracks between the towns of Jõhvi and Rakvere in northern Estonia in 1916. The results of chemical and technological studies showed the promise of

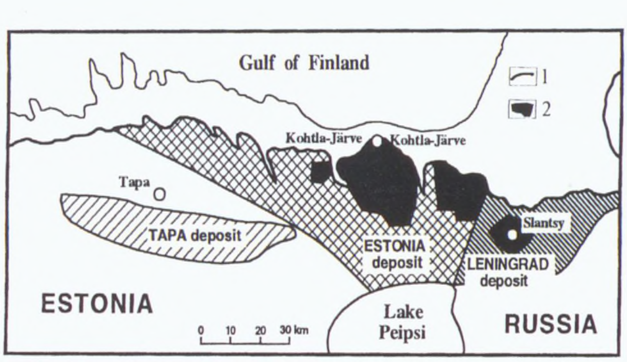


Fig. 203. Location of oil shale deposits in the Baltic Oil Shale Basin: 1 - Recent erosional boundary of kukersite oil shale; 2 - mined out areas and fields of active mines.

Photo 63. Aidu quarry. Section of the commercial bed. A, B, C, D, E, F$_1$ and F$_2$ are the kukersite oil shale beds. A/B, C/D, *etc.* are limestone interbeds. *Photo by V. Piir.*

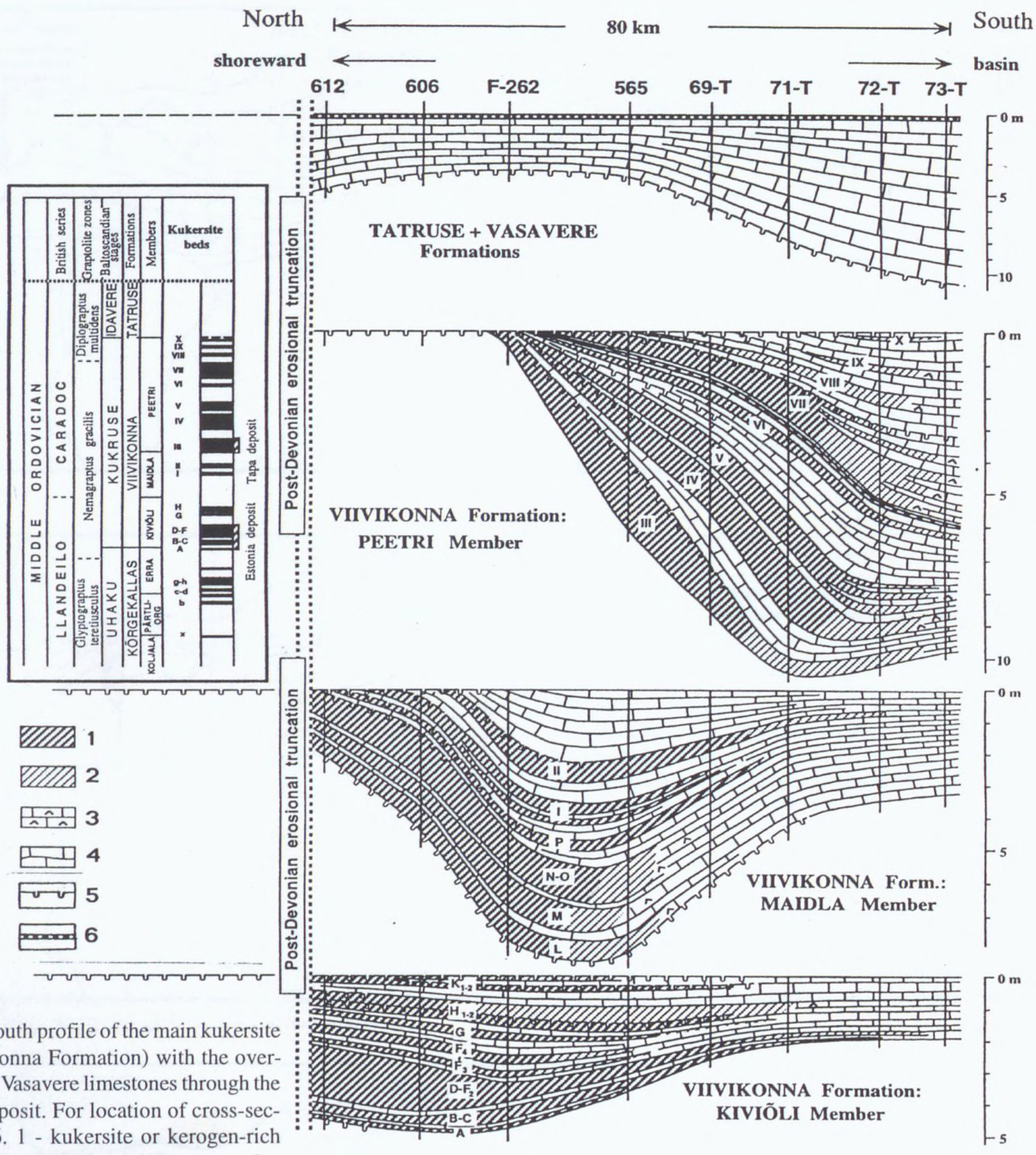


Fig. 204. North-south profile of the main kukersite sequence (Viivikonna Formation) with the overlying Tatruse and Vasavere limestones through the Tapa oil shale deposit. For location of cross-section see Fig. 206. 1 - kukersite or kerogen-rich limestone; 2 - kerogen-lean limestone; 3 - biomicritic limestone with a low content of organic matter; 4 - variously argillaceous limestone; 5 - hardground; 6 - K-bentonite bed.

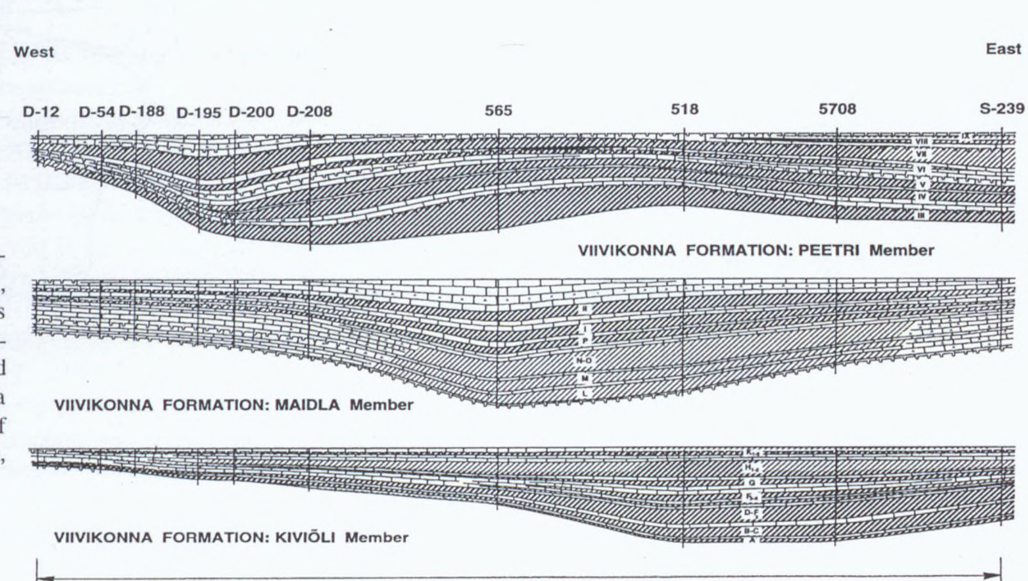


Fig. 205. East-west cross-section of the Kiviõli, Maidla and Peetri members of the Viivikonna Formation through the Tapa and central part of the Estonia deposits. For location of cross-sections see Fig. 206, for legend Fig. 204.

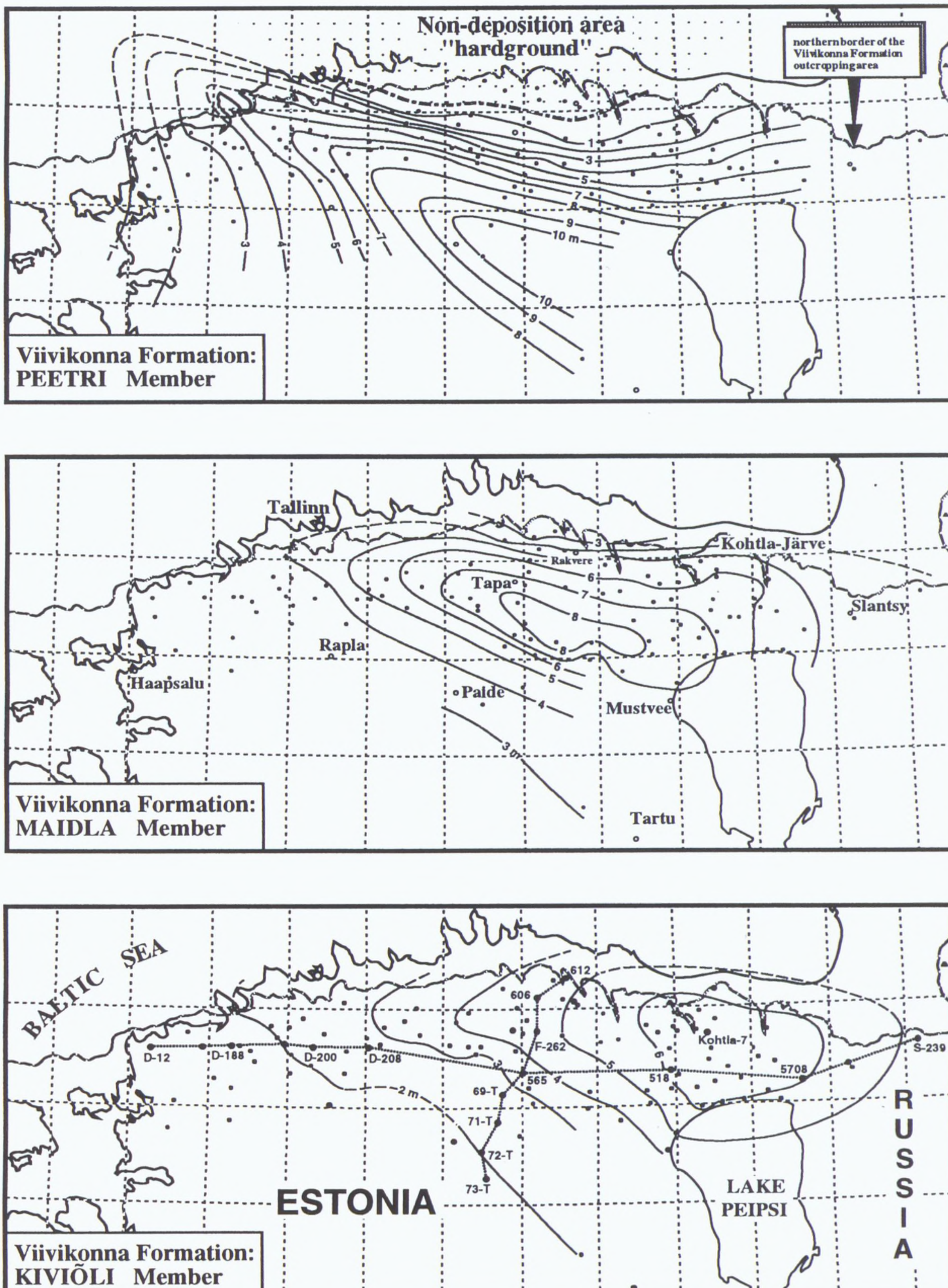


Fig. 206. Thickness maps of the Kiviõli, Maidla and Peetri members of the Viivikonna Formation in northern and central Estonia with locations of N-S and E-W cross-sections.

kukersite oil shale as a raw material for the chemical industry (Погребов 1919). Openpit mining was soon initiated at Pavandu (1919) and Vanamõisa (1920) and underground mining at Kukruse (1920) and Käva (1924). In the following years (Fig. 207), both Estonian companies and foreign capital launched several new opencast pits at Kiviõli (1922), Küttejõud (1925), Ubja (1926), Viivikonna (1936) and Kohtla (1937).

The first oil shale mines were located in areas where kukersite oil shale was either exposed on the ground or occurred rela-

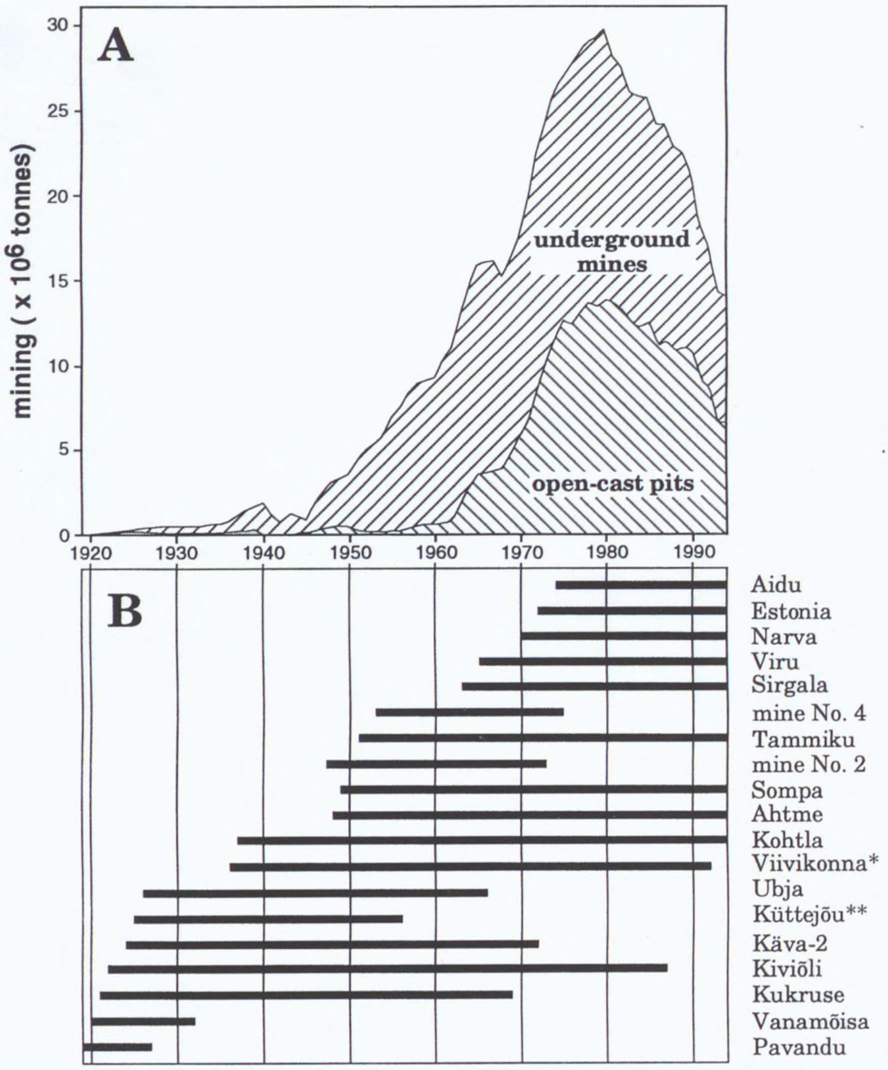


Fig. 207. Output of kukersite oil-shale in Estonia since 1919 (A), including list of all present and past open-cast pits and underground mines (B). Black bars show the years of activity.

*The Viivikonna open-cast pit was incorporated with the Sirgala open-cast pit.
**The Küttejõu open-cast pit was incorporated with the Kiviõli Mine.

tively close to the surface. As all kukersite seams are gently dipping southwards, underground mining could be introduced in areas where the thickness of overburden exceeded 5 - 8 metres. By 1940, the total output of oil shale had reached about 11 x 10^6 tonnes and the annual output 1.7 x 10^6 tonnes.

A new epoch in kukersite oil shale exploration started immediately after World War II when exploration works covered most of northeastern Estonia. In 1947, the explored oil shale resources reached 0.9 x 10^9 tonnes and in 1960 already 3.3 x 10^9 tonnes. Opening of several new mines (Ahtme - 1948; Sompa and No. 2 - 1949; Tammiku - 1951; No. 4 - 1953) brought about a rapid increase in an annual mining output (Fig. 207).

Launching of two large oil shale-feeded power plants - the Baltic Thermal Power Plant in 1966 and the Estonia Thermal Power Plant in 1973, increased the demand for kukersite oil shale as fuel. Therefore, several new open-cast pits (Sirgala - 1963, Narva - 1970, Aidu - 1974) and underground mines (Viru - 1965, Estonia - 1973) were taken into use. Annual oil shale output continued to increase rapidly (Fig. 207).

Launching of a nuclear power plant at Sosnovyi Bor in the Leningrad District, Russia, lowered the need for electricity produced in the oil shale-based power plants. As a result, the annual output of oil shale has steadily decreased during the last decade: from 25.7 x 10^6 in 1985 and 21.2 x 10^6 in 1990 to 13.0 x 10^6 tonnes in 1996.

Characterization of the Estonia deposit subdivisions

According to the degree of exploration, thickness and quality of the commercial bed and its depth from the surface, the Estonia deposit has been arbitrarily subdivided into five areas: central, eastern, western, northwestern, and southern (Fig. 208).

The c e n t r a l area of the Estonian deposit covers about

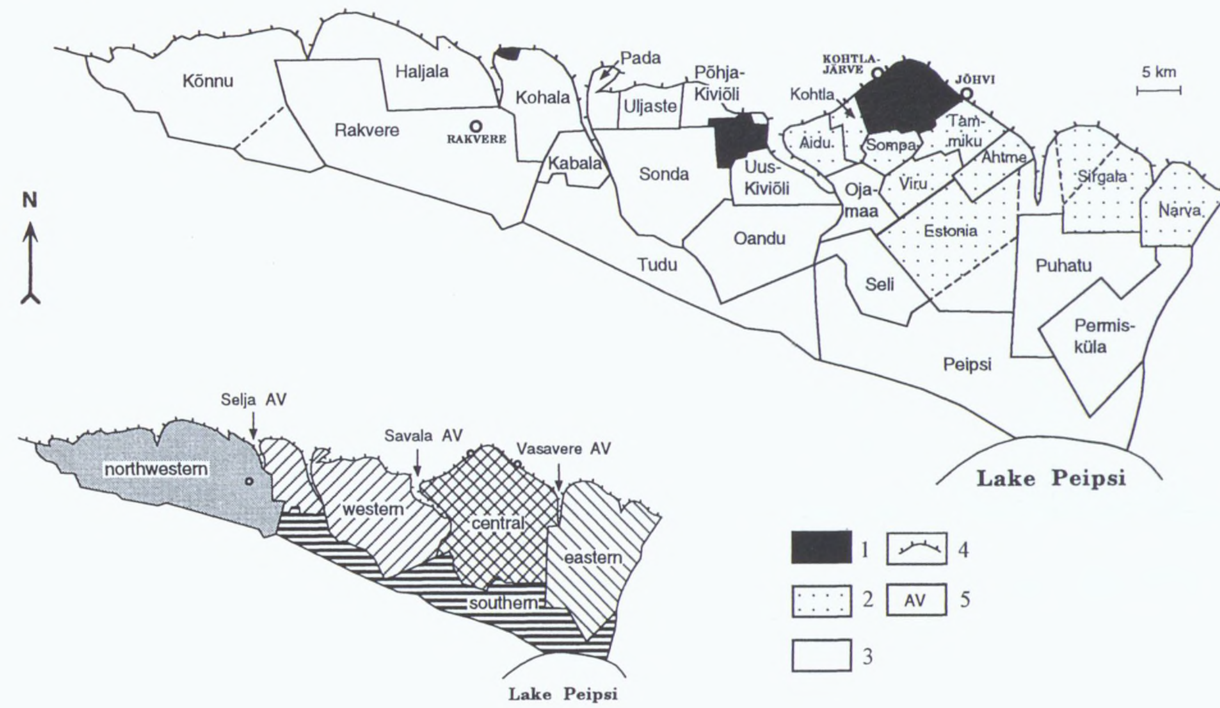


Fig. 208. Subdivisions of the Estonia deposit: 1 - closed mines; 2 - fields of active mines and open-cast pits; 3 - explored fields for further mining; 4 - recent erosional boundary of the kukersite oil shale formation; 5 - ancient valleys.

660 km². Its eastern and western boundaries are limited by the Vasavere and Savala ancient valleys, respectively. Both valleys are filled up with Upper Pleistocene and Holocene sediments and extend southward from the northern erosional boundary of the Viivikonna Formation deposits. In this area, the commercial kukersite seams A-F_1 are at their thickest. Several mines in the northernmost part of the central area, including Kukruse, Käva-2, mines No. 2 and No. 4 with the area totalling 120 km², have already been closed down. Currently, six underground mines (Kohtla, Sompa, Viru, Tammiku, Ahtme and Estonia) and one open-cast pit (Aidu, Photo 64) operate in this area. Within the limits of the currently operating mines, kukersite oil shale has already been mined out in an area of about 180 km². Therefore, further mining can be developed in an area of about 360 km² with recently revised total resources of about 1.2 x 10^9 tonnes. As oil shale mining is not permitted within the limits of the Seli and Ojamaa nature reserves, the mineable (active) resources account only for 0.7 x 10^9 tonnes.

The e a s t e r n area of the Estonia deposit extends from the Vasavere ancient valley eastwards up to the Narva River (Fig. 208) and holds the currently operating Sirgala and Narva open-cast pits and the well-explored prospective Puhatu and Permisküla mining fields. The total extent of the eastern area is about 530 km², of which about 90 km² has been mined out already. The mining conditions are similar to those in the central area and the mineable oil shale quality is only slightly lower than in the center of the Estonia deposit. The estimated oil shale resources reach 1.3 x 10^9 tonnes (incl. 0.5 x 10^9 tonnes of oil shale within the limits of nature reserves).

The w e s t e r n area of the Estonia deposit is located between the Savala and Selja ancient valleys (Fig. 208). Of the total of 620 km², about 30 km² have been mined out. There are no currently operating mines in this area, however, it is considered as the most prospective area for launching new open-cast pits and underground mines in the near future. The total oil shale resources are estimated at about 1.7 x 10^9 tonnes. Because of several environmental restrictions, only 0.5 x 10^9 tonnes of oil shale are mineable todate.

The n o r t h w e s t e r n area (590 km²) is restricted to the westernmost part of the Estonia deposit. The oil shale resources are estimated at 1.5 x 10^9 tonnes. Since the oil shale quality parameters are rather low, particularly due to the low OM content and the diminished thickness of mineable seam, this is regarded as a low-prospective area for future oil shale mining.

The s o u t h e r n area (570 km²), like the northwestern one, has a little prospective to become an oil shale mining area. The main reasons are the gradual southward increase of overburden thickness and the lowering of calorific value in oil shale seams. The exploration estimates suggest about 1.3 x 10^9 tonnes of oil shale resources for the southern area of the Estonia deposit.

Geology of the commercial bed A-F_1

The commercial bed A-F_1 consists of seven indexed kukersite seams (from the bottom: A; A'; B; C; D; E; F_1) alternating with five host limestone interbeds (A'/B; B/C; C/D; D/E; E/F_1). Among host limestone interbeds, the beds A'/B and C/D are almost pure limestones with a low percentage of OM while the other limestone interbeds (B/C; D/E; E/F_1) are enriched in OM (Fig. 209).

The lithology of this interbedded sequence for the whole Kiviõli Formation, as exposed in the wall of the Kohtla open-cast pit (central part of the Estonia deposit) is depicted in Figure 209. As is seen, the role of kerogenous limestone nodules varies within kukersite oil shale beds. The thickness of individual elongated nodules and kerogen-rich nodular

Photo 64. Opencast mining of oil shale in the Aidu quarry. *Photo by V. Piir.*

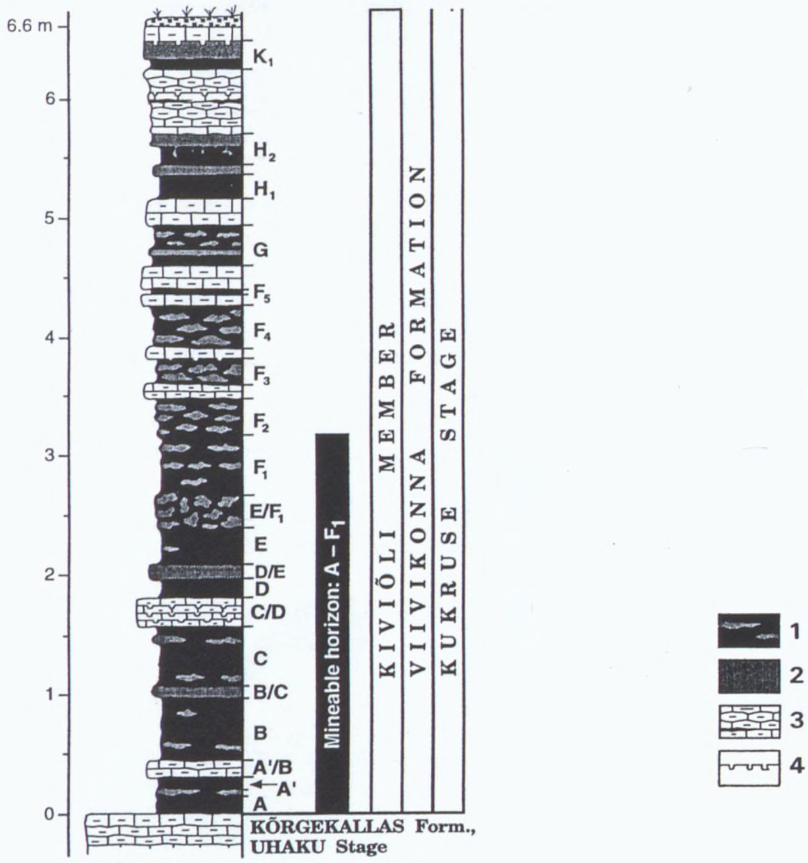


Fig. 209. Typical lithology of the Kiviõli Member beds at the Kohtla open-cast pit (modified from Kõrts & Einasto 1990): 1 - kukersite with kerogenous limestone nodules; 2 - kerogenous limestone; 3 - variably argillaceous limestone with kukersite lenses; 4 - hardground.

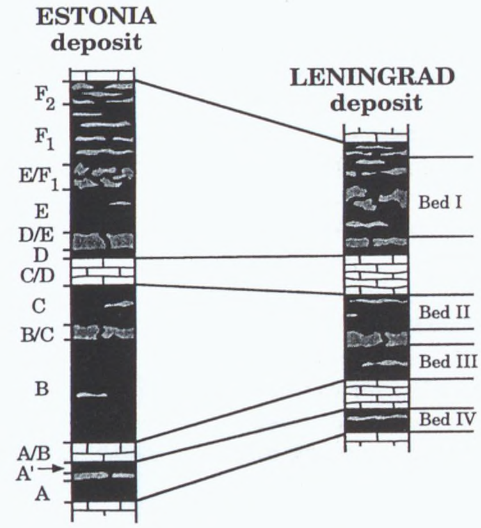
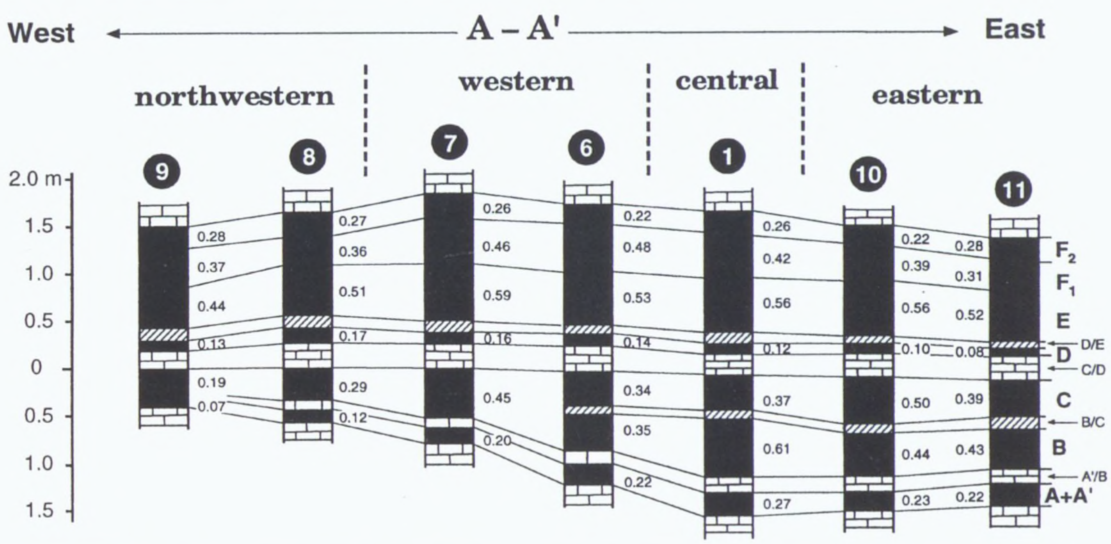


Fig. 209a. Nomenclature and correlation of mineable beds between the Estonia and Leningrad deposits. For legend see Fig. 209.

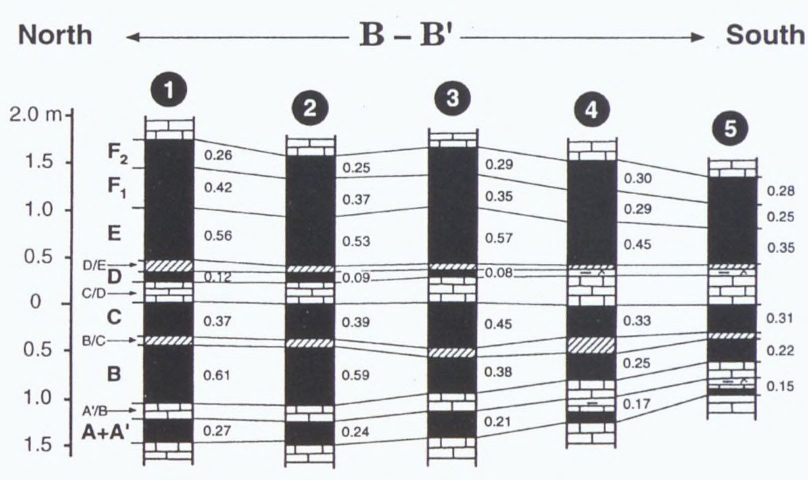
interbeds may reach 15 cm. The content of such nodules may reach 50-60% in some oil shale seams (F_1). An increase of the percentage of nodules, accompanied with the decrease of OM, is observable towards the western margin of the basin. The maximum thickness of oil shale beds may reach 60-70 cm, while limestone interbeds can be as thick as 30-35 cm.

It should be pointed out that all kukersite seams retain most of their lithological characteristics (degree of bioturbation, content and shape of nodules, *etc.*) throughout the whole Estonia deposit. The changes in lithology occur only in the western and southern peripheral parts of the deposit where these are caused either by changes in the composition (the clay content increase in the south and southeast direction; increase in the role of limestone nodules in the west direction) or by gradual thinning out of seams (Figs. 204, 205, 210). Abrupt changes in synsedimentary accumulation of kukersite organic muds or carbonate muds within the Estonia deposit have not been recorded (Каттай 1990). However, post-depositional karstification in certain areas, mainly along linear zones of tectonic disturbances, has resulted in the formation of karst clays (Каттай 1986).

The commercial bed A-F_1 is at its thickest (2.7-3.0 m) in the northern part of the central and eastern areas of the Estonia deposit (Fig.211). The thickness decreases gradually both southward and westward down to 1.8-2.0 m and 1.3-1.5 m, respectively. The calculations suggest that southward the thinning out is twice as fast (3-5 cm/km) as westward (1-3 cm/km).



A - F_1 (m)	1.57	1.98	2.40	2.62	2.93	2.84	2.56
Σ kukersite beds	1.20	1.45	1.86	2.02	2.35	2.22	1.95



A - F_1 (m)	2.93	2.77	2.71	2.37	2.06
Σ kukersite beds	2.35	2.21	1.84	1.56	1.27

Fig. 210. West-east (A-A') and north-south (B-B') cross-sections of the commercial bed A-F_1 through the Estonia deposit. Kukersite bed F_2 is not mined because of its low organic matter content. Location of both cross-sections is shown in Fig. 211. 1 - Kukersite oil shale; 2 - kerogenous limestone; 3 - variably argillaceous limestone; 4 - kerogenous marlstone.

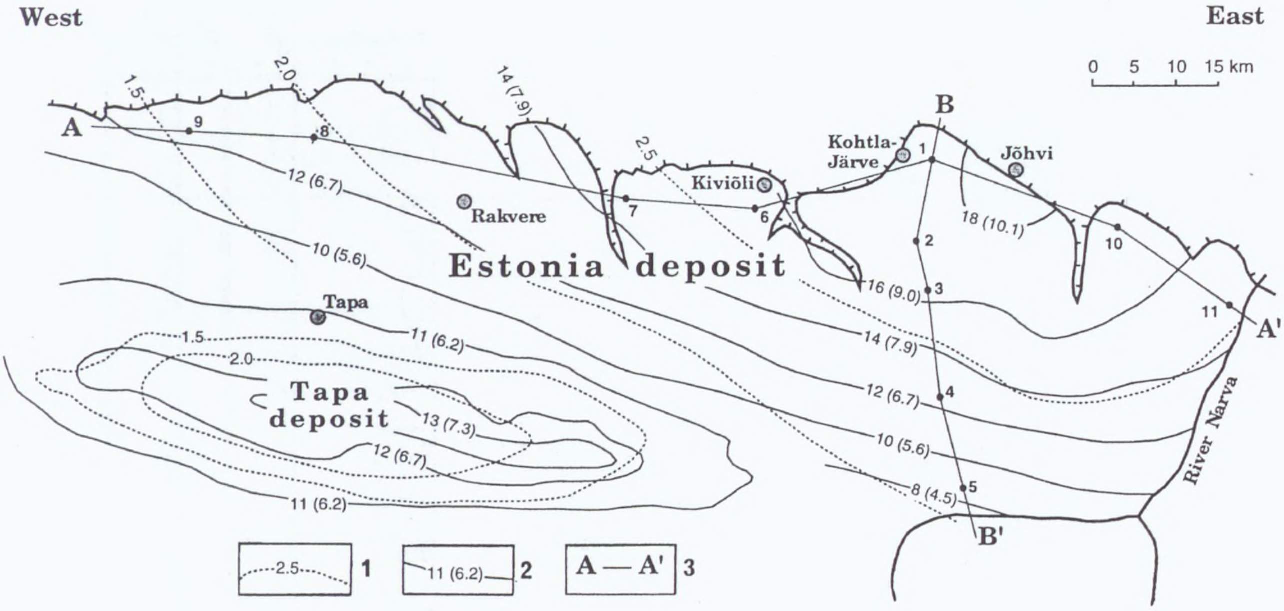


Fig. 211. Characteristics of the commercial bed A-F_1 of the Estonia deposit and the mineable bed III of the Tapa deposit: 1 - Thickness isolines for commercial bed A-F_1 (Estonia deposit), II kukersite oil shale bed (Tapa deposit); 2 - oil yield by Fisher Assay (T, %) and calorific value (Q, MJ/kg) isolines (calorific values are given in brackets); 3 - location of cross-sections shown on Fig. 210.

The total thickness of the kukersite oil shale seams in the commercial bed A-F_1 is the following: 2.0-2.4 m in the northern part of the central area of the Estonia deposit, 1.2-1.4 m in the marginal southern and 0.8-1.0 m in the marginal northwestern part. The total thickness of host limestone beds is rather uniform (0.5-0.6 m in the central, 0.4-0.5 m in the marginal northwestern and 0.7-0.8 m in the southern area). Evidently, the factors favouring the accumulation of organic matter were different from those responsible for carbonate sedimentation.

Mineralogy and geochemistry of kukersite oil shale

The kukersite seams contain OM, carbonate and clastic materials in various proportions (Fig. 212). The carbonate material consists mainly of pore-filling micritic carbonate mud together with a variable content of fine to coarse skeletal debris. The carbonate content ranges from 20-70% in different kukersite seams (Аалоэ А.О. 1986, Kattai & Puura 1988). In general, the carbonate content increases mostly westward and also slightly southward (balancing a diminishing OM content). The main carbonate mineral is calcite, while dolomite usually remains within 1-15% limits. The content of dolomite is higher in the northern part of the central area and also in the eastern area.

The semiquantitative X-ray diffractometry and thin section examination have revealed that the clastic matrix is composed mainly of silt-size quartz and illite (Вингисаар и др. 1984). The minor clastic minerals are feldspars and chlorite (Fig. 213). Besides, pyrite is a rather common authigenic mineral in kukersite oil shale. The content of clastic minerals varies from 15-60%.

Kukersite OM may occur as dispersed in host limestone or as concentrated into individual lenses and seams. The OM consists mostly of accumulations of the microfossil *Gloeocapsomorpha prisca* Zalessky 1917 (Kõrts 1992, Kõrts & Veski 1994). Based on apparent morphologic similarities with extant species, *G. prisca* has been previously assigned either to the cyanobacteria or to the Chlorophyta. Recent morphological and biogeochemical studies (Foster *et al.* 1989) have suggested that *G. prisca* has strong similarities to the modern cyanobacterium *Entophysalis major* which is an intertidal to very shallow subtidal, mat-forming cyanobacterium. Modern occurrences of *E. major* are known from Shark Bay, western Australia, where it forms widespread mats (Logan *et al.* 1974). It has also been reported from the hypersaline Laguna Mormona, California (Stuermer *et al.* 1978).

The OM content varies significantly with kukersite seams, ranging from 10-15 to 50-60% in seams B and E in the northern part of the central area of the Estonia deposit. This is the

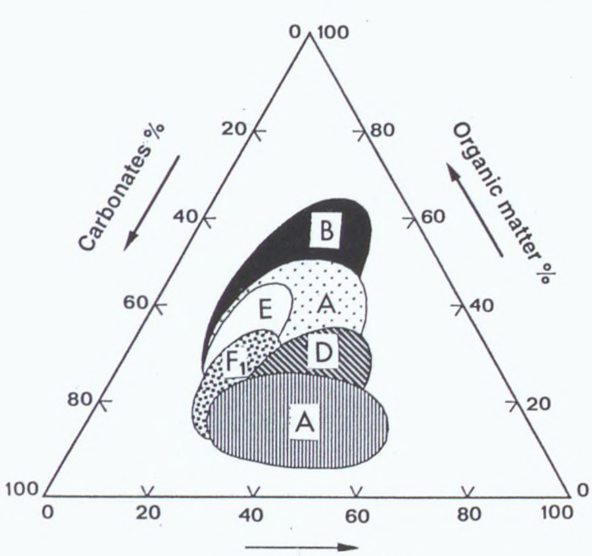


Fig. 212. Composition of oil shale seams in the commercial bed A-F_1 of the Estonia deposit (Kattai & Puura 1988).

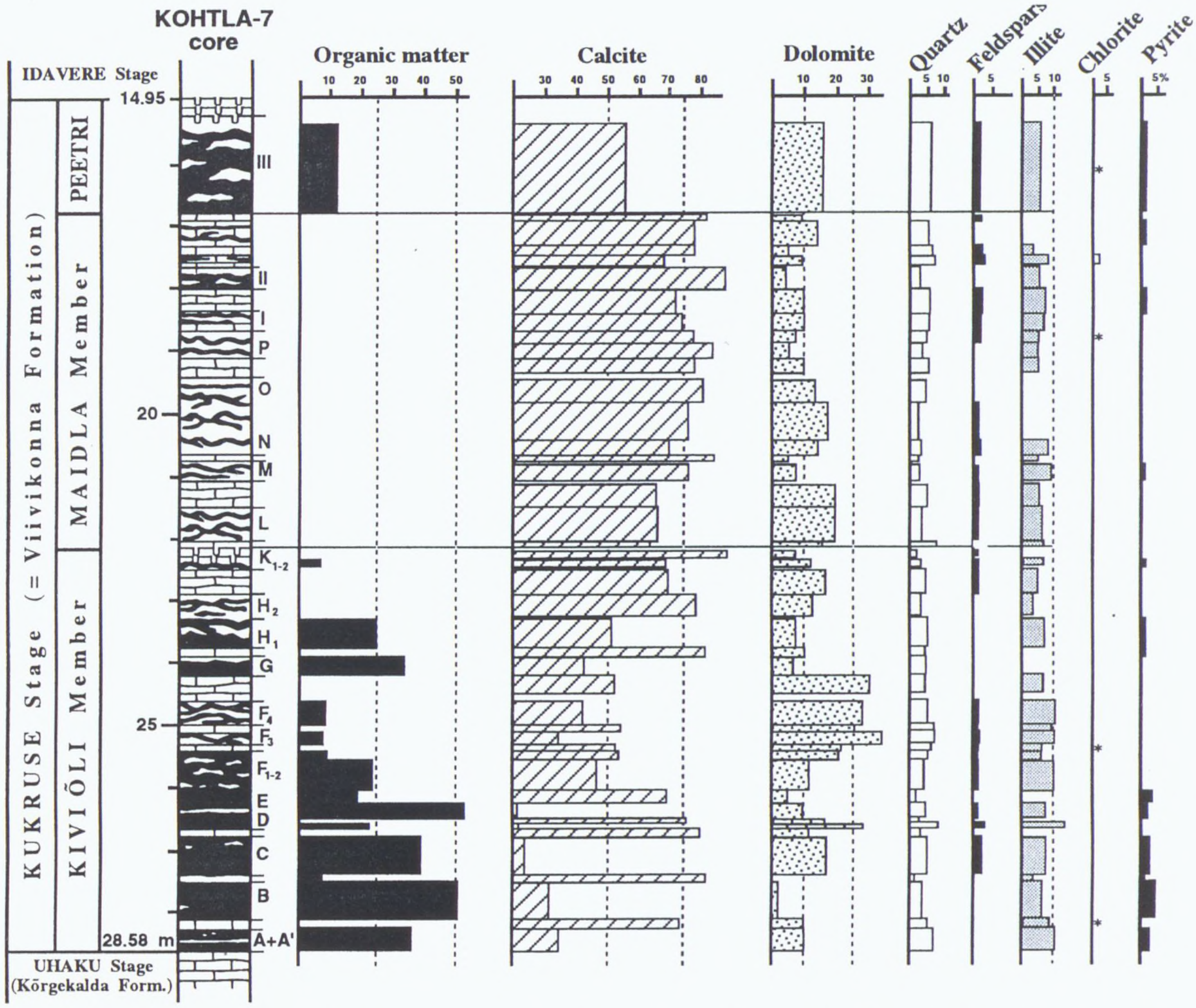


Fig. 213. Semiquantitative X-ray diffractometry data on the mineral composition of the Viivikonna Formation carbonate rocks and oil shale beds. The borehole Kohtla-7 was drilled in the central area of the Kohtla mine; for location of the mine see Fig. 208. Asterisks denote traces of chlorite.

area with the maximum total thickness of kukersite beds in the commercial bed A-F_1. The kukersite OM is characterized by a very low bitumen content and the kukersite kerogen has the following elemental composition (Уров и Сумберг 1992): C = 77.3%; H = 9.8%; O = 10.8%; S = 1.7 and N = 0.4%. Pyrolysis studies (Dyni *et al.* 1989, Foster *et al.* 1989) indicate that kukersites have a significant hydrocarbon potential (S_1 & S_2 = 300-350) and are characterized by a high hydrogen index (HI = 675-960). These hydrogen and oxygen index data suggest the prevalence of Type I kerogen in kukersite OM.

Very little is known about the minor and trace element abundances in kukersite oil shale and interbedded limestones. However, preliminary instrumental neutron activation analysis (INAA) data from a few samples of the Baltic Thermal Power Plant feedstock oil shale (Пец и др. 1985) indicate that the Middle Ordovician kukersite oil shale is not enriched in heavy metals (Table 50), whereas the Cambrian organic-rich alum shale in Sweden and the Lower Ordovician *Dictyonema* argillite in northern Estonia are known to contain these elements in highly elevated concentrations (Andersson *et al.* 1985, Пукконен 1989). At the same time, the INAA analysis revealed high concentrations of Cd, As and Cl relative to the concentrations in an average shale (Виноградов 1962) and average carbonate rocks (Turekian & Wedepohl 1961). Concentrations of major elements in the ashed oil shale samples from various parts of the Estonia oil shale deposit vary within the following range (average value is given in brackets):

SiO_2:	20-50% (30%)	K_2O:	2-6% (3.5%)
Al_2O_3:	5-15% (9.5%)	Na_2O:	0.1-0.5% (0.2%)
Fe_2O_3:	2-9% (5.5%)	TiO_2:	0.2-1.2% (0.7%)
CaO:	30-60% (41.5%)	SO_3:	3.0-6.5% (4.5%)
MgO:	1-6% (3%)		

Calorific value and Fisher Assay data for kukersite oil shale

At present, slightly over 80% of mined oil shale is directly burnt as a pulverized fossil fuel in power plants, while the chemical industry uses about 15% of mined oil shale for shale oil retorting. A minor oil shale user (about 3%) is also the cement industry.

A calorific value is considered as a main characteristic for fossil fuels. The lower limit for a calorific value of feedstock oil shale in power plants is set for a range 7.8-8.8 MJ/kg, whereas the moisture content must be <14% (equals to 10.2-11.5 MJ/kg for dry oil shale).

Table 50. Contents of chemical elements by INAA (Пец и др. 1985) in a feedstock kukersite oil shale of the Baltic Thermal Power Plant (= average for oil shale seams A-F_1)

Element	Content in kukersite oil shale by INAA (ppm)	Ef - clays	Average content in shales and clays (ppm)	Ef - lime-stones	Average content in limestones (ppm)
Cl	1850	11.6	160	12.3	150
Sc	4.2	0.4	10	4.2	1
V	28	0.2	130	1.4	20
Cr	37.5	0.4	100	3.4	11
Co	2.9	0.1	20	29	0.1
Ni	21	0.2	95	1	20
Cu	17	0.3	57	4.3	4
Zn	48.7	0.6	80	2.4	20
Ga	8	0.3	30	2	4
As	7.6	1.2	6.6	7.6	1
Se	<0.1	<0.2	0.6	<1	0.08
Br	100	16.6	6	16.1	6.2
Rb	39	0.2	200	13	3
Sr	151	0.3	450	0.2	610
Y	5.4	0.2	30	0.2	30
Zr	49	0.2	200	2.6	19
Nb	4.5	0.2	20	15	0.3
Mo	3	1.5	2	7.5	0.4
Cd	4	13.3	0.3	114	0.035
Sb	0.5	0.2	2	2.4	0.2
Ba	140	0.2	800	14	10
Ce	22	0.4	50	1.9	11.5
Yb	0.8	0.3	3	1.6	0.5
Hf	1.8	0.3	6	6	0.3
W	0.6	0.3	2	1	0.6
Hg	0.1	0.3	0.4	2.5	0.04
Pb	23.5	1.2	20	2.6	9
Th	3.4	0.3	11	2	1.7
U	2.9	0.9	3.2	1.3	2.2

Average concentrations in shales and clays are from the paper by Vinogradov (Виноградов 1962) and average concentrations in limestones are from Turekian and Wedepohl (1961). E_f- enrichment/impoverishment factor: relative to average clays/shales and relative to average limestones. Shaded rows show elements which have E_f-shales>1.

For retorting crude oil from kukersite oil shale, crushed oil shale (diameter of particles 25-125 mm) with higher OM content (estimated Fisher Assay 23-25%; = calorific value 13-14 MJ/kg) is used. There is a clear relationship between a calorific value (Q) and an oil yield by Fisher Assay (T), where:

T = 1.78Q (with a correlation coefficient - 0.975).

Studies have also shown that a 10% increase of OM in the kukersite oil shale, will raise a calorific value by 3.7 MJ/kg and will increase an oil yield 6.7% (Каттай 1991). The sulfur content is relatively low in kukersite oil shale (average 1-3%) and therefore, sulfur is not regarded as a strictly limiting factor for oil shale utilization.

The highest quality oil shale is mined in the northern part of the central area of the Estonia deposit. The following data characterize an average calorific value and the Fisher Assay results for the whole commercial bed (A-F_1) in this area:

A) Commercial bed (limestone interbeds included):
- calorific value: 9.0-10.0 MJ/kg;
- Fisher Assay: 16-18%.

B) Commercial bed (limestone interbeds excluded):
- calorific value: 12.6-13.4 MJ/kg;
- Fisher Assay: 21-23%.

The quality of kukersite oil shale gradually decreases toward peripheral parts of the deposit. Along the southern boundary, the commercial bed is characterised by the following values:

A) Commercial bed (incl. limestone interbeds):
• calorific value: 4.0-5.0 MJ/kg;
• Fisher Assay: 7-9%.

B) Commercial bed (limestone interbeds excluded):
• calorific value: 7.5-8.5 MJ/kg;
• Fisher Assay: 13.4-15.1%.

In the marginal north-west, both the average calorific value and oil yield values have lowered:

A) Commercial bed (limestone interbeds included):
• calorific value: 6.0-7.0 MJ/kg;
• Fisher Assay: 10.7-12.5%.

B) Commercial bed (limestone interbeds excluded):
• calorific value: 8.4-9.2 MJ/kg;
• Fisher Assay: 15.0-16.4%.

It should also be noticed that both oil shale characteristics for individual oil shale seams vary significantly both seam by seam and laterally within the deposit: the calorific value ranges from 4-5 to 20-21 MJ/kg and the oil yield from 6-8 to 35-37%.

The lower limit set for a calorific value for both a power plant feedstock oil shale (10.2-11.5 MJ/kg for dry oil shale) and a crude oil retorting shale (calorific value 13-14 MJ/kg) is higher than a calorific value for most individual kukersite seams in a mineable group of seams. Therefore, either a selective mining or a later beneficiation of mined oil shale must be involved. As the density of kukersite oil shale (1.3-1.8 g/cm^3) differs notably from the density of limestone (2.0-2.5 g/cm^3), a flotation-based beneficiation of mined oil shale is commonly used.

Mining obstructions and hydrogeological conditions

The Ordovician-Silurian succession of carbonate rocks which includes Middle Ordovician kukersite oil shale is resting on a slightly southward dipping slope of the Fennoscandian Shield. As a result, the kukersite oil shale outcropping along the northern erosional cut, is overlain by about 100-120-m-thick overburden in the southern marginal part of the Estonia deposit (Fig. 214). The overall simple geological structure of the oil shale deposit is somewhat complicated by erosional cuts of ancient valleys, linear zones of tectonic disturbances, karst, dome-like structures and shallow depressions, which together with varying hydrogeological conditions may render oil shale mining in some areas impractical.

The Kunda, Savala, Vasavere, Selja and other **ancient valleys** (Fig. 208) are deep (10-20 to 60-70 m) erosional cuts formed during the pre-Quaternary in Lower Palaeozoic rocks. Those valleys are hundreds of metres to 1.5-2.0 km wide and may extend 10-25 km southward into the Baltic Oil Shale Basin. Northward, they reach the Gulf of Finland. Ancient valleys are filled up with glacial, glaciofluvial and glaciolacustrine deposits. Those ancient valleys cut either partly or completely through kukersite oil shale seams and form natural boundaries which subdivide the Estonia oil shale deposit into main oil shale mining areas (Fig.208).

Several tens of kilometres long but rather narrow (1-4 km) zones of **tectonic disturbances** (Aseri, Ahtme, Viivikonna, *etc*., Fig. 214) extend down to the crystalline basement and split the homocline into several blocks of different size. The vertical displacement of oil shale beds ranges between 5-25 metres (Вахер и др. 1962, Пуура и др. 1986). All rocks within these zones are highly fractured and often transferred into karstic clays (Каттай и Вингисаар 1980).

Besides being common along long linear zones of tectonic disturbances, a phenomenon of oil shale alteration to karstic clays is also observed in the areally limited northeast-extending (45-75°) fracture zones. The width of such **karstification zones** ranges from several tens of metres to 100-200 m, whereas the residual karstic clay zones are less than 10 m wide. Usually, such karstified zones are hundreds of metres, occasionally even 3-4 kilometres long.

Several areally limited (areas ranging from hundreds of square metres to 10-12 km^2) **dome-like structures** as well as shallow oval or **ring-shaped depressions** have been established in the northern area of the Estonia deposit (Каттай

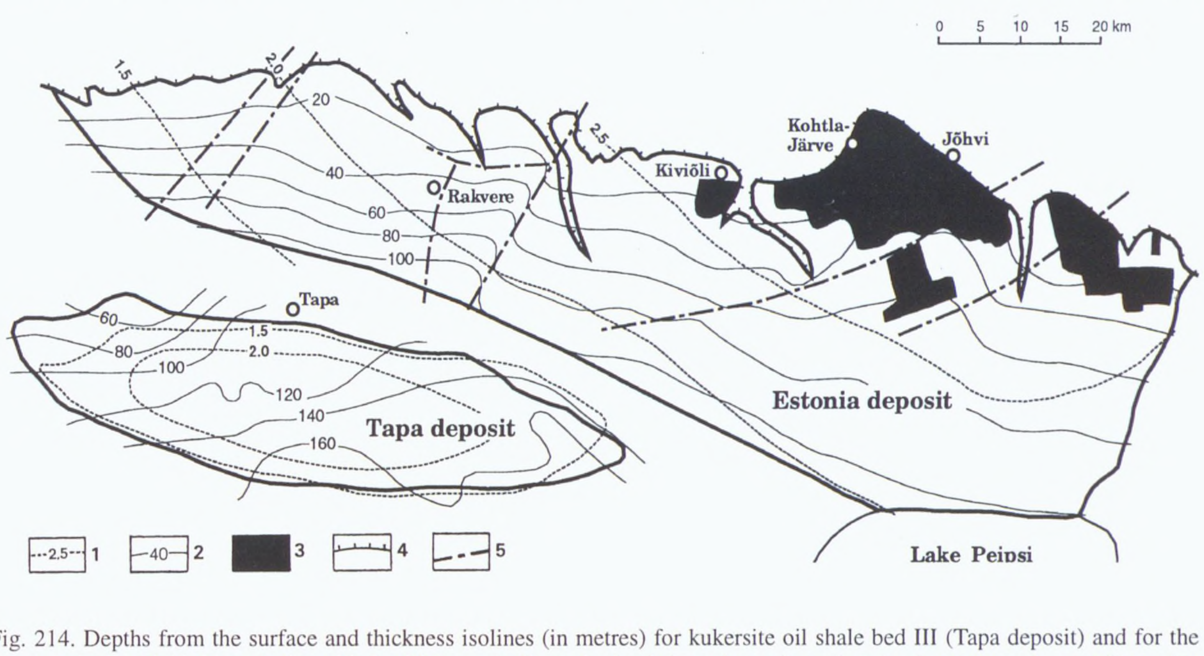


Fig. 214. Depths from the surface and thickness isolines (in metres) for kukersite oil shale bed III (Tapa deposit) and for the commercial bed A-F$_1$ (Estonia deposit): 1 - thickness isolines; 2 - depth from the surface; 3 - mined-out areas; 4 - post-Devonian erosional cut; 5 - linear zones of main tectonic disturbances.

1990). In the case of positive structures, kukersite oil shale seams may have been wedged out (the Sonda-Uljaste Dome), while increased water influx into shallow depressions usually complicates mining operations.

The kukersite oil shale mining is highly influenced by local **hydrogeological conditions** which mainly depend on the hydraulic conductivity of host rock, on the thickness of overburden and on seasonal variation in precipitation. The hydraulic conductivity is mainly determined by a degree of fracturing and by local karst phenomena. As a rule, the hydraulic conductivity of host rock is at its highest near the surface and lowers gradually downwards. This causes the highest influx of shallow groundwater into open-cast pits and into shallow mines. The atmospheric precipitation, which is considered as a main source for groundwater replenishment, swiftly passes through the thin cover of glaciofluvial deposits and infiltrates into the relatively highly fractured Ordovician carbonate rocks. Three water-saturated zones (see Ch. VI) have been established within the Ordovician carbonate sequence:

- in limestones of the Nabala - Rakvere stages (southern part of the deposit);
- in limestones of the Keila - Kukruse stages (throughout the area);
- in limestones of the Lasnamägi - Kunda stages (throughout the area).

The natural groundwater circulating in the Ordovician aquifer is weakly mineralized and characterized by a high content of hydrocarbonate and calcium ions. Because of the need to drain open-cast pits and underground mines, on an average 800 000 m^3 of water is pumped out from mines every day (Vallner 1994). The pumping data have shown that depending on local hydrogeological conditions and on seasonal variation, 5 to 56 m^3 of water (average 14-15 m^3) must be pumped out per every ton of mined kukersite oil shale. Almost all pumped-out water (96-97%) is directly discharged either into the Gulf of Finland, the Narva River or Lake Peipsi. The calculations by Erg and Punning (1994), however, have shown that because of a high permeability of surface rocks, more than 40% of pumped-out water may infiltrate back into underground mines and open-cast pits.

The oil shale mining brings about changes in the groundwater regime and its composition. As a result of extensive draining of open-cast pits and mining shafts, the groundwater table has noticeably lowered in the area of the Estonia deposit (Fig. 215). The most influenced is the Lasnamägi-Kunda water saturated zone where the depression extends 7-8 km outward from the areas of the currently active underground mines. In the areas of open-cast pit mining, a 10-20 m deep groundwater depression has formed (Erg & Punning 1994) and its influence is felt several kilometres apart from open-cast pits. In cases of deep subsurface mining, the depth of groundwater depression depends on depths of mining shafts which may reach down to 70 metres from the surface.

The infiltration of drainage water through overburden back to underground mines and open-cast pits brings about changes in the composition of water. The most noticeable change is a sharp increase of sulphate anions up to 600 mg/l, a natural background for SO_4^{2-} being between 2-10 mg/l (Erg & Punning 1994, table 4). Evidently, the rise of sulphate anions in such waters is caused by oxidation of pyrite in well-aerated water which percolates down through the disturbed overburden. In addition to the sharp increase in sulphate content, those waters have become enormously hard and are often polluted by oil from leaking mining machinery (Vallner 1994).

There are several **nature reserves** with a total area of about 1000 km^2 where oil shale mining is not permitted within the limits of the Estonia oil shale deposit (Fig. 215). In most of the western and northwestern part of the deposit, no mining is permitted because of the Pandivere Water Protection Area (Pandivere watershed area). Besides, the following nature reserves are located in this region:

- eastern part of the Lahemaa National Park;
- landscape reserves (Mõdriku-Roela, Võlumäe-Linnamäe, Uljaste);
- bog reserves (Sirtsi, Muraka).

There are a few nature reserves with a total area of about

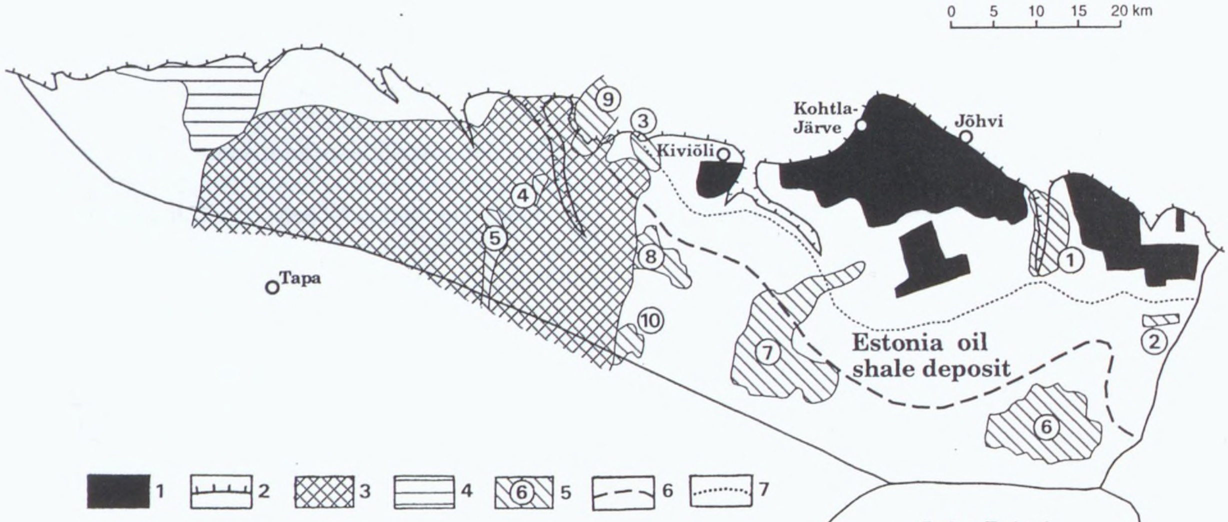


Fig. 215. Location of water and nature reserves in the area of the Estonia oil shale deposit. The dashed lines indicate the extent of water depression formed because of extensive pumping from open-cast pits and from underground mines. Landscape reserves: 1 - Kurtna; 2 - Porkuni; 3 - Uljaste; 4 - Võlumäe; 5 - Mõdriku-Roela. Bog reserves: 6 - Agusalu; 7 - Muraka; 8 - Sirtsi; 9 - Sämi-Kuristiku; 10 - Tudu-Järvesoo.

70 km^2 in the central and eastern parts of the Estonia deposit:
- landscape reserves (Porkuni, Kurtna);
- bog reserves (part of the Muraka and Agusalu reserves).

The rest of both bog reserves (about 230 km^2) lies within the southern area of the Estonia deposit.

Oil shale mining and utilization

The Estonia kukersite oil shale deposit is almost completely explored and has been mined since 1919. According to the present assessment, the total amount of resources is about 7 x 10^9 tonnes. Currently, kukersite oil shale is mined in three open-cast pits and six underground mines; in 1996 the annual output was about 13 x 10^6 tonnes. About half of the whole output comes from open-cast pits. Surface mining is economically feasible if the mineable bed is at a depth of less than 25 metres below the surface. The other half of the output comes from underground mines where shafts reach a depth of 60-65 metres from the surface. The calculations suggest that during the past 80 years of mining, about 850 x 10^6 tonnes of kukersite oil shale has been brought to the surface from a mining area of about 400 km^2.

It must be emphasized that all former and all currently operating mines are located in the northern part of the central and eastern areas of the Estonia deposit (Fig. 208) which is characterized by the best technological and mining parameters and by the lowest mining costs. The prospective, but less feasible mining areas are father in the south and west.

In general, because of a high ash content (40-50%), kukersite oil shale is classified as a low-grade fossil fuel. However, in Estonia this fossil fuel is still considered as being capable of competing with other high-grade fossil fuels due to:
- low mining costs;
- proximity of mines to power plants;
- great distance to other sources of high-grade fossil fuels (oil, natural gas, coal).

Currently, kukersite oil shale is mined in underground mines either by room-and-pillar method or by shearer mining. In three currently operating open-cast pits, either a selective strip mining is carried out in three steps (Sirgala and Narva open-cast pits) or the whole commercial bed A-F_1 is mined out at once (Aidu open-cast pit), followed by later beneficiation of oil shale.

The grinding and beneficiation (based on a density difference between oil shale and host limestone) process allows to obtain the following size-class material:
- 0-25 mm - for pulverized combustion in power plants;
- 25-125 mm - for oil retorting;
- 20-300 mm - tailings (limestone aggregate).

The limestone tailings can be used in road construction and in the building industry, but so far they have found only a little useage because of their low frost-resistance.

The conventional oil shale mining may have a serious impact on the surrounding environment. The most common features of possible environmental damages are:
- pollution of surface and groundwater by polluted mine drainage waters;
- lowering of groundwater level;
- formation of water-filled depressions on the ground (be cause of collapses of mining shaft roofs);
- changing of soil properties and the overall landscape (result of wide-scale strip-mining);
- formation of waste dumps near beneficiation facilities (where the residual OM is prone to self-ignition).

Another concern in kukersite oil shale mining relates to the high mining losses. To support the roof of mining shafts, about 25-30% of mineable oil shale is left intact as pillars. The supplementary losses form during hauling and beneficiation steps, increasing the net losses to as high as 50%.

The main kukersite oil shale consumers are two power plants - the Baltic Thermal Power Plant (project capacity: 1624 MW) and the Estonian Thermal Power Plant (project capacity: 1610 MW). Together with some smaller local power plants they use up more than 80% of mined oil shale. Both power plants are of environmental concern due to gaseous emissions of NO_x, SO_2, HCl, fly ash, cancerogeneous polycyclic aromatic hydrocarbons, *etc.* (Ots 1992).

About 15% of mined oil shale is used for crude oil retorting. Because of its specific fractional and chemical composition (mainly oxygen-containing compounds), it is not an appealing source for refining motor fuels, but has a high value as a raw oil for chemical industry (Joonas *et al.* 1994). The number of various products derived from crude oil totals about 50. Besides, for manufacturing fuel oils and gas for household use, several other products like wood-impregnation oils, electrode coke, roofing and construction mastics, soil conditioners, rubber softeners, casting binders, *etc.* are produced. The shale oil phenols are used as a feedstock material for various resins, synthetic tanning agents, rubber modifiers, *etc.* (Yefimov *et al.* 1994).

The retorting of oil shale and upgrading of shale oil to commercial products has a severe impact on the surrounding environment. Both the discharge of retorting wastewaters into local streams and leaching of organic compounds and toxic elements from a retorted spent shale heaps into the ground- and surface water is causing a large-scale environmental pollution around retorting sites.

Tapa oil shale deposit

An extensive survey for new oil shale prospects was carried out south of Rakvere and Tapa towns in 1967-68. As a result, a new - Tapa oil shale deposit was established (Fig. 203). This deposit is based on the III kukersite seam which lies 5-8 metres above the kukersite seam F_1 in the upper part of the Viivikonna Formation (Figs. 204, 205).

The economically most prospective part of this deposit covers an area of about 1150 km^2 (length *ca.* 80 km, width *ca.* 15 km). The thickness of the III kukersite bed ranges between 2.0-2.3 metres in the central part of the deposit and it thins out toward the limits of the deposit to 0.6-1.0 metres (Fig. 211). Within the Tapa deposit, the III kukersite bed lies at a depth of 50-160 m below the surface (Fig. 214).

Lithologically, this seam resembles to nodular limestones where kukersite OM forms a matrix between kerogenous limestone nodules. These limestone nodules make up 25-60% of the total volume. The main mineral constituents of the III kukersite seam vary within the range:
- OM content: 10-25%;
- carbonates: 60-70%;
- clastic minerals: 14-20%.

The analytical data indicate that while OM content diminishes toward the distal areas of the deposit, carbonate content increases westward and clay material content increases both in a southern and eastern direction. The technological quality of oil shale in the III kukersite bed is rather low: calorific value ranges 6-8 MJ/kg and oil yield 9-13% (Fig. 10). The preliminary estimated oil shale resources of the Tapa deposit reach 2.6 x 10^9 tonnes. Because of several inhibiting factors like:

- oil shale is rather low-grade;
- thick overburden (50-160 m);
- environmental restrictions,

oil shale mining activities in the Tapa oil shale deposit are not regarded feasible at present (Каттай и Рейнсалу 1991).

Dictyonema ARGILLITE

Dictyonema argillite (*Dictyonema* shale, alum shale, graptolitic argillite) of the Estonian Tremadock is a fine-grained sedimentary rock, rich in organic matter (OM) and pyrite (Luha 1946). It is characterised by a high concentration of C and S and the presence of several metals, first of all Mo, U and V. In Estonia, *Dictyonema* argillite occurs only in the northern part of the Republic (Fig. 216). It belongs to the extensive formation of the Cambrian - Ordovician black shales which form patches in the latitudinal zone extending from Lake Onega in the east to the Jutland Peninsula in the west. *Dictyonema* argillite correlates with the upper part of alum shale section of the Cambrian - Ordovician system in central and southern Sweden (Andersson *et al.* 1985).

Dictyonema argillite exposes in several locations between the Pakri Peninsula and Narva Town. The deposit has a southward inclination (3 m per km). The boundary with the underlying layer is sedimentary and lithologically sharp. The upper boundary is erosional. *Dictyonema* argilllite reaches a maximum thickness of approximately 6 m (8 m on Osmussaar Island) in northwestern Estonia (Бауков 1968). To the east and west the thickness of the deposit decreases (Fig. 216) and the number of light-grey quartzitic silt interlayers, from a fraction of millimetre up to several centimetres in thickness, increases.

Dictyonema argillite contains prevailingly 65-75% mineral particles, 15% roentgenoamorphous and 15-20% organic matter (Утсал и др. 1982) and 2.4-6.0% pyrite. In the crystalline mineral matter, pelitic fraction (< 0.01 mm) averages 79.67% (Table 51). Its share decreases in those parts of the section where silt interlayers are numerous. Sand fraction accounts for less than 2% in *Dictyonema* argillite. According to Kleesment (Клеесмент и Курвитс 1987), this fraction is mostly composed of fragments of fossils (graptolites, conodonts, brachiopods), the content of minerals is negligible. In the vertical section, the percentage of pelitic fraction decreases from top to bottom. The amount of sand and silt grains increases from the south-west to the north-east.

The data of X-ray analysis have shown (Утсал и др. 1982) that the crystalline mineral part of *Dictyonema* argillite consists prevailingly of clay minerals, feldspars and quartz (Table 51). The percentage of micas (mouscovite, green and brown biotite), chlorite and other minerals is low. In general, terrigeneous minerals prevail with their sum accounting for 59%; the sum of clay minerals makes up only 40%. Of clay minerals, montmorillonite-illite forms the major part of the finest fraction and has the highest concentration. The percentage of chlorite is insignificant. Kaolinite is absent or very low (<1%). The most recent, yet unpublished data of X-ray analysis show that K-feldspar is partially represented by sanidine.

On the boundary of <2μm and >2μm fractions, the grain-size and mineral composition of *Dictyonema* argillite changes distinctly. The percentage of 2-5μm fraction increases abruptly and clay minerals are substituted with K-feldspar and quartz. K-feldspar is often fresh, idiomorphic and poorly orientated (Утсал и др. 1982, Петерсельль и др. 1987).

The content of pyrite in *Dictyonema* argillite is highly variable, usually 1.5-9.0%, prevailingly 2.4-6%. Pyrite forms fine-crystalline disseminations, lenses, thin interlayers and concretions with different forms and sizes. The diameter of concretions does not usually exceed 2-3 cm. Part of concretions are complex in structure and contain crystals of galenite, sphalerite and calcite. Pyrite comprises also marcasite. Un-

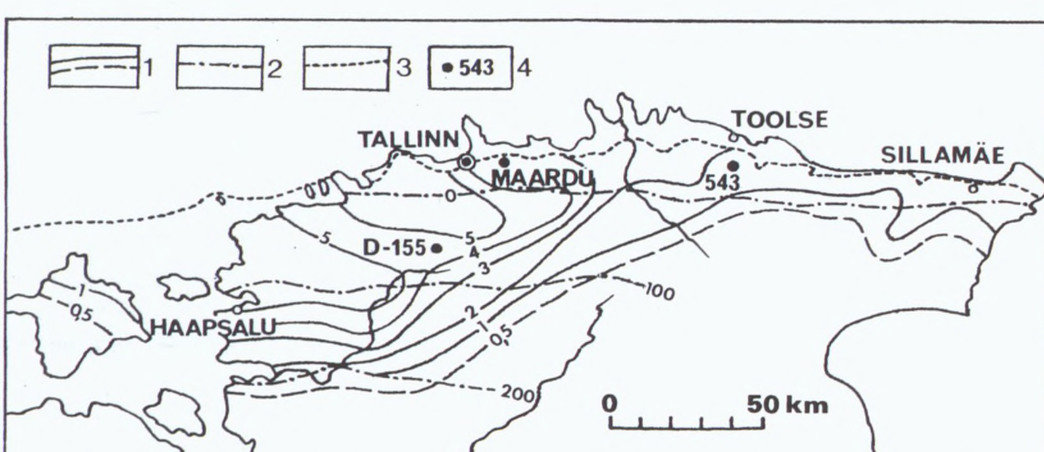


Fig. 216. Scheme of the occurrence of *Dictyonema* argillite in Estonia: 1 - isopachyte of the stratum, m; 2 - isobath of the stratum's upper surface, m; 3 - erosional northern boundary of the strata; 4 - boreholes in Fig. 217.

Table 51. Mineral and grain-size composition of the crystalline part of *Dictyonema* argillite

Fraction	Grain-size, µm	Content, %	Mineral composition, %*					
			Montmoril-lonite-illite	Swelling illite, illite	Micas	Chlo-rites	Quartz	Feldspars
Pelite	<0.2	15.54	100					
	0.2-0.35	0.83		93.57		1.43	2.14	2.86
	0.35-0.5	1.09		92.14			5.00	2.86
	0.5-0.75	1.59		93.57			2.14	4.29
	0.75-1.0	2.29		84.29			7.14	8.57
	1.0-2.0	6.0		76.43			8.00	14.14
	2.0-5.0	35.95		40.71		1.43	24.29	34.29
	5.0-10	16.38			19.29	0.71	35.43	43.71
Silt	10-100	20.33			10.29	0.57	37.86	50.14
	<0.2-100	100	15.54	24.56	5.42	0.51	22.96	30.54

* The sensitivity of mineral analysis was approximately 1%

der atmospheric conditions, this part of pyrite is rapidly substituted with jarosite and anhydrite. *Dictyonema* argillite contains also light-brown phosphatic ooids, a few millimetres in diameter, which have cracked as a result of dehydration. Besides sulphides, *Dictyonema* argillite comprises also auxiliary minerals, including zircon, tourmaline, garnet, rutile, glauconite and carbonates (mainly siderite). Medium- to high-resistant minerals, such as corundum, amphiboles and disthen have preserved in western Estonia (Клеесмент и Курвитс 1987). According to Baukov (Бауков 1968), apatite, titanite, alkali amphibole (aegirine), baryte and diopside have also been found in *Dictyonema* argillite.

OM is fine-dispersed and spread rather evenly. In the middle of the section, its percentage somewhat increases. The OM concentration is at its lowest in those parts of the section where the interlayers of carbonate minerals or silt are high in number (Fig. 217). OM binds the shale and makes it swelling resistant.

Authigenic, phosphatic, siliceous and sulphatic formations and carbonate minerals are associated with silt interlayers. Sulphides are dominated by pyrite, the content of which reaches 30-40%. Some interlayers abound in sphalerite. The occurrence of galenite and calcopyrite is significant only from

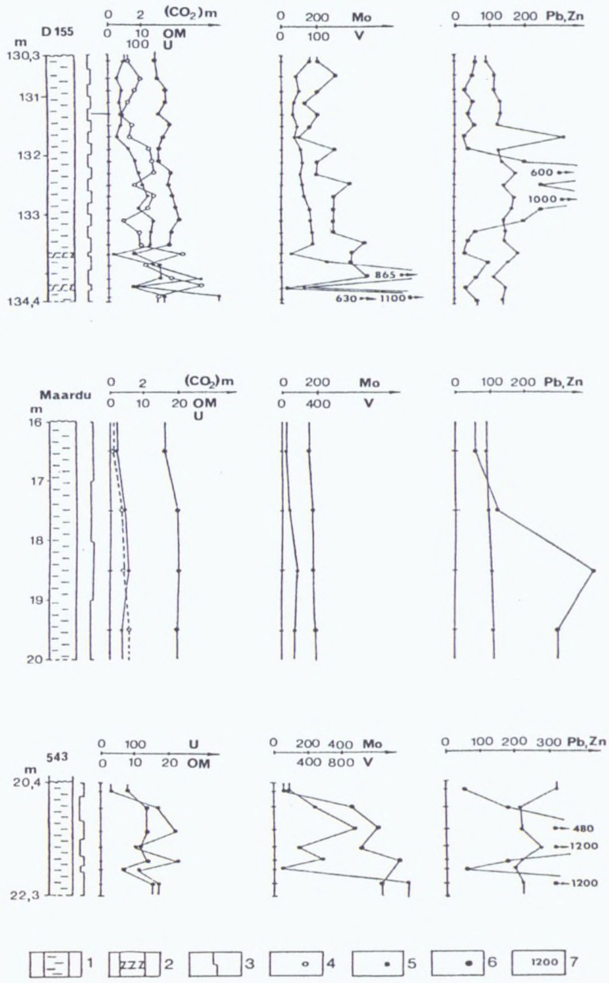


Fig. 217. Distribution of OM and $(CO_2)_m$ and microelements in *Dictyonema* argillite: 1 - *Dictyonema* argillite; 2 - carbonate-rich interlayer; 3 - sampling interval; 4 - CO_2 (%); 5 - OM (%), V and Pb (ppm); 6 - U, Mo and Zn (ppm); 7 - the highest contents (ppm). For location of boreholes see Fig. 216.

mineralogical point of view. Argillite contains also galenite and sphalerite with numerous small calcite veins (Петерселль и др. 1987).

Carbonate minerals occur sporadically. They cement terrigenous material in patches or form concretions: calcite in the western and dolomite in the eastern sections. The occurrence of phosphatic cement or even lumps of the earlier formed chemogeneous phosphatic layer are characteristic of some silt interlayers (Loog *et al.* 1995).

The section of *Dictyonema* argillite, particularly in the eastern part of the distribution area, comprises small lense- or nest-shaped silica interlayers, complex in structure and composition. The main bulk is made up of white, in places grey (due to the admixture of clay minerals, pyrite, OM, *etc.*), porous, soft silica and terrigeneous quartz with the grain-size of fine sand (Loog & Petersell 1995).

The chemical composition of *Dictyonema* argillite is of great interest. Its K and S content is much higher and the content of Na and Ca lower than the clark for clays (Table 52). The concentration of K in *Dictyonema* argillite is higher than could be expected on the basis of the composition of the known shale-forming minerals. This suggests the presence of some so far undetermined minerals or compounds with a high K concentration. With the increase of the volume of silt interlayers from the west to the east, the concentration of Si, Ca and P increases, while that of Al, K and Mg decreases

The concentration of S is variable. Prevailingly it is in a range of 2 - 6%, of that 0.6-0.8% is comprised in OM, ca. 0.3% is sulphatic, and the remaining part sulphitic S. Some 58% of the samples analysed had S with the isotopic composition ($\delta^{34}S$ +2.9....-3.8‰) close to meteorite standard (Петерселль и др. 1987).

OM in *Dictyonema* argillite is sapropelic in origin (Pukkonen & Rammo 1992) and rich in N, S and O (Table 53). The ratio of C and H in OM is ca. 9, but the calorific value is still low, being 5.5-6.0 MJ/kg as an average.

More than 12 microelements have been determined in *Dictyonema* argillite, the concentration of which is 2-100 ppm, occasionally even more times in excess of the clark for clays (Лоог 1962, Петерселль и др. 1981, Пукконен 1989). The value varies laterally and with the regions and parts of the section (Table 54, Fig. 217). The concentration of commercially important elements Mo, U and V is high in north-eastern Estonia where the thickness of *Dictyonema* argillite does not exceed 1-2 metres. In the western and northwestern parts of the Republic, where the deposit is thick, only its lowermost two metres are enriched with those elements. Occasionally, at some sites the concentration of microelements in the enriched layer may be as high or even higher than in the Toolse area.

Zn is of peculiar distribution. Predominantly its content is lower than the average in clays. However, in several parts of the section, the content of Zn is higher than 1000 ppm (Fig. 217). Exceptionally, in western Estonia the proportion of Zn reaches 0.5 - 0.6%. In the area between Keila and Viitna, *Dictyonema* argillite is low in microelements.

Table 52. Chemical composition of *Dictyonema* argillite, %

Compound	West Estonia		Maardu (Киррет и др. 1959)		Toolse (Раудсеп 1987)		Clark for clays
	Argillite	Anorganic part	Argillite	Anorganic part	Argillite	Anorganic part	
SiO_2	48.92	62.32	52.09	63.52	51.15	64.11	50.93
Al_2O_3	13.09	16.68	13.09	15.96	9.76	12.28	19.75
Fe_2O_3	5.61	7.15	5.68	6.93	8.03	10.06	5.23
CaO	0.49	0.63	0.82	1.00	2.82	3.52	3.54
MgO	1.49	1.90	1.42	1.74	1.08	1.34	2.24
Na_2O	0.08	0.10	0.56	0.68	0.09	0.11	0.89
K_2O	7.89	10.05	7.47	9.11	5.73	7.18	2.75
TiO_2	0.73	0.93	0.64	0.78	0.73	0.91	0.77
P_2O_5	0.20	0.24	0.23	0.28	0.39	0.49	0.16
L.O.I.	21.39		17.83		20.60		
Σ	99.89	100.0	99.83	100.0	100.38	100.0	
S_{tot}	2.62		3.01		5.07		0.24
FeS_2	2.99		8.81				

Table 53. Composition of the organic part of *Dictyonema* argillite

Locality	Ash (A),%	Organic matter (OM),%	Concentration of elements in organic matter, %				
			C	H	N	S	O
Maardu	79.2	16.8	70.5	8.3	2.5	2.4	16.3
Toolse	80.5	15.4	69.0	7.6	2.1	2.1	20.2

In recent years, anomalously high Au and Pt concentrations, more than 1 ppm, have been registered in some places. Considering the small amount of explorations carried out so far and the frequency of anomalous concentrations, these elements seem to have received less attention than they actually deserve. Occasionally, the concentration of Re is very high, amounting to 3 ppm in the Toolse area.

Mineral occurrences of Mo, Re, U and V have not been identified so far and these elements are supposed to occur prevailingly as metal-organic compounds. The contrastic anomalies of polymetals are due to their sulphitic form; usually they occur together with pyrite or are scattered in the latter (Loog & Petersell 1994). More than 50 % of anomalous Au concentrations are caused by extremely fine-grained pure gold which occurs either in sulphites or is scattered in *Dictyonema* argillite. It may be assumed that Au is partially represented by compounds of tellurite and selinite, since the concentration of the latter is high in *Dictyonema* argillite (Table 54).

The genesis of *Dictyonema* argillite is not unambiguously clear. In all likelihood, it took place under unaerobic conditions in a body of stagnant water. The input of abundant clay minerals, fine terrigeneous material from the erosion area and volcanic ash as a result of volcanic activity promoted the development of extensive clay facies. The low wearing degree of K-feldspar and quartz grains and montmorillonite platelets, the frequent absence of their orientation within the layer as well as the occurrence of sanidine-like feldspar are indicative of the presence of volcanic matter (Утсал и др. 1982, Петерсельь и др. 1987). Caledonic orogenese and the accompanying magmatic activity promoted the development of deep faults in the Tremadoc (Petersell & Levchenkov 1994), which opened the way for the hydroterms into the Baltic Palaeobasin (Лоог и Петерсельь 1986). From the geological point of view, short-term (some millennia) specific palaeoecological conditions formed which favoured the development of only single groups of organisms (bacteria, algae, *etc*.). In the course of time, part of these organisms was enclosed in sediments and during the succeeding diagenesis formed an organic-bearing layer of *Dictyonema* argillite. The persistent composition of rock-forming minerals and even the distribution of several elements (K, Rb, Sr, Th) in *Dictyonema* argillite suggest calm and balanced conditons of sedimentation in the basin. The distribution of Mo, Pb, Re, U, V and Zn, local and uneven in character, cannot be unambiguously associated with the accumulation of organisms. First of all, the character of polymetallic mineralization and the closeness of $\delta^{34}S$ to sulphur of deep origin refer to partial input of metal-rich endogenic matter during the accumulation of sediments.

Table 54. Approximate average concentration of microelements in *Dictyonema* argillite, ppm

Element	West Estonia	Maardu	Toolse	Clark for clay
Au	0,008	0.02	0.004	0.001
As	<60	<60	38	13
Cu	94	80	75	45
Mo	162	53	406	2.6
Ni	185	75	140	68
Pb	130	98	120	20
Re	0.10	0.08	0.18	<0.01
Se		1.6	4.3	0.6
Te		0.05	0.1	0.01
Th		7.4	14.5	1
U	86	36	162	3.7
V	724	350	1040	130
Zn	222	220	170	95

Dictyonema argillite is a potential complex mineral resource. Its reserves are estimated at 60 billion tons. In terms of the distribution, energy and metals accumulated in the rock, there are only a few rock formations in the Earth's crust comparable with *Dictyonema* argillite and its analogs - Swedish alum shales. *Dictyonema* argillite is a low-quality oil-shale, ore of U, Mo, V and other valuable metals. Its calorific value ranges from 4.2 - 6.7 MJ/kg (Pukkonen & Rammo 1992), volume weight is changeable and high, up to 1.8 - 2.5 g/cm^3. Negative correlation between the volume weight and calorific value, so typical for oil shales, is absent or weekly expressed in *Dictyonema* argillite.

The energetical potential of U, enclosed in *Dictyonema* argillite, is close to the amount of the energy derived through burning of one tonne of high-quality coal. The concentration of Mo, which exceeds 300-400 ppm together with U, V and other elements, would be of commercial interest if there were economically acceptable and environmentally sound technologies.

From 1948 to the 1960s, *Dictyonema* argillite was excavated at Sillamäe as raw material for producing uranium. In the mined out *Dictyonema* argillite, the concentration of U was about 300 ppm and it was extracted in a sulphuric acid medium.

The possibilities of producing microelements from

Dictyonema argillite have been studied during many years by the researchers of the Estonian Academy of Sciences (Маремяэ 1989, Маремяэ и др. 1991) and different All-Union institutions. *Dictyonema* argillite and its ash, rich in K, P, Mo, U, Zn, Cu and several other elements, could find use in agriculture for raising the fertility of soils, poor in these elements.

PHOSPHORITE

Research into the *Obolus* sandstone (Upper Cambrian to Ordovician in age) in Estonia goes back well over a century. Nevertheless, its utilization lagged behind the studies. In 1861, C. Schmidt stressed the significance of *Obolus* sandstone as a possible raw material in the manufacture of fertilizers which is easy to enrich through sifting. However, it was not until the end of World War I when an acute shortage of resources made the German geologists to get down to studying *Obolus* sandstone as a potential phosphorite ore.

The high prices of phosphate fertilizers immediately after World War I, compelled the Ministry of Agriculture of Estonia to initiate relevant investigations. In 1920, the joint-stock company "Eesti Vosvoriit" ("Estonian Phosphorite") was established and the first geological investigation was started near Tallinn. It went on for 2 years. As a result, *Obolus* sandstone was rated as phosphorite ore. In 1924, a mine and plant started to operate at Ülgase. The plant was in operation until the fire in 1938. A new mine and an enrichment plant were built at Maardu in 1940.

During World War II, the annual output of phosphorite was low. However, in the post-war period the Soviet authorities ordered widening of the phosphorite mines at Maardu. The highest annual output reached 850,000 tonnes of raw material, most of which was used for producing an uneffective fertilizer - phosphorite meal. A small amount of concentrated phosphorite was added to superphosphate for neutralization of acid soils. Apatite, the raw material for superphosphate production, was brought from the Kola Peninsula. At the same time, the whole of northern Estonia was subject to large-scale explorations during which the already known deposits at Maardu, Tsitre and Aseri were studied and several new ones – Narva, Toolse and Rakvere – were discovered (Raudsep 1982).

In the 1980s, environmental problems gained increasing attention. In the Maardu mining area air, soil and water were contaminated. *Dictyonema* argillite (alum shale), which on phosphorite mining was removed with overburden and deposited in waste dumps, tended to self-ignition due to pyrite oxidation. The burning shale emitted hazardous gases and radioactive substances which were carried with the winds and water over a distance of many kilometres.

In 1991, considering the above circumstances and exhaustion of phosphorite resources at Maardu, the mining and enrichment of phosphorite were terminated. During the years in operation, a total of 25 million tons of phosphorite ore were extracted at the Maardu deposit.

At the end of the 1980s, the Estonian Government, on the motion of scientists and under the pressure of public opinion, succeeded in rejecting the demands of Soviet officials to start the building of gigantic mines in the Rakvere Phosphorite Region, including the Toolse and Kabala deposits. In doing so, the Estonian Government and publicity relied above all on environmental and socio-economic considerations (Raudsep *et al.* 1991, Raudsep 1994).

Geology

Estonian phosphorite is a yellowish-light or dark-grey fine- or coarse-grained slightly cemented sandy deposit. This typical shelly phosphorite (Пуура 1987, Raudsep *et al.* 1991) occurs at the Upper Cambrian/Lower Ordovician boundary (Kallavere Formation). The basic rock-forming minerals are quartz and biogenic phosphate (flourcarbonate apatite), represented by remnants of brachiopods. The proportion of these minerals varies with layers and deposits. The content of remnants of brachiopods in the rock ranges from 5-10 to 80-90%. The brachiopod shells and detritus contain up to 35-37% P_2O_5 (Table 55), in the whole phosphorite layer the content of P_2O_5 is in a range of 6-20%.

Phosphate rocks also contain dolomite, calcite, pyrite, glauconite and ferrous hydroxides which occur in small amounts, but are sometimes very important for the purposes of ore enrichment.

The sandy deposits of the Kallavere Formation are spread nearly all over Estonia, except the narrow southwestern and

Table 55. Average composition of inarticulate brachiopod valves from the Kallavere Formation

Components	Deposits / content, %				
	Rakvere		Toolse	Aseri	Maardu
	dark valves	light valves			
P_2O_5	35.37	37.11	33.85	34.59	34.66
Insoluble residue	1.04	1.20	3.91	1.18	0.71
MgO	0.63	0.67	0.45	0.47	0.21
Fe_2O_3	3.15	1.17	2.45	3.47	1.59
FeS_2	3.36	0.21	x	4.46	3.20
SO_3	0.47	0.32	x	0.27	1.26
SiO_2	0.79	1.17	3.16	0.64	x
Al_2O_3	0.51	0.46	0.31	0.73	x
CaO	50.21	52.34	47.56	49.78	50.58
CO_2	3.46	3.68	3.41	3.57	x
S_{total}	1.20	0.24	1.53	2.66	x
F	2.03	2.01	2.07	2.21	2.67
Na_2O	0.54	0.47	0.89	0.61	x
K_2O	<0.10	<0.10	traces	<0.10	x
Number of analyses	6	12	12	15	1

northeastern zone which extends from Saaremaa Island to Lake Peipsi (Fig. 218). The thickness of the formation ranges from 1 to 20 m. The commercial phosphorite bed at a depth of 5-200 m is 1-12 m thick. The monoclinally bedded sedimentary complex has a slight southward inclination (10-15'). Some tectonic dislocations occur against the generally peaceful background.

The phosphorite layer is covered by *Dictyonema* argillite, rich in kerogen and pyrite, and by the Lower Ordovician clay, glauconitic sandstone, limestone and dolomite. The Quaternary cover is usually 0.5-3 m, in the buried valleys 20-90 m thick.

The major **phosphorite deposits** occur in northern Estonia. In order of size they are Rakvere, Toolse, Aseri, Tsitre and Maardu (Fig. 218).

Rakvere deposit

This is the biggest phosphorite deposit in Europe. It contains roughly 700 million tonnes of P_2O_5. The thickness of the layer ranges from 2 to 12 m and it lies at a depth of 42-210 m over an area of 1000 km^2. The content of P_2O_5 in the phosphorite is 10-20%, mainly 7-15% (Table 56).

Most of the deposit is situated in the Lääne-Viru County, part of it also in the Ida-Viru and Järva counties. The deposit is divided into several regions: Kabala (subdivided into West-Kabala and East-Kabala), Rägavere, Assamalla, Sonda, Southeastern and Western regions (Fig. 219).

The exploration of the Rakvere deposit was initiated in the mid-1970s and it lasted until 1988 when a detailed study of the Kabala mining field (West-Kabala) was completed.

In terms of the geological structure, the Rakvere deposit is similar to the other Estonian deposit, but its overburden is much thicker and crossed by dislocations. As a matter of fact, this is a complex deposit, the overburden of which contains remarkable resources of oil shale, sand, clay and peat.

According to geological peculiarities, the deposit may be divided into three parts (Fig. 219):

1) the northern and northeastern parts (Kabala, Rägavere and Sonda regions) where the commercial oil shale (kukersite) beds of the Estonian Oil-Shale Deposit occur 30-35 m above the phosphorite layer;

2) the central part (the northern part of the Western, Assamalla and Southeastern regions) where the oil shale reserves are unfit for use;

3) the southern and southeastern parts (the southern part of the Assamalla and Southeastern and Western regions) where the oil shale bed of the preliminarily estimated Tapa deposit occurs 40-50 m above the phosphorite layer.

As mentioned above, the thickness of the overburden is substantial, ranging between 50 and 200 m. In the northern part of the deposit, there are several zones of tectonic disturbances, karst phenomena and buried valleys (Fig. 219). In the Kabala mining field (cat. T reserves) the karst spreading through the limestones of overburden has partially affected the dolomitization of phosphorite.

In this deposit, both the thickness of the phosphorite layer and the reserves (Table 56) are greatest in Estonia: maximum 12 m, average ca. 8 m in the Kabala Region. The P_2O_5 con-

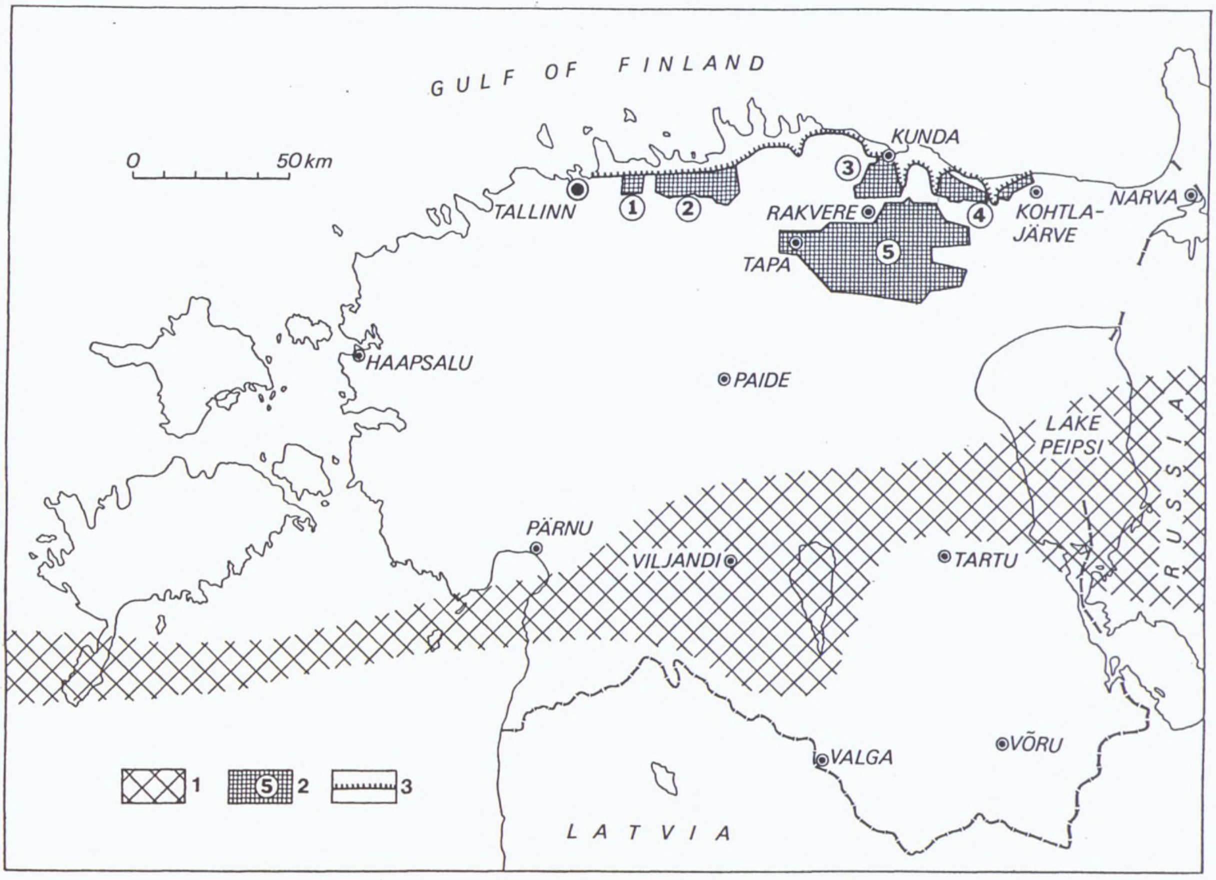


Fig. 218. Phosphorite deposits of Estonia: 1 - area where *Obolus* sandstone is lacking; 2 - phosphorite deposits: 1 - Maardu, 2 - Tsitre, 3 - Toolse, 4 - Aseri, 5 - Rakvere; 3 - klint.

Table 56. Average characteristics of Estonian phosphorite deposits

Deposit, area	Thickness of phosphorite layer	P_2O_5 content	Productivity	Depth of phosphorite layer	Reserves of P_2O_5 (as of 1989)
	(m)	(%)	t P_2O_5/m^2	(m)	(million tonnes)
Maardu	0.5	13.0	0.24	5-20	Cat. T - 3.4?
Tsitre	1.5	8.5	0.25	7-32	Cat T - 4.1
Toolse	2.9	10.6	0.61	5-55	Cat. T - 27.4
Aseri	1.1	9.6	0.21	1-34	Cat. T - 22.5
Rakvere	3.1-7.6	7.1-14.9	0.40- 1.28	42-210	Cat. T - 25.0 Cat. R - 251.4Cat. P - 458.7
Among this, Kabala area	6.0	12.3	1.62	42-107	Cat. T - 25.0 Cat. R - 163.0
Total					Cat. T - 82.3 Cat. R - 251.4 Cat. P - 458.7

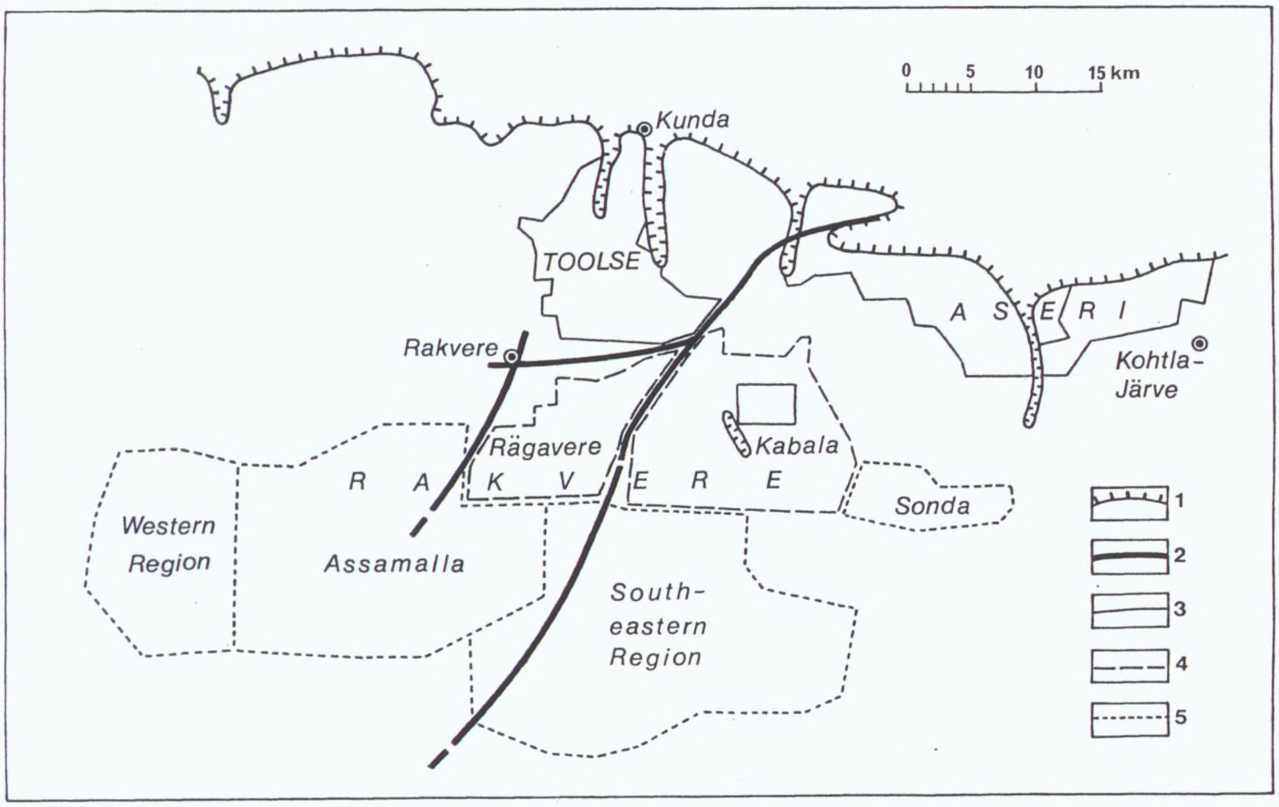


Fig. 219. Rakvere Phosphorite Region: 1 - northern limit of the present distribution of phosphorite; 2 - zones of disturbances; 3 - cat. T; 4 - cat. R; 5 - cat. P.

tent in separate strata is 3-28 % and, like the thickness of the layer, it decreases southwards.

The phosphorite of the Rakvere deposit is comparatively rich in P_2O_5 (Table 56). Its quality is partly aggravated (especially in the southern part of the deposit) by a high content of magnesium and ferro compounds.

In underground mining, serious technical problems may arise due to the brittle glauconite sandstone (1-2 m) resting straight on the phosphorite and the oil-shale bed lying still higher in the section. *Dictyonema* argillite forms a thin layer only in the northern part of the deposit. For instance, in the Kabala mining field the 2.5-m-thick commercial bed of oil shale (reserves 116 million tonnes) lies 20-35 m (Пуура 1987) above the phosphorite.

Hydrogeological conditions are complicated because the greater part of the deposit is located on the Pandivere Upland or at its foot where most of the groundwater rises in northern Estonia and any change would affect water regime over a large area. The main aquifers are as follows:

- Silurian-Ordovician groundwater complex containing an oil-shale formation;
- Ordovician-Cambrian groundwater layer containing a phosphorite layer;
- Cambrian-Vendian groundwater complex.

Hydrogeological conditions have better been studied in the Kabala mining field. It is predicted that on the event of complex extraction of phosphorite and oil-shale, the two upper aquifers in the Kabala mining field will have to be drained. As a result, large cones of depression will form.

Toolse deposit

The deposit was discovered during the investigations in 1957-60 (Raudsep 1982, Пуура 1987), though A. Öpik had described *Obolus* sandstone in the Toolse riverside as early as the 1920s. The geological mapping in the 1950s imparted new information on the spread of phosphorite layer in that region. The Toolse deposit was explored during 1966-71, the results were summarised in a geological report in 1971.

The deposit is situated in the northern part of the Lääne-Viru County — in an advanced agricultural and industrial (Kunda Cement Works, several enterprises at Rakvere and Tapa) area with a dense network of roads. Geologically, it is located on the outskirts of the large Rakvere Phosphorite Region (Fig. 219) on the southern slope of the Fennoscandian Shield. The thickness of the sedimentary complex with a S-SE inclination (10-15') increases southwards. In the north, the deposit borders on the klint and in the east the layer of phosphorite is intersected by the Kunda buried valley (Fig. 219). The southern and western boundaries of the deposit are transitional: in these directions the layer of phosphorite becomes less lavish.

The Toolse buried valley cuts into the bedrock (inc. the phosphorite layer) in the middle of the deposit (maximum depth 25-30 m). The studied area covers about 90 km^2 and forms a complex deposit which also contains top-quality limestones, glauconite sandstone, clay, *Dictyonema* argillite and peat.

The geological section of the deposit is as follows (from bottom upwards, Fig. 220):

1) The phosphatic formation rests on Lower Cambrian sandstone and siltstone (Tiskre Formation, 10-15 m thick).

2) The phosphatic sandstones of the Upper Cambrian - Lower-Ordovician Kallavere Formation (the so-called *Obolus* sandstone): detritic and highly detritic quartziferous sandstone, detritites, coquina (the so-called *Obolus* conglomerate). The thickness of the formation is about 1.5 - 8.0 m (average 4.5 m). The yielding layer in the lower part of the formation is 1.0 - 5.1 m (average 2.9 m) thick and contains, on an average, 10.6% P_2O_5, 0.4% MgO and 1.5% Fe_2O_3. In terms of the quality parameters, the resources of the *Obolus* phosphorite in the Toolse deposit may be rated as average (Table 56).

3) The rest of the Lower Ordovician is represented by the kerogenous *Dictyonema* argillite of the Türisalu Formation, that lies immediately on the *Obolus* sandstone and has an average thickness of 1.4 m. The *Dictyonema* argillite is overlain by:

- silty clay of the Varangu Formation (average thickness - 2.4 m);
- glauconite sandstone of the Leetse Formation (average thickness - 1.1 m);
- glauconitic clayey limestone of the Toila Formation and dolomitized limestone containing glauconite and oolitic limestones of the Sillaoru and Loobu formations (average total thickness - 9.5 m).

4) Middle Ordovician fine crystalline limestones, dolomitized limestones and marls forming a 0-40 (average 20) m thick complex. The thickness increases towards the south. The rocks of the Aseri Stage and Väo Formation, which form the lower part of the complex, are of the highest quality and provide an excellent construction material.

5) Quaternary deposits are represented by till, sand, clay and peat. Commonly, the formation is up to 2-3 m, in buried valleys up to 35-40 m thick.

Hence, the total thickness of the overburden of the phosphorite formation is 5-55 (average 24) m. Hydrological conditions complicate both the opencast and underground mining.

On the basis of the composition, four types of phosphorite have been distinguished: silicon, weathered, ferrogenous and silicon-ferrogenous. Of the above varieties, the silicon ("normal") phosphorite is most common accounting for about 70% of the deposit's total area.

Technological experiments have shown that the "normal" phosphorite can be easily enriched by flotation (Table 57): the received concentrate contained 28% (occasionally even more than 30%) of P_2O_5, whereas the content of harmful components (first of all MgO and ferro compounds) was low. The worst results were obtained in enriching the "weathered" phosphorite. It was impossible to get a normal concentrate which could be used in producing double superphosphate or ammonium phosphate.

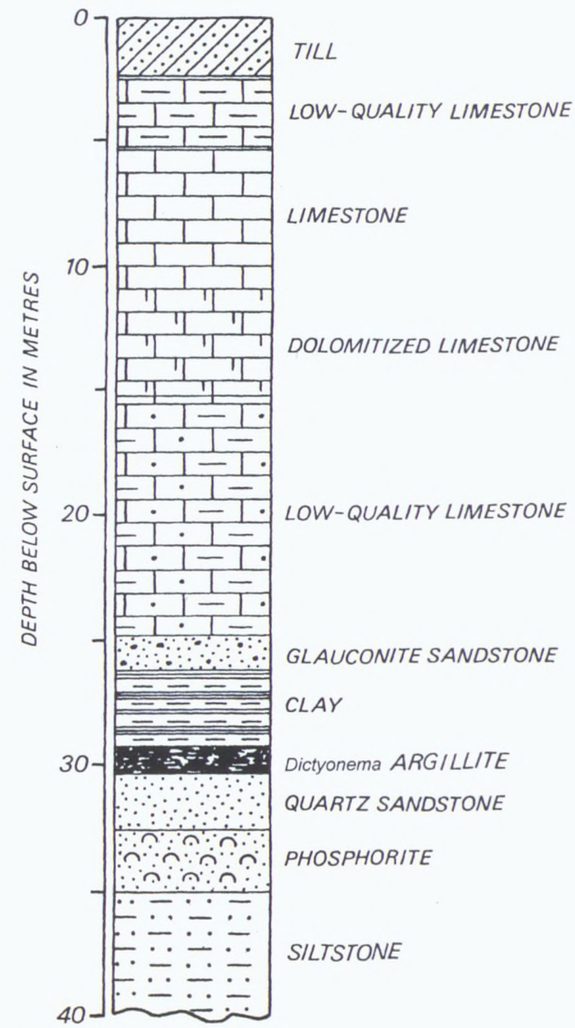


Fig. 220.Generalized section of the Toolse phosphorite deposit (Raudsep 1994).

Compared to other Estonian phosphorite deposits, the productivity of the Toolse phosphorite may be rated as average (Table 56)

During the exploration work, the rocks of the overburden were described and sampled. The results showed that the limestone of the Lasnamäe and Uhaku (the lower part) stages can be used in the manufacture of cement and for building purposes, glauconite sandstone for producing colour-pigments and silicate concrete, silty clay for ceramcite (brand 500-800). Rare elements (U, Mo, V, *etc.*), alum, sulphur, ammonium sulphate and several other products can be derived from *Dictyonema* argillite.

The **reserves** of phosphorite are 27.4 million tonnes (Table 56). The estimated reserves of minerals in the overburden are as follows:

Limestone:

	for cement (thousand m^3)	building stone (thousand m^3)
- cat. T reserves	13989	11274
- cat. R reserves	87946	37846

Glauconite sandstone:
- cat. R reserves 60.6 million tonnes

Silty clay:
- cat. R reserves 242.2 million tonnes

***Dictyonema* argillite:**
- cat. R reserves 141.4 million tonnes (content of organic matter: 10.9-14.8%, average 12.7%).

Among this:
molybdenum - 57.4 thousand tonnes (content 406 ppm);
vanadium -147.1 thousand tonnes (content 1040 ppm).

Initiation of phosphorite mining at other deposits (Fig. 218, Tables 55-57) in the nearest future is excluded. The **Aseri** deposit is situated in a densely populated area and the phosphorite is low in quality. The reserves of the **Tsitre** deposit are small and located partly within the Lahemaa National Park. The small residual reserves of the **Maardu** deposit are partly located within a nature reserve and, besides, the ecological damage caused by phosphorite mining is still posing problems in this area.

Enrichment and utilization

Experimental enrichment involving the production of Estonian phosphorite combines both anionic and cationic flotation (Пуура 1987). The received concentrates of the Toolse and Rakvere deposits contain more than 28% P_2O_5, occasionally even over 30% (Tables 57, 58). The composition of the concentrates varies depending on the chemical and mineral composition of the ore. The concentrates differ in the content of quartz, but also in the compounds of magnesium and iron, having a detrimental effect on the technological processes in obtaining fertilizers and on the quality of products. Magnesium occurs mainly in dolomite. Further reduction of the dolomite content is difficult. Economically, the attempts to remove carbonates by means of flotation have not proved successful anywhere in the world. Carbonates can be removed from sedimentary ores by calcination at temperatures higher than 830^0 C ("Preliminary Technical and Economic Assessment of Developing the Toolse Phosphate Deposit, Estonia", report prepared by IFDC, 1993).

Some iron compounds (particularly Fe_2O_3) contain pyrite which is insoluble under sulfuric and phosphoric acid attack, but dissolves in oxidizing medium (nitric acid). Non-

Table 57. Results of enrichment (based on geological reports)

No.	Deposit	Method of flotation	Type of ore	Ore, %			Concentrate, %			
				P_2O_5	MgO	Fe_2O_3	Yield of P_2O_5	P_2O_5	MgO	Fe_2O_3
1.	Toolse	anionic	silicon	12.30		1.19	83.6	28.20		2.57
2.	Toolse	anionic	weathered	11.58		0.94	53.0	25.76		
3.	Toolse	cationic	ferrogenous	11.65		2.38	83.2	28.00		3.18
4.	Toolse	combined	silicon	9.06			74.6	33.50		
5.	Toolse	combined	ferrogenous	10.38			71.0	33.50		
6.	Aseri	anionic	ferrogenous	10.03		2.13	60.4	25.22	3.51	3.98
7.	Aseri	cationic	ferrogenous	8.70		2.11	54.2	28.75	3.26	3.62
8.	Aseri	combined	ferrogenous	10.03	1.28	2.13	56.3	28.20	2.82	3.27
9.	Rakvere	anionic	rich	11.70	1.05	1.80	70.0	28.60	2.50	2.90
10.	Rakvere	cationic	rich	11.70	0.99	1.74	75.2	28.40	1.62	3.00
11.	Rakvere	cationic	rich magnesion	12.38	1.43	1.45	70.5	28.20	2.42	2.52
12.	Rakvere	combined	rich magnesion	12.14	1.61	1.55	56.5	33.40	0.40	2.98

pyritic iron dissolves completely when processed with sulphuric and phosphoric acids.

Quartz is the most abundant impurity in the concentrates (up to 25-26%), but its further reduction should be possible. The chlorine content is very high (more than 300 ppm); the cadmium content is low (5-6 pmm).

The processes of obtaining mineral fertilizers from Estonian phosphorite concentrates have been studied in laboratories. It has been established that the obtaining of good-quality fertilizers (triple superphosphate, ammonium phosphate, *etc.*) is possible.

Nitric acid decomposition of Estonian phosphorite concentrates is accompanied by the emission of gases (CO_2, NO_x) and the foaming of pulp. Elimination of foam and gas emission can be attained by thermal treating of concentrates (Veiderma 1993).

Problems of development

The opening of new phosphorite mines and plants in Estonia (Toolse and Rakvere deposits) has recently been widely discussed. The problems to be addressed may be divided into four groups: technological and technical, environmental, economic and sociological.

Technological and technical problems of development differ with deposits. The phosphorite of the Toolse and Rakvere deposits gets easily enriched by flotation. Some types of ore (weathered, high silicate, carbonate) are difficult to enrich and it is impossible to get normal concentrates out of them. In general, all the concentrates have high total impurity (MgO, Fe_2O_3, Cl) contents. There is a possibility that the areas contributing to the high impurity content will not be mined. Additional studies are needed.

Table 58. Average composition of Estonian phosphorite ore and concentrate

Component	Maardu deposit		Toolse deposit	
	Ore	Concentrate	Ore	Concentrate
		%		
P_2O_5	10.5	28.8	10.6	28.2
CaO	16.0	41.2	16.0	41.3
MgO	0.8	1.1	0.8	1.1
Fe_2O_3, total	1.8	2.9	1.9	3.4
Fe_2O_3,pyrite	1.2	2.3	1.3	2.4
Al_2O_3	0.7	0.4	0.5	0.3
S, total	0.8	1.6	1.1	1.8
CO_2	1.5	3.4	1.9	3.9
F	0.9	2.7	0.8	2.7
SiO_2	65.0	15.5	64.0	17.3

Trace Elements
U - 38-50 ppm
Pb - 50-110 ppm
Cd - 5.5- 6.3 ppm

Another group of problems relates with the mining of phosphorite. In the Rakvere deposit, where the phosphorite seam lies at a great depth, only underground mining is possible. The main difficulty is the lower non-solid rock overburden. The oil-shale layer some 20-35 m above the phosphorite can be removed and utilized.

Open-cast mining is possible only in the Toolse deposit, however, there are several problems to be addressed before it can be initiated, among those the efficient use or, at least, safe disposal of the *Dictyonema* argillite and other minerals found in the overburden.

The above is a topical environmental problem. A terrifying example is the mined-out area at Maardu where the loose *Dictyonema* argillite in the waste dumps is liable to spontaneous ignition. Due to its sulphur content (2-3%), the burning *Dictyonema* argillite contaminates the atmosphere with sulphur dioxide. Heavy and radioactive metals, rendered more soluble on ignition, also find their way into surface and groundwater. The development of satisfactory methods for dealing with the associated rocks, especially *Dictyonema* argillite, is a matter of the highest priority (Veiderma 1993). It is extremely important to find a solution to this problem, because this determines whether opencast exploitation of the other Estonian phosphorite deposits (particulary at Toolse) will be introduced or not.

The Toolse and Rakvere deposits are situated in a densely populated and economically advanced region. On the one hand, the convenient location of the deposits deserves attention. On the other hand, one has to consider that these deposits are located in the Pandivere Upland from where a lot of rivers flow out and which is a water recharge area for groundwater aquifers. Introduction of phosphorite mining at the Rakvere deposit would call forth a situation where over a large area water basins would dry and the groundwater level of several aquifers would noticeably sink. Thus, for several reasons the agriculture and the whole life in a comparatively large area would be destroyed. The situation in the Toolse deposit is less complicated.

From the economic point of view, the production of concentrates from the Toolse and Rakvere deposits will not pay off. Production costs at Toolse may approach or exceed the level quoted for Kingisepp (Russia) concentrates: 38-56 USD per tonne (IFDC Report, 1993) which is considerably higher than, for instance, in Florida (less than 30 USD per tonne). One factor that might decrease the production costs at Toolse is the potential for low-cost electrical power and labour in Estonia.

Sociological factors could be very important for the further development of the Toolse deposit. Impact on demographic situation in the region will be minor if local people are hired. Some expropriation of land and relocation of residents are anticipated. However, the number of families directly affected by planned mining activities will not be known until a mining plan is prepared and a detailed land use survey is conducted (Environmental Review for Toolse Phosphorite Deposit, KBN, USA, 1993).

PEAT

A mire is part of the landscape where owing to permanent abundance of water and deficiency of oxygen some of organic matter remains decomposed and is deposited as peat. Peat is a fibrous substance, produced by the decay of vegetation in mires and it contains a high proportion of water (80...94%). Peat consists of oxygen, carbon and hydrogen; it also contains nitrogen, phosphorus and non-burning substances (Table 59).

Table 59. Characteristics of Estonian natural peat (Orru 1995, manuscript report)

	Deposit		
	Möllatsi	Keressaare	Laukasoo
C, %	52.4	52.9	58.6
H, %	5.7	6.8	6.9
N, %	2.7	1.0	2.2
S, %	0.27	0.13	0.22
Ash content, %	9.5	3.9	5.6
Calorific value (MJ/kg) at 50% moisture content			
	7.5	8.0	9.2

The composition and characteristics of peat depend on the conditions under which it was formed (nutrition of soil by groundwater, precipitation, or both), and on the location of the deposit in the landscape (lowland, hollow, river valley, *etc.*).

In comparison with the fuel peats of Ireland which belong to those most intensively studied in the world, Estonian peats have a higher ash content (Table 60). They also contain a lot of stumps (Photo 65), while the peats of Ireland have almost none. The fuel peats are represented mainly by grass peat in Ireland and wood-grass peat in Estonia.

Table 60. Fuel peat characteristics (Orru 1995, manuscript report)

	Estonia	Ireland
C, %	53.5	57.5
H, %	6.2	5.5
N, %	2.4	1.5
S, %	0.23	0.3
Ash content, %	8.0	1.5
Degree of humification	H_4-H_5	$H_{3.5}$-$H_{4.5}$
Temperature of ash deformation	1100°C	1100°C
Calorific value, MJ/kg	7.8	7.7

According to E. Lappalainen, *Sphagnum* and *Hypnum* are the main peat-forming species in Finland, but also sedge, cotton-grass and wood peats occur. In raised bogs, *Sphagnum* peat with a degree of humification $H_{3.5}$ - $H_{4.5}$ and the ash content 1-3%, forms up to 70% of peat deposits in Finland; in fens, sedge peat is most widespread. The fens of Estonia contain besides sedge also reed and wood peats. The peat deposits of Estonian raised bogs are prevailed by poorly-humified moss peat, well-humified peat being of a rather limited distribution. In Finland and Belarus, well-humified *Sphagnum* and pine-cotton-grass peat are most common.

Peat is used for making fuel briquettes, for soil improvement and as a substrate in horticulture. Potential fields of peat application are much more numerous (Fig. 221). In Estonia, peat has been used as fuel since long. From the beginning of the 19th century, it has also been used as litter. The very first peat industry was established at Ülemiste in 1913. However, about organized peat production one can talk not until 1919, when several peat societies were founded (Luha 1946). In the pre-war Estonia, there were two peat fuel based power stations - Ellamaa (1923-66) and Ulila (1923-55). The first peat briquette producing plant was put into operation at Tootsi in 1938, afterwards also at Oru (1964) and Sangla (1976).

Until the 1970s, during the years under Soviet occupation, 60% of the peat produced was used as litter for livestock and 40% as fuel. Of ca. two million tonnes produced annually in later years, some 10% was used as a substrate in horticulture and as a fertilizer.

Estonia is considered as a country richest (Fig. 222) in peatlands in North Europe. The total area, occupied by 9836 mires, covers one million hectares. Of 1626 peat deposits of commercial interest (*ca.* 0.9 million hectares), 520 are large mires with a thick layer of peat which are of major significance in a number of respects, including commercial and agricultural use, and nature conservation. The remaining 1106 peat deposits are mostly fens with a thin layer of peat. Peat deposits with an area of more than 10 ha and the thickness of peat layer over 0.9 m belong to the deposits of commercial importance (Orru 1995). The average thickness of the peat layer in Estonian mires is 3-4 metres.

Formation of mires started after the territory of Estonia was freed of the ice cover, first in Upper and later in Lower Estonia. Through the whole postglacial period, climatic conditions have favoured the rise and development of mires. In Estonia, mires formed mostly as a result of paludification of mineral soils or filling up of shallow bodies of water. On the basis of the origin and peat-forming species, mires are classified as fens and raised bogs with a transitional mire between them. A fen forms in a low-lying area where the soil is rich in nutrients, provided by groundwater, and a luxuriant carpet of

Photo 65. Estonian peat often contains stumps. Stump piles in the Imsi bog. *Photo by M. Orru.*

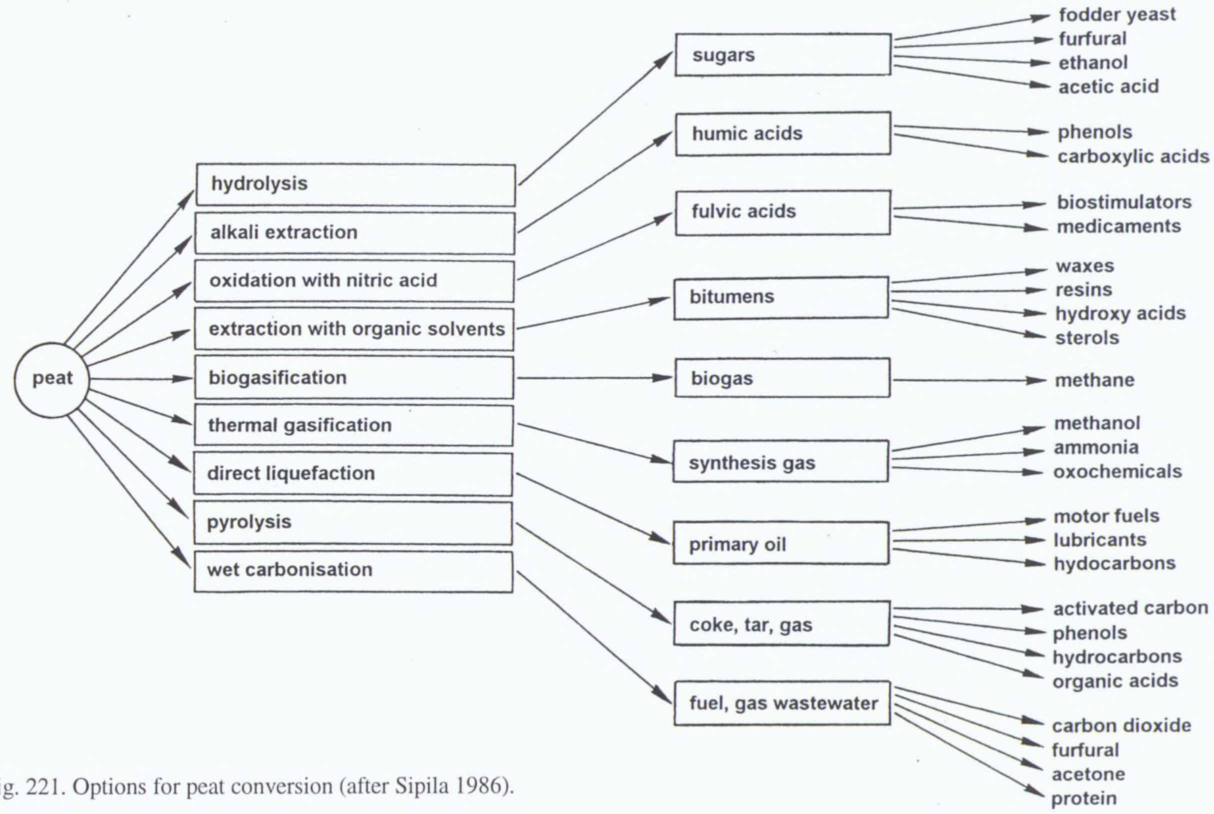


Fig. 221. Options for peat conversion (after Sipila 1986).

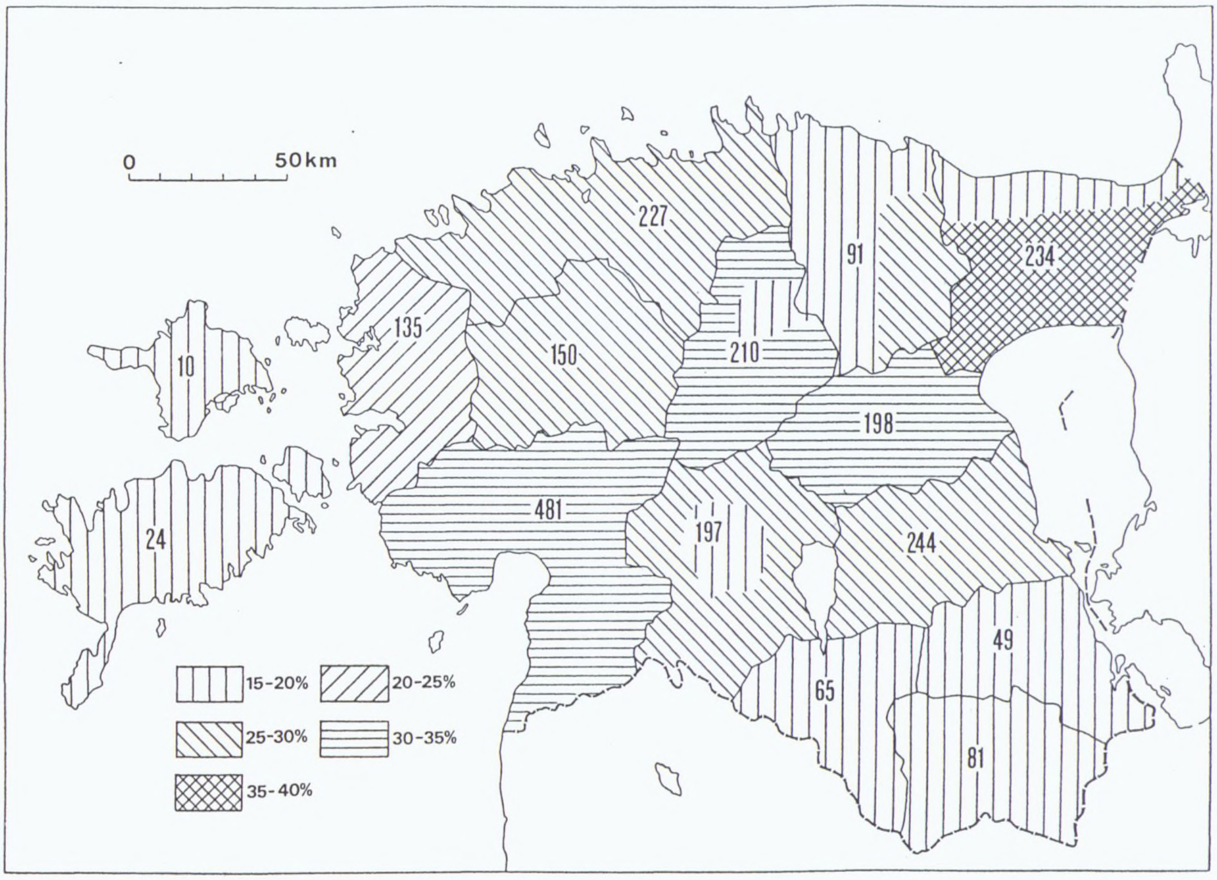


Fig. 222. Paludification (%) and the total amount of peat resources (million tonnes at 40% moisture content) in Estonia by counties (after Orru 1996).

numerous species of hydrophilous plants grows on it. The main peat-forming plants are sedges, reed, horsetails, mosses and several species of trees. The ash content of eutrophic peat is high. A raised bog's only supplementary source of minerals is precipitation, the nutrition provided by this kind of "airlift" is inevitably rather meagre and all plants growing in bogs have to cope with a substrate that is extremely poor in nitrogen and lime. Peat mosses flourish under such conditions and provide the major ingredient for peat. To a lesser extent, there occur horsetail, *Scheizeria* and pine. The ash content of ombrotrophic peat is low. A transitional bog is fed by groundwater and precipitation, and promotes the development of both fen and raised bog plants. It also provides favourable habitats for cranberries.

The study of mires goes back to the 19th century. The first studies (Bode 1836, 1837) dealt mostly with the mire vegetation. In 1910, an experimental station was founded at Tooma with an aim of studying the prospectives of mire usage for agricultural purposes. Vellner (1922) published the data and maps on 517 largest mires in Estonia. The later studies dealt with peat increment, palynology, vegetation, hydrology, genesis and ecology of mires (Thomson 1933, Allikvee & Masing 1988, Masing 1988b, Ilomets 1995a). More detailed data on the distribution, area, thickness, characteristics and reserves of peat deposits are available mainly in manuscript exploration reports and, in generalised form, also in several publications (Raudsepp 1946, Труу и др. 1961, Valk 1988, Orru 1992, Eesti sood 1993, Orru 1995).

PEAT DISTRICTS

A peatland is a multi-faceted and developing landscape phenomenon. Peatlands are classified on the basis of a variety of features, such as stratigraphical, geobotanical, the composition of peat layers, genesis, *etc.* The Estonian peatlands are divided into 20 districts (Orru 1992, Fig. 223).

1. North-Estonian Coastal Lowland

The district contains transitional mires, mixed and raised bogs with only a thin layer of peat. The eutrophic and mesotrophic deposit consists of well-humified (up to 50%) wood and reed peats, with the moisture content 78-84%, ash content 12-25% and a slightly acid reaction (pH 3-6). The mixed deposit is well humified (up to 40%). The moisture content is 88-92% and the concentration of minerals is low (ash content 1-2%). A somewhat higher mineral matter concentration (2%) in the upper part of the deposit is evidently due to the admixture of dust originating from the Kunda Cement Works. The ombrotrophic peat deposit consists prevailingly of poorly or moderately (8-28%) humified wood-*Sphagnum* and *Sphagnum* peats with acid reaction (pH 3-4). The moisture content is constantly 90-91% throughout the deposit, the average ash content is 1%.

2. North-West Estonian Plateau

The eutrophic and mesotrophic peat in the district is made up of well-humified (38-40%) wood peat, rather low in moisture (80-88%), high in minerals (ash content up to 25%) and slightly acid (pH 5.5-6). The mixed and ombrotrophic deposit consists of medium- to well-humified *Fuscum* peat with a stable moisture content (93-94%) and high concentration of minerals (ash content up to 15%). Both the mixed and ombrotrophic peats are well humified - 28 and 40%, respectively.

Viru, Rae, Pääsküla, Valdeku, Ellamaa, Valgjärve, Kadasoo and Hagudi are the most characteristic mires in the area (Figs. 224, 225).

3. North-East Estonian Plateau

On the North-East Estonian Plateau, larger peat deposits occur only in the areas where the runoff on the flat terrain is obstructed, for instance, by kame fields. The eutrophic and mesotrophic deposit consists of well-humified (40-45%) wood-reed peat, in which the moisture content is low (72-85%), the concentration of minerals is high (ash content 10-24%) and the reaction is slightly acid (pH 5-6). The mixed deposit is rather thin (0.7-0.8 m), formed of well-humified (up to 40%) wood-*Sphagnum* peat, changeable in moisture content (72-88%). The concentration of minerals (ash content) is 2-10%, and the reaction is acid. Oligotrophic peat forms the major part of the area's peat reserves. The lower part of the deposit consists of moderately (25%) and the upper part of poorly (10%) humified moss peat. The moisture content of ombrotrophic peat is 90-92%, the ash content averages 2% and the pH value is around 3.

Rannu, Kõrgesoo, Uljaste, Hiie, Peeri, Kure, Voorepere, Kunda, Varuli and Laukasoo are the major mires in the area under consideration (Fig.224).

4. Kõrvemaa

The **Kõrvemaa watershed area** is a region of large mires, including Kastna, Mukre, Palasi, Selja and Nõlvasoo, where the thickness of the layer of peat reaches 5.0-5.5 m (Figs. 223, 224).

The eutrophic and mesotrophic deposit consists of a thin

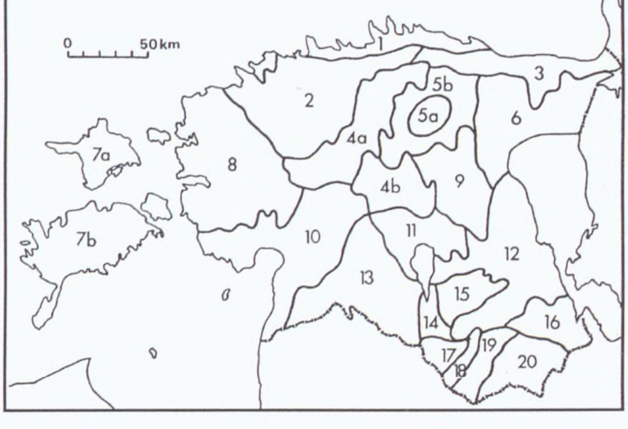


Fig. 223. Estonian peat districts: 1 - North-Estonian Coastal Lowland; 2 - North-West Estonian Plateau; 3 - North-East Estonian Plateau; 4 - Kõrvemaa (4a - watershed area, 4b - upper course of the Pärnu River); 5 - Pandivere Upland (5a - central part, 5b - slopes); 6 - Alutaguse; 7 - West-Estonian Archipelago (7a - Hiiumaa, 7b - Saaremaa); 8 - West-Estonian Lowland; 9 - Vooremaa; 10 - Pärnu Lowland; 11 - Võrtsjärv Lowland; 12 - South-East Estonian Plateau; 13 - Sakala Upland; 14 - Väike-Emajõgi Valley; 15 - Otepää Heights; 16 - Palumaa; 17 - Valga Lowland; 18 - Karula Upland; 19 - Võru-Hargla Depression; 20 - Haanja Heights.

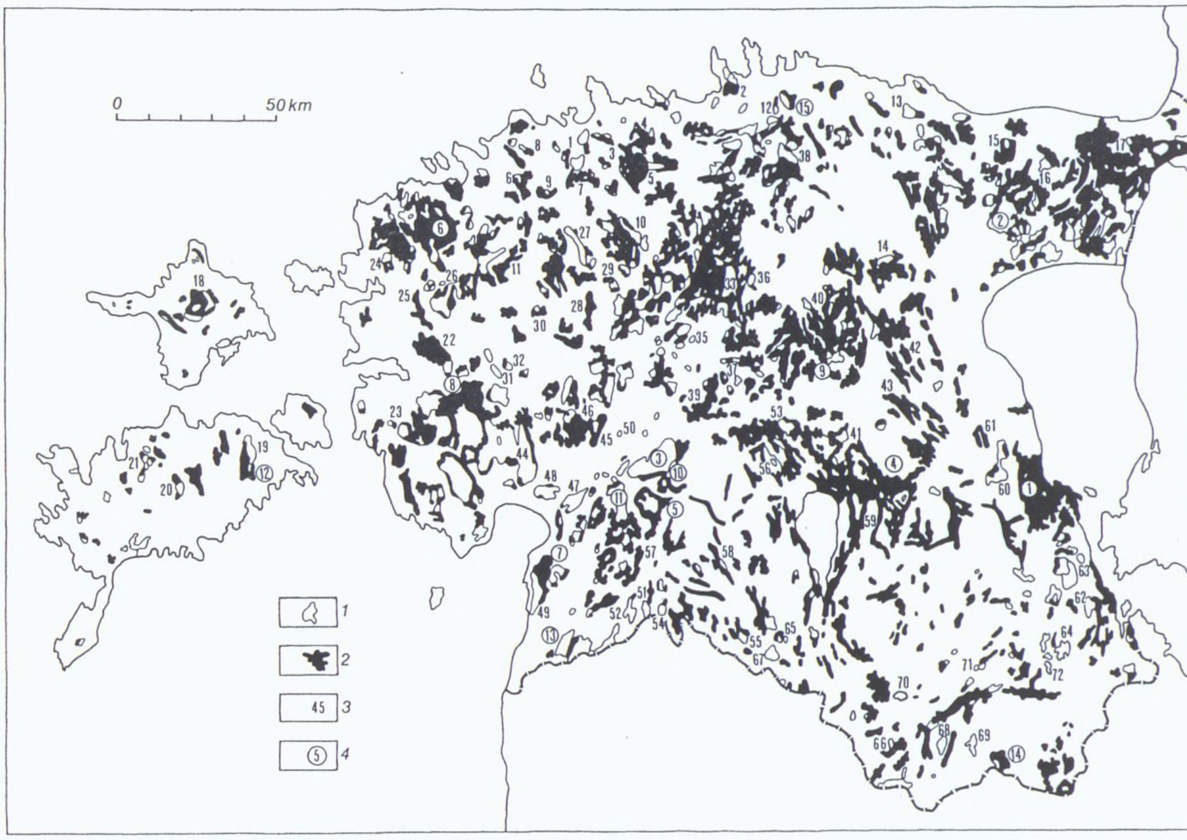


Fig. 224. Estonian peat deposits: 1 - fen, 2 - raised bog, 3 - number of the peat deposit in use; 4 - number of the protected mire. Peat deposits in use: 1 - Pääsküla; 2 - Hara; 3 - Rae; 4 - Kostivere; 5 - Peningi; 6 - Ohtu; 7 - Sausti; 8 - Vääna; 9 - Ääsmäe; 10 - Mahtra; 11 - Ellamaa; 12 - Uuemõisa; 13 - Varudi; 14 - Peetla; 15 - Hiiesoo; 16 - Peeri; 17 - Puhatu; 18 - Pihla; 19 - Koigi; 20 - Piila; 21 - Pelisoo; 22 - Laiküla; 23 - Kõverdama; 24 - Niibi; 25 - Turvalepa; 26 - Õmma; 27 - Hagudi; 28 - Keava; 29 - Hõreda; 30 - Orgita; 31 - Tõnumaa; 32 - Päärdu; 33 - Epu-Kakerdi; 34 - Retla; 35 - Lokuta; 36 - Tondissaare; 37 - Epa-Vassaare; 38 - Ohepalu; 39 - Kallissaare; 40 - Endla; 41 - Umbusi; 42 - Kivijärve; 43 - Visusti; 44 - Lavassaare; 45 - Mõrdama; 46 - Pööravere; 47 - Kõrsa; 48 - Rääma; 49 - Tolkuse; 50 - Kavasoo; 51 - Viirasoo; 52 - Mõksi; 53 - Soosaare; 54 - Pätsi; 55 - Ikepera; 56 - Parika; 57 - Napsi; 58 - Õisu; 59 - Sangla; 60 - Laukasoo; 61 - Keressaare; 62 - Tuurapera; 63 - Meelva; 64 - Meenikunno; 65 - Helme; 66 - Kantsi; 67 - Lagesoo; 68 - Roosa; 69 - Põdrasoo; 70 - Kungjärve; 71 - Kurgsoo; 72 - Pindi. Main protected mires: 1 - Emajõe-Suursoo; 2 - Muraka; 3 - Kuresoo; 4 - Emajõe-Pedja; 5 - Ördi; 6 - Suursoo; 7 - Tolkuse; 8 - Avaste; 9 - Endla; 10 - Valgaraba; 11 - Kikepera; 12 - Koigi; 13 - Nigula; 14 - Kellamäe (Vanamõisa), 15 - Laukasoo.

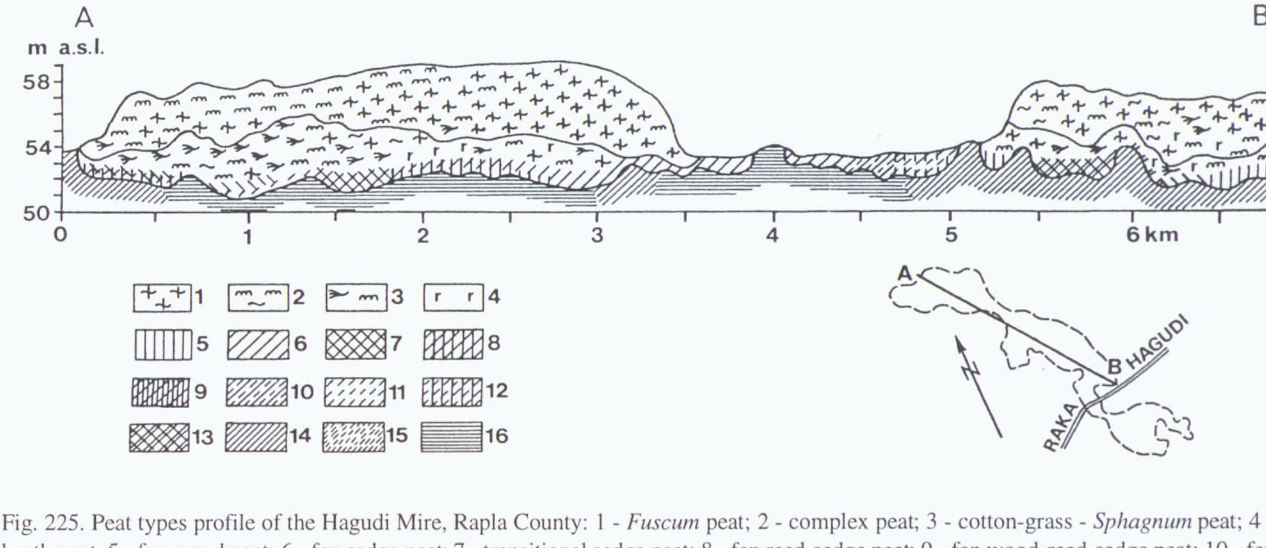


Fig. 225. Peat types profile of the Hagudi Mire, Rapla County: 1 - *Fuscum* peat; 2 - complex peat; 3 - cotton-grass - *Sphagnum* peat; 4 - heath peat; 5 - fen wood peat; 6 - fen sedge peat; 7 - transitional sedge peat; 8 - fen reed-sedge peat; 9 - fen wood-reed-sedge peat; 10 - fen sedge-*Hypnum* peat; 11 - fen *Hypnum* peat; 12 - fen wood-*Hypnum* peat; 13 - transitional grass peat; 14 - loam; 15 - sandy loam; 16 - clay.

(0.4-0.5 m) well-humified (40-45%) wood peat layer. Its moisture content reaches 80%, the content of minerals (ash content) is 3-14% and the pH value is 4-5. The layer of mixed peat is thin (0.7-0.8 m) and of limited distribution. Its moisture content averages 90%, the ash content is 1-4%, and the pH value is commonly 3. Ombrotrophic peat dominates forming some 80% of the deposits in most of the mires. It consists of poorly humified (11-15%) cotton-grass - *Sphagnum* and *Sphagnum* peats. The moisture content (90-92%), ash content (1-4%) and the pH value (3-4) are constant throughout the deposit and the district.

The area on **the upper course of the Pärnu River** (Fig. 223) holds the lands of the western and southwestern parts of the Järva County and the southeastern part of the Rapla County. In the region, raised bogs and ombrotrophic peat deposits cover about one third of the total area occupied by mires and peat deposits. In places, ombrotrophic peat forms the entire thickness of the deposit. Eutrophic and mesotrophic, mainly wood and wood-reed peats, form only a small part (0.5-0.6 m) of the deposit. The moisture content is 85-90%, the content of minerals is 3.5-11% and the reaction is acid (pH - 5). Mixed peat forms a thin (0.6-0.7 m) layer, which consists of medium- to well-humified wood-*Sphagnum* peat. The degree of humification fluctuates within a wide range (20-40%), while the moisture content is stable enough (88-92%), the ash content is low (1-1.5%) and the reaction is acid (pH 1.5-3).

Rumbi, Idva, Retla, Kallissaare - Lubjaahju, Nõmme, Matussaare and Laugasoo are the most typical mires in the area (Fig. 224).

5. Pandivere Upland

In the **central part of the Pandivere Upland** the groundwater table is deep below the ground, and the surface water percolates directly into the fissured and karsted bedrock. The hydrographic network is poorly developed and the number of peatlands is small. The largest, Savalduma Mire is situated in the vicinity of the Savalduma Karst Field. The other mires are concentrated in the area of the Neeruti eskers and the Loobu River. The eutrophic and mesotrophic deposit (0.9-1.2m) is represented by well-humified wood and wood-sedge peats. The deposit has a low moisture content (80-85%), the ash content is 1.5-1.7% and the reaction is acid (pH 4-5). Mixed and ombrotrophic peats are not so widespread as eutrophic and mesotrophic peats, because the majority of mires have not yet reached the raised bog stage. Only in larger mires the poorly humified *Sphagnum* peats form up to half of the deposit's thickness. The moisture content is 90% throughout the deposit, the ash content is 1.5-4% and the pH value is 3-4.

The **slopes of the Pandivere Upland** hold the lands of the central and eastern parts of the Järva County and the central part of the Lääne-Viru County. The area abounds in large mires. On the slopes of the upland, the water appears at the surface in the form of springs supporting the development of fens. The fens in small depressions, which have already reached the raised bog stage, have merged to form spacious mosaic mire systems. Epu-Kakerdi, Mahtra, Juuru, Viirika, Pakasjärve, Vistjärve-Jalametsa, Peetla, Punasoo, Sämi, Kabala, Tudu, Saara, Neeva-Prandi and Kallissaare-Lubjaahju are the most typical representatives of such fens (Fig.224).

The deposit of eutrophic and mesotrophic peat (1.2-1.4 m) consists mainly of medium-humified sedge and reed-sedge peats. The moisture content is constantly around 90%, the ash content is 4-12% and the pH value is 4-5. Mixed and ombrotrophic peats account for about two thirds of the total thickness of the deposit. They are made up of poorely decomposed (15-30%) cotton-grass and moss peats. The moisture content is constantly around 92%, the content of mineral substances is low (ash content 1.5-6%), the reaction is strongly acid (pH 2-3).

6. Alutaguse

The Alutaguse District extends over the northern part of L. Peipsi depression, the evolution of which was highly controlled by proglacial lakes located in this area. In the northern part of the area, which was freed from the waters of the ice-dammed lake rather rapidly, the bottom relief of the peat deposits is undulating and generally follows the bedrock topography. Puhatu (57,000 ha), the largest mire system in Estonia, and the Sirtsi (4680 ha) and Muraka (12,790 ha) mires are situated in the northern and eastern parts of the area. As to the percentage of land covered by mires (50%), the Alutaguse peat district ranks first in Estonia (Figs. 222, 223). The central and southern parts of the area are featured by eskers, kames and numerous old beach ridges of L. Peipsi, now covered with dunes. In places, the long esker ridges are jutting out into the raised bogs as long peninsulas of mineral soils. Fens and transitional mires occur on the fringe of the esker chains. Small mires where ombrotrophic peat forms the whole thickness of the deposit, are numerous in the area of dunes.

Eutrophic peat consists of poorly humified (30-31%) sedge and reed-sedge peat with the moisture content of 88-93%. Like in all eutrophic deposits, the ash content varies within a wide range (5-12%), the reaction is acid (pH 4-5). Mesotrophic peat forms thin (0.3-0.4 m) layers of limited distribution. The degree of humification is 30-35%, the moisture content is 90% and the reaction is acid (pH 4). The mixed and ombrotrophic deposits consist of cotton-grass - *Sphagnum* and *Sphagnum* peat layers. The degree of humification is 10-25%, the moisture content 90-93% and the reaction is more acid (pH about 3) than in the above-described deposits.

7. West-Estonian Archipelago

The eutrophic and mesotrophic deposits consist mostly of well-humified (35-50%) reed and wood-reed peats and, to a lesser extent, also of less-humified reed - *Sphagnum-Scheizeria* peats. The thickness of the peat deposit is 1.2-1.5 m, the moisture content 85%, the ash content 8-16%, and the pH value 4.5-5.5. Mixed and ombrotrophic peats in the uppermost part of the deposit are represented by poorly- humified (16%) *Fuscum* peat which in the lower part of the deposit is substituted by moderately (20-35%) to well (50%) decomposed mesotrophic wood peat. Throughout the deposit, the average degree of humification is 20-35%, the moisture content 90% and the ash content 2.5-2.6%. The ash content in ombrotrophic peat is higher than the average (1-2%).

Pihla, Määvli and Õngu are the largest mires in Hiiumaa, and Koigi, Piila and Pelisoo in Saaremaa (Fig. 224).

8. West-Estonian Lowland

The eutrophic and mesotrophic peats of the West-Estonian Lowland belong to the swamp and forest-swamp subtype. The thickness of eutrophic peat ranges from 2.5-3.0 m. Mesotrophic peat is either absent or forms a very thin (0.3-0.4 m) layer. Generally, the deposits consist of moderately- to well-humified (occasionally up to 40%) sedge, reed-sedge, sedge-*Sphagnum* and sedge-*Hypnum* peats. The degree of humification varies in a wide range throughout the deposit. As an average, the moisture content is 88%, the ash content is 5-8%, reaching 13-22% only in the bottommost layers. The reaction is slightly acid (pH about 5).

The mixed peat deposit in the mires of the West-Estonian Lowland belongs mostly to the swamp subtype and is variegated in structure. Peat of bog hollows and *Fuscum* peat make up most of the ombrotrophic deposit.The mesotrophic deposit consists mostly of *Sphagnum* and sedge-*Sphagnum* peats and occurs as a thin (up to 0.5 m) layer under the ombrotrophic deposit, or extends as far as the mineral land. The mixed deposit has an average moisture content 92-93%, it contains 6% mineral substances and has a acid reaction (pH 3-4).

The ombrotrophic deposit usually occurs in large mires - Marimetsa, Niibi, Mustjärv, Palivere (Fig.224) and consists mainly of *Fuscum* peat. In some places, the ombrotrophic peat is underlain by a thin (0.5-0.6 m) layer of mesotrophic peat, dominated by *Sphagnum* and heath-*Sphagnum* peats. Eutrophic peat, forming the lowermost part of the section, consists of sedge, wood-sedge and wood peats. The ombrotrophic deposit, as a whole, is characterised by an evenly high moisture content (up to 96%), the content of mineral substances is 2-3% and the pH value is 3-3.5.

9. Vooremaa

In the Vooremaa district, the eutrophic and mesotrophic deposit consists of wood and wood-reed peats, the moisture content is 85-86%. The content of ash varies significantly (6-15%), evidently due to the surface water percolating from the slopes of drumlins into the mires. The eutrophic and mesotrophic deposit has an average thickness of 2.5 m. The degree of humification ranges from 28 to 40%, the pH value is 4.5-5.5. In the region under consideration, mixed deposit is rather widespread. The upper and central parts of the deposit consist prevailingly of poorly-humified *Fuscum* peat, the bottommost layers of poorly decomposed mesotrophic sedge-*Sphagnum* or sedge peats, to a lesser extent, also of eutrophic sedge peat. The content of moisture and ash, as well as the pH value of the mixed peat is much the same as in ombrotrophic peat.

Practically in all mires, the lowermost 3.5-3.8 metres of ombrotrophic peat consist of well-decomposed wood, often also of reed-*Hypnum* and *Hypnum* peats. The remaining part of the deposit is made up of poorly-humified (10-12%) moss peat (*Fuscum, Angustifolium, Medium etc.*). The moisture content of ombrotrophic peat is usually 90-92%, the ash content is around 3%. Eutrophic peat prevails throughout the Vooremaa area, however, the mixed deposit is also rather widespread.

10. Pärnu Lowland

The Pärnu Lowland is primarily a region of large mires (32 % of its area), which include Lavassaare, Nigula, Võlla, Tolkuse, Rääma, Kõrsa (Figs. 224, 226), Avaste and Mõrdama. Ombrotrophic peat, mostly poorly humified, makes up 72% of the whole peat deposit in the Pärnu Lowland. In fens the peat layer is thin, with a dissected contour and islands of mineral soils. The eutrophic and mesotrophic deposit consists for the most part of rather poorly humified (20-30%) reed-sedge and sedge peats. The moisture content (88-90%) is stable throughout the section. The content of mineral substances is rather low (ash content 5.5-8%), the pH value is in a range of 3.5-4.5. The mixed and ombrotrophic deposits have formed of medium- to poorly-decomposed pine-cotton - *Sphagnum* and moss peats (*Sphagnum, Fuscum, Medium*). The moisture content of the deposit is constantly 90-91%, the concentration of mineral substances 20% and the reaction is strongly acid (pH 2-3).

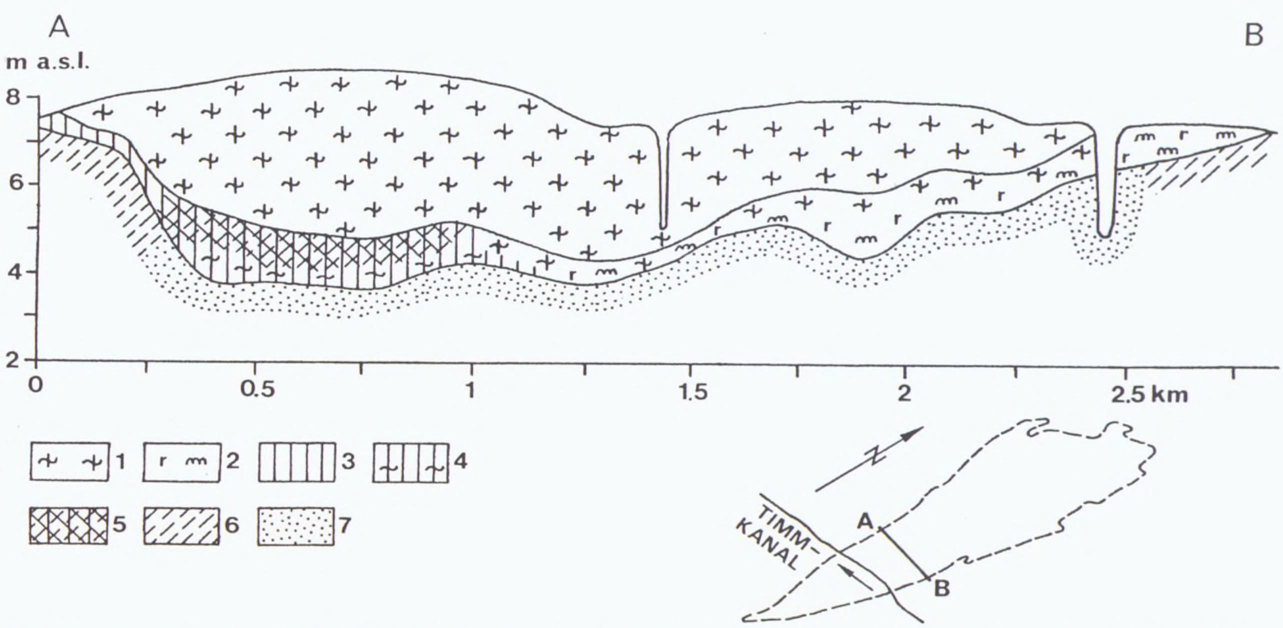


Fig. 226. Peat type profile of the Tolkuse Mire, Pärnu County: 1 - *Sphagnum-Fuscum* peat; 2 - heath-*Sphagnum* peat; 3 - fen wood peat; 4 - fen wood-*Sphagnum* peat; 5 - transitional wood-*Sphagnum* peat; 6 - loamy sand; 7 - sand.

11. Võrtsjärv Lowland

Mires cover some 30 per cent of the low-lying area around L. Võrtsjärv. The largest mires are Soosaare, Sangla, Parika, Umbusi and Kohvisoo (Fig. 224).

The upper layers of the eutrophic and mesotrophic deposit consist often of moderately humified heath-*Sphagnum* and wood-sedge peats, which are underlain by well-humified wood-reed and wood peats. The moisture content of eutrophic peat is rather high, reaching occasionally 90%. The ash content is generally 5.5-9.5%, but in the mires within river valleys it may even reach 21-22%. The pH value of eutrophic and mesotrophic deposits is 5-6.

In the Võrtsjärv Lowland, mixed peat is not so widely spread as ombrotrophic peat. The upper layers of the mixed deposit consist of moderately humified *Magellanicum* and pine-cotton peats underlain by well-humified wood-grass peat. The bottom layers of the deposit consist of moderately- to well-humified sedge peat. The moisture content of the whole deposit is ca 90%, the ash content is 1.5-4.5% and the pH value is about 3. The ombrotrophic deposit consists mostly of poorly humified *Fuscum* peat, followed by pine-cotton - *Sphagnum* peat. Ombrotrophic peat rests upon mesotrophic *Scheizeria*, grass-*Sphagnum*, sedge, wood-reed or *Hypnum* peat. The average moisture content of the ombrotrophic peat is 90%, the ash content is permanently 1.5-2.0% and the pH value is 3-3.5. In conclusion, it may be said that the ombrotrophic peat deposit in the Võrtsjärv Lowland is low in the moisture content and the proportion of the *Sphagnum* peats is high.

12. South-East Estonian Plateau

In this area, all larger valleys are occupied by mires, mostly fens like, for instance, Emajõgi-Suursoo. Raised bogs, including Laukasoo, Keressaare, Rihtemetsa, and Essaksoo, have developed in the watershed areas dissected by valleys (Fig. 224). In raised bogs, the peat layer reaches 7.5 m in thickness.

Eutrophic peat, 1.9-2.7 m thick, varies in composition. In the Emajõgi-Suursoo mire it is represented by moderately humified swamp peat. In other fens, peats of the forest subtype are also encountered. Usually, the upper layers of the deposit consist of sedge, wood-sedge, wood-reed and occasionally also of wood and wood-*Hypnum* peat. The lowermost layers of the deposit are dominated by well-humified wood-reed, wood-*Hypnum*, reed and *Hypnum* peats. The moisture content of the eutrophic deposit is 80-85% and the pH value is 5.5-6.5. The concentration of mineral substances varies significantly within the section (7-17%) due to flood water.

The mesotrophic peat deposit is not so widespread as ombrotrophic peat and consists of wood-*Sphagnum* peat, 0.7-0.8 m in thickness. In terms of the quality indices, it is close to the mixed peat deposit.

The mixed deposit (average thickness 4.5-5.5. m) consists of poorly- to moderately-humified *Fuscum* and *Medium* peats in the upper part and pine-cotton and *Sphagnum* peats in the lower part. The ash content is 4-5%, the pH value 3.5-4.5

13. Sakala Upland

In the Sakala Upland, which comprises the central and southern parts of the Viljandi County, mires cover some 15% of the land surface. The central part of the plateau-like upland is dissected by numerous ancient valleys. The thickness of the Quaternary cover ranges from 3-10 m in the northern to 40-50 m in the southern part of the upland (Раукас 1978). Most of the upland – the areas between the ancient valleys – has a plateau-like, poorly dissected topography. As a result, the number of mires in the upland is much smaller than in the areas with a highly dissected topography, *e.g.* the Otepää, Haanja and Karula heights. The most characteristic mires in the area include Raudna, Vennissaare, Õisu, Halliste, Saaretsi, Umbsoo, Veelikse, Napsi, Pahuvere and Lillesoo (Figs. 224, 227). Napsi, with the thickness of the peat layer up to 11.0 m, is one of the deepest mires in Estonia.

The topography of the eastern part of the upland is more levelled than in the central part. The areas with a greater relative height hold small mires, up to 10 hectares in area. Large raised bogs, like Rubina, Ikepera and Lagesoo have formed on the undulated till plains skirting the Väike-Emajõgi Valley. In the eastern part of the Sakala Upland, the eutrophic and mesotrophic deposit consists prevailingly of well-humified (30-45%) wood-sedge peat. The thickness of the peat is usually 3.0-3.5 m, the content of mineral matter ranges from 8 to 19%, the pH value is 5-7, and the moisture content is around 80%.

In the central part of the Sakala Upland, the eutrophic and mesotrophic deposits consist of well-humified (up to 50%) swamp, wood and wood-sedge peats. The moisture content of the deposits is rather low (80-88%). The concentration of mineral matter fluctuates from 3 to 12.5%, in the river valley mires it may reach even 21% due to the input by flood water.

Mixed and ombrotrophic deposit occurs mostly in the watershead area (Umbsoo, Saaretsi, Pätsi, *etc.*). It is represented by poorly humified (18-28%) *Fuscum* or *Sphagnum* peat. The lowermost part of the deposit consists of eutrophic or mesotrophic peat (wood-sedge or sedge-*Sphagnum* peat). The moisture content of the mixed and ombrotrophic deposit is constantly around 90%, the content of mineral matter is 1.5-4% and the pH value is 3-4.

The mixed peat deposit (1.5-1.6 m) consists mainly of

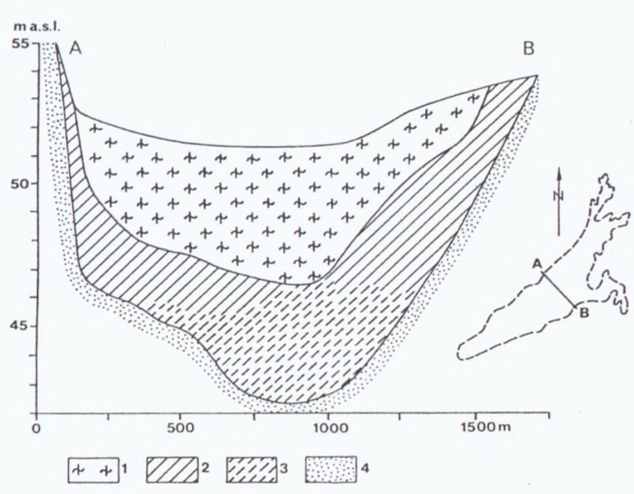


Fig. 227. Peat type profile of the Napsi Raised Bog, Viljandi County: 1 - *Sphagnum-Fuscum* peat; 2 - fen sedge peat; 3 - fen *Hypnum* peat; 4 - sand.

poorly (16%) humified ombrotrophic peat with a moisture content 85%. Its lower part is represented by moderately- to well-humified (20-45%) wood-sedge peat. In the mixed deposit the concentration of mineral matter is constantly 3-9% throughout the section.

The ombrotrophic deposit is composed of poorly-humified (10-20%) cotton-grass, grass and moss peats with an average thickness of 4.1-4.3 m. The moisture content is 85-89%, the concentration of mineral matter is 1-1.5% and the pH value is about 3.

14. Väike-Emajõgi Valley

The Väike-Emajõgi Valley covers part of the low-lying area between the Otepää and Sakala highlands. Variety is added to the scenery by kames, small lakes and abundant mires. Mires cover about 26% of land surface (Fig. 222). Most common are fens, with a strongly elongated contour and a deep layer of peat; of those Väike Emajõgi is the largest (Fig. 224). Moderately to well humified (24-45%) wood and wood-reed peats are widespread, sedge peats are also encountered. The ash content of eutrophic peat is around 10%. Ombrotrophic deposit consists of *Fuscum* and complex peats, the degree of humification is 15-30% and the ash content 2-4%. In quality, the peats of the Väike-Emajõgi Valley are close to the peats of the Emajõgi Valley.

15. Otepää Heights

Palupera, Maru, Pühajärve, Vidrike, Truuta and Pori are the most important mires on the Otepää Heights (Fig.224).

The eutrophic and mesotrophic deposit (thickness 3-4 m, maximum 7.5 m) is dominated by wood-sedge peat rich in reed. The moisture content of the peats is rather low (78-85%), the content of mineral matter is variable (4-20%) but in general high. The reaction of the deposit is slightly acid (pH 5-5.5). The ombrotrophic and mixed peats form thin (1.7-2.1 m) layers in limited areas of single mires. The lower part of the deposit consists of well-humified eutrophic peats, which are overlain by ombrotrophic peats. In the mixed deposit the degree of humification is up to 40%, while in the ombrotrophic peat it is 20-25%. The moisture content of the whole deposit is around 90%, the content of mineral matter ranges from 2-5.5% and the pH value is 3-4.

16. Palumaa

The Palumaa district is situated between the Haanja Heights and Lake Pihkva depression in the southeastern part of the Põlva County (Fig. 223). The Quaternary cover consists mostly of till and, to a lesser extent, of glaciolacustrine sediments. The terrain is flat owing to the wide distribution of abrasion and accumulation plains. Single valleys dissect the area into plateau-like patches. Under such conditions, the groundwater nutrition of the majority of mires between the river valleys was not possible and the only supplementary source of minerals was precipitation. Therefore, the development of the mires in this area started right from the raised bog stage. Raised bogs make up some 80% of the mires' total area in this region. Fens occur mostly in the river valleys rich in springs. The most typical mires in the area are Meelva, Meenikunno, Riha, Tedremäe, Timo and Valgesoo (Fig.224).

The eutrophic and mesotrophic deposit in the Palumaa area is represented by a thin (1.0-1.2m) layer of wood and wood-reed-sedge peats. The moisture content of the deposit is low (78-80%), the content of mineral matter ranges from 7 to 12%, the pH value is 5.5-6.5. The content of minerals in the lowermost part of the deposit is highly variable. Mixed and ombrotrophic peats make up about 80% of the deposits at Palumaa. The ombrotrophic peats are prevailed by poorly humified (20%) cotton-grass -*Sphagnum* and grass peats. The moisture content is more or less constant (92%) and the content of minerals is low (0.8-1.0%) throughout the section.

The Palumaa area is notable for its impressive raised bogs.

17. Valga Lowland

The Valga Lowland, as a southward extension to the Väike-Emajõgi Valley, stretches up to the Karula Upland. Administratively, it is divided between the Valga and Võru counties. The generally hilly and hollow topography determines the location of different types of mires in the district. The central part of the lowland holds flood plain mires with well-humified eutrophic peat, formed in the nutrient-rich environment. Small raised bogs occur in the marginal areas. Larger mires, like Väike-Emajõgi, Priipalu, Rulli, Pedeli, Kohvisoo and Korva, are fed by groundwater, in some places also by flood water. The raised bogs, such as Mehiksoo and Struuga are fed by precipitation (Fig. 224).

The eutrophic and mesotrophic deposits (2.0-3.0 m) consist mostly of well-humified (40-45%) wood or wood-sedge-reed peats. As a result of partial draining of the mires in the river valleys (Korva, Pedeli, Priipalu), the moisture content of the deposits is rather low (78-80%). The eutrophic peat has a high but fluctuating mineral matter content (ash content 5-24%); the pH value is 5.5-6.

The ombrotrophic deposit occurs mainly in the southern part of the lowland (Mehiksoo and Struuga mires). The lower part of the deposit consists of wood-rich grass peat, while moss peat forms the middle and upper parts of the deposit. The average thickness of the ombrotrophic peat is 4.5 m, maximum 8.5 (in the Mehiksoo mire). The moisture content is 90-92%, the degree of humification 15-22%, the content of mineral matter is rather low (ash content 1.5-2.0%), the reaction is highly acid (pH 2.5-3). Mixed deposit does not practically occur in the Valga Lowland mires.

18. Karula Upland

The Karula Upland has an extremely rugged topography, featured by numerous hills, tiny lakes and small and medium mires in the interstitial hollows. Both raised bogs and fens are encountered. Fens are numerous and fed by the surface water flowing down the steep hillsides. The most typical mires of the area include Kantsi, Lauksilla, Koobassaare, Kuutsi, Jahusoo and Sarbasoo (Fig. 224).

Eutropic and mesotrophic peats form the deposits with an average thickness of 4.5-5.5 m (max. 6.5 m) which consist mostly of wood and wood-reed peats. The moisture content of the deposits is generally 80-87%, the content of minerals is high and variable (5-13.5%), the reaction is slightly acid with the pH value being 5-5.5.

The lower part of the mixed deposit consists of moderately humified eutrophic peat, rich in sedge, overlain by cotton-grass or pine-rich ombrotrophic peat. The average thickness of the deposit is 2.8 m, the moisture content is constantly 90% throughout the section, the pH value is 4-5.

Ombrotrophic deposit is of the most limited distribution, being worthy of mentioning only in the Kantsi and Koobassaare mires on the southern fringe of the Karula Upland (Fig. 224). The deposit has an average thickness of 3.5-4.5 m and consists mostly of *Fuscum* peat which is characterised by a low degree of humification (15-16%), high moisture content (90-91%), and a low mineral matter content (ash content 1%). The reaction is acid (pH 3-3.5).

19. Võru-Hargla Depression

The Võru-Hargla Depression, which runs from northeast to southeast through the Valga and Võru counties, separates the Haanja Heights from the Karula and Otepää highlands. It is also traceable in the pre-Quaternary topography. In the Lateglacial it served as a stream bed for the meltwater flow. In the lower part of the depression, the river valleys hold mires where eutrophic peat dominates. The slopes of the depression, *i.e.* areas with a greater absolute height, provide good conditions for the surface water flow and, therefore, the mires being mostly fed by precipitation are in the bog stage. Raised bogs of medium size are frequent north of Nõnova and south of Tsooru. About 30% of the mires on the slopes of the depression are raised bogs, the percentage of transitional mires is almost as high. The most typical mires of the area are Kääpa, Kerreti, Pindi, Paidra, Võru, Kaugjärve, Roosa, Kurgsoo, Mustjõe, Põdra, Kuutsi and Pärnika (Figs. 224, 228).

The eutrophic and mesotrophic deposits, with an average thickness of 2.9 m, have formed of well-humified (32-41%) wood peat. The moisture content of the eutrophic peat is 78-84%, in the mesotrophic peat it is a bit higher. The eutrophic peat is rich in minerals (10-24%), but their content varies with the parts of the section. The mesotrophic deposit is more acid than the eutrophic deposit, the pH values being 4-5 and 5-5.5, respectively.

In the area under consideration, mixed deposit is represented by well-humified (40%) wood-sedge peat and poorly-humified (16%) *Fuscum*, *Medium* or complex peats. The deposit has an average thickness of 2.4 m, the moisture content 90%, the ash content 1-2% and the pH value 3-4.5.

The ombrotrophic peat consists mostly of poorly humified (15%) *Fuscum* peat and, to a lesser extent, of up to moderately-humified (18-25%) wood-bearing ombrotrophic peat. Throughout the ombrotrophic deposit, the moisture content is constantly 90-91%, the content of minerals is low (ash content 1-2%)) and the reaction is highly acid (pH 2-3).

20. Haanja Heights

The closed hollows in the Haanja Heights abound in small mires fed by groundwater. Eutrophic peats cover 80% and ombrotrophic peats 20% of the mires' area. The mires which are situated in ancient valleys and in larger depressions have a complex configuration. The mires in the contemporary river valleys are partially fed by flood water. The most characteristic mires of the Haanja Heights include Tika, Hino, Vanamõisa, Luha, Murati, Kirikumäe and Palu (Fig. 224).

The eutrophic deposit consists of forest and swamp peat subtypes, prevailed by well-humified (40%) wood peat. The moisture content of the layer is 85%, the content of mineral matter fluctuates within a wide range (ash content 5-15%) and the pH value is about 5. The content of mineral matter is, as a rule, highest in the lower layers of the deposit. The mesotrophic deposit is represented by a very thin (0.3-0.4 m) layer of peat or is non-existent. The mixed deposit is also of limited distribution. In mires, where it exists, it consists of moderately- to well-humified (35%) peats belonging to the swamp subtype. The ombrotrophic deposit is made up of cotton-grass - *Sphagnum* peat, with a thickness of 2.7-3.6 m. The moisture content of the deposit is 90%, the content of minerals 1-3% and the pH value is about 3.

In Estonia, peat is at its thickest (about 17 m) in the Vällamäe Mire at the foot of the Suur-Munamägi Hill in the Haanja Heights. The mire has an area of one hectare and has reached the raised bog stage. The deposit consists prevailingly of moss peat.

Peat resources

Estonia's total peat resources in 1626 deposits are estimated at 15.24 milliard m³ or 2.37 milliard tonnes, of which

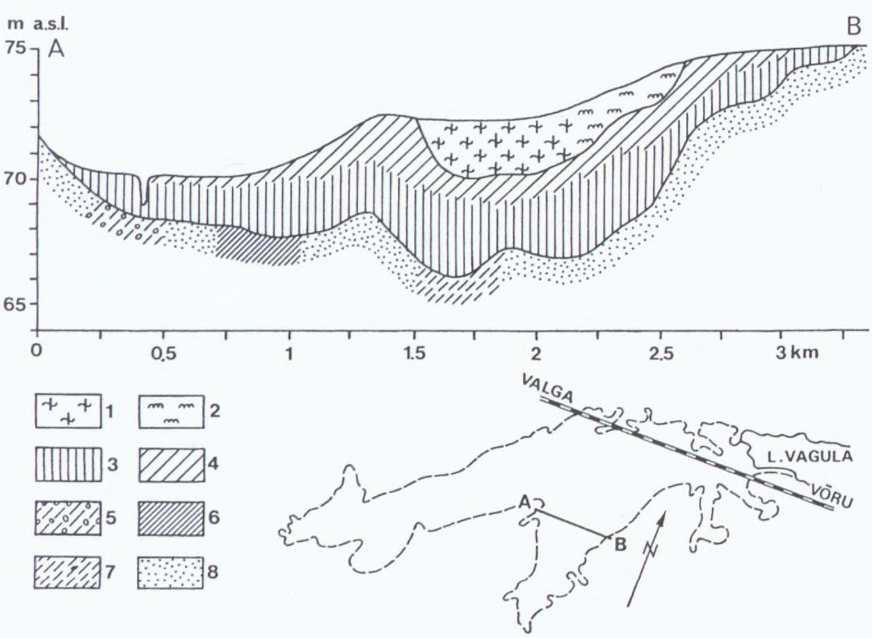


Fig. 228. Peat type profile of the Kerreti Mire, Võru County: 1 - *Sphagnum-Fuscum* peat; 2 - *Sphagnum-Medium* peat; 3 - fen wood peat; 4 - fen sedge peat; 5 - till; 6 - sandy loam; 7 - loamy sand; 8 - sand.

active (economically exploitable) resources constitute 1.52 milliard tonnes. 0.85 milliard tonnes occur under fields and meadows (Orru 1995). Currently, 69 mires (156 000 ha by area) are under conservation (Fig. 224, Kallas 1993). The reserves of peat suitable for use in horticulture and as litter are estimated at 0.2 milliard tonnes (Orru 1995). About 72 deposits are exploited. The peat is produced for litter, fuel and use in horticulture. The average thickness of peat is usually 3...4 m. The distribution of peat resources by counties is uneven (Fig. 222).

Peat is produced for use as litter in animal husbandry, substrate in horticulture and for fuel. It is also mixed with manure and used for improving agricultural soils. The area of peat milling fields (Photo 66) is 16,000 ha. In 1994, 8300 tonnes of sod peat, 490,000 tonnes of milled peat for litter and use in horticulture and 555,000 tonnes for manufacturing peat briquettes were produced (Juske 1995). The annual output of fertilizer peat is still modest: 190,000 tonnes from 16 deposits. Every year about 50,000 tonnes of Estonian milled peat is exported to Norway, Netherlands, Belgium, Spain and other European countries. During the last 3 years, the production of peat for litter, horticulture and fuel has decreased some 50%. The current extraction and drainage technology enable to use only 0.80 milliard tonnes or 40% of peat resources. The lower part of the deposit is below the recipient-river level and its extraction is possible only in the event of complicated polder drainage system, however, this would cause several technical, economic and environmental problems.

Besides the above-mentioned peatlands, there are a lot of small mires with a thin layer of peat which cover an area of 107,453 ha. Peat reserves of those small mires have not yet been estimated.

In Estonia, 156,562 ha or 16% of the mires are under protection (Fig.224). The total peat reserves of the protected mires are estimated at 3.5 milliard m^3 (Orru 1995).

Photo 66. Milling field in the Puhatu Bog. *Photo by M. Orru.*

As the natural accumulation of peat in mires is slow and natural sites for its accumulation are dwindling rapidly, the Government of the Estonian Republic has adopted the Regulations of Sustainable Use of Peat which enacts annual output quotas for every county.

It is expected that peat will find new uses in future, since it shows a promise for producing growth stimulators and peat wax. There are rich resources of well-humified eutrophic peat suitable for making growth stimulators (about 57 million tonnes), and the production technology is not very complicated either. The resources of cotton-grass peat for producing peat wax are estimated at 1.6 million tonnes.

Extensive peat production may damage the natural ecological balance. When planning peat milling in new fields, one has to take into consideration the ground-water regime and other local conditions to avoid the destruction of natural cranberry bogs. Cranberry plantations are to be laid out in the exhausted peat milling fields. Experiments of this kind have already been made in Estonia.

CRYSTALLINE ROCKS

Maardu granite deposit

The only explored granite deposit in the crystalline basement of Estonia occurs within the limits of the Neeme Pluton (intrusive) consisting of potassium-granites. The deposit is located at Maardu, 12 km east of Tallinn, to the north from the Tallinn-Narva highway on the coast of Muuga and Ihasalu bays where the depth of the crystalline basement is the smallest (minimum 120 m from the surface).

In the course of geological investigation, 36 boreholes reaching the porphyry-granites were made within the limits of the Neeme Pluton. The latter can be easily followed by the minimum values of magnetic and gravitational fields.

Like elsewhere in Estonia, two structural stages can be distinguished in the area of the deposit: the deeply eroded Precambrian crystalline basement is unconformably overlain by a monoclinal sedimentary cover (2-4 m/km, azimuth 160-180°). Structurally, the crystalline basement of this area belongs to the Tallinn Zone of the Svecofennian orogenic complex. In the crystalline basement, two major complexes with different genesis, age and structure can be distinguished (Fig. 229):

1) the Jägala Complex of Palaeoproterozoic metamorphic rocks (gneisses), primarily of sedimentary and sedimentary-volcanogenic origin which have been strongly folded and migmatized;

2) anorogenic intrusive rocks dating from the Mezoproterozoic and cut into the Jägala Complex - the potassium granites of the Neeme Pluton.

The Neeme Pluton resembles other porphyry plutons in northern Estonia (Naissaare, Ereda, Taebla) and the small potassium-granite massifs in southern Finland. The area of the pluton is about 120 km^2. There is no direct data available on its thickness but, in any case, it ought to be more than 1 km. The deepest borehole (F-503) penetrates to a depth of 210 m into the pluton.

Two types of granite (Fig. 230) can be observed within the pluton :

1) porphyry-potassium-granite forming the main part (95%) and occurring in 34 boreholes out of 36;

2) granite porphyry in the form of subvertical veins, the thickness of which ranges from some tens of centimetres to about ten metres. The contacts of the veins and the main granite

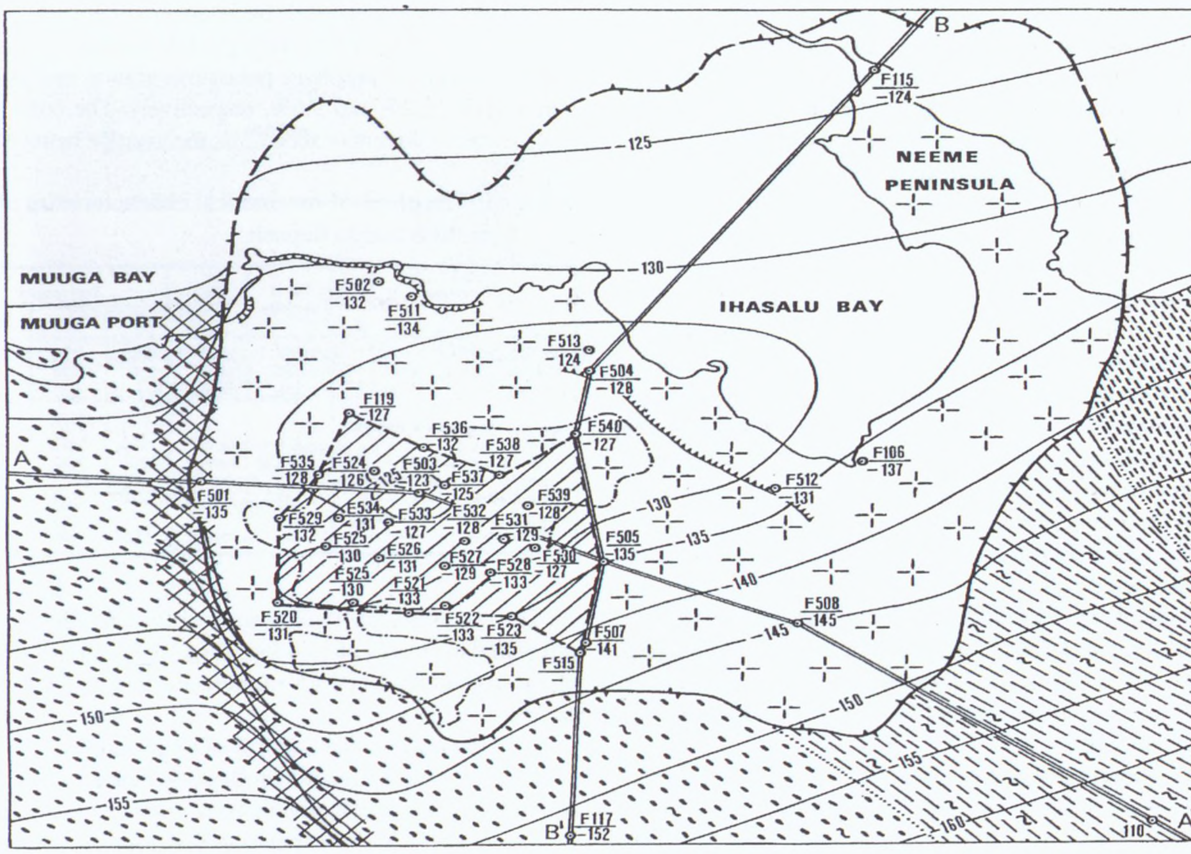


Fig. 229. Sketch-map of the Maardu deposit. For legend see Fig. 230.

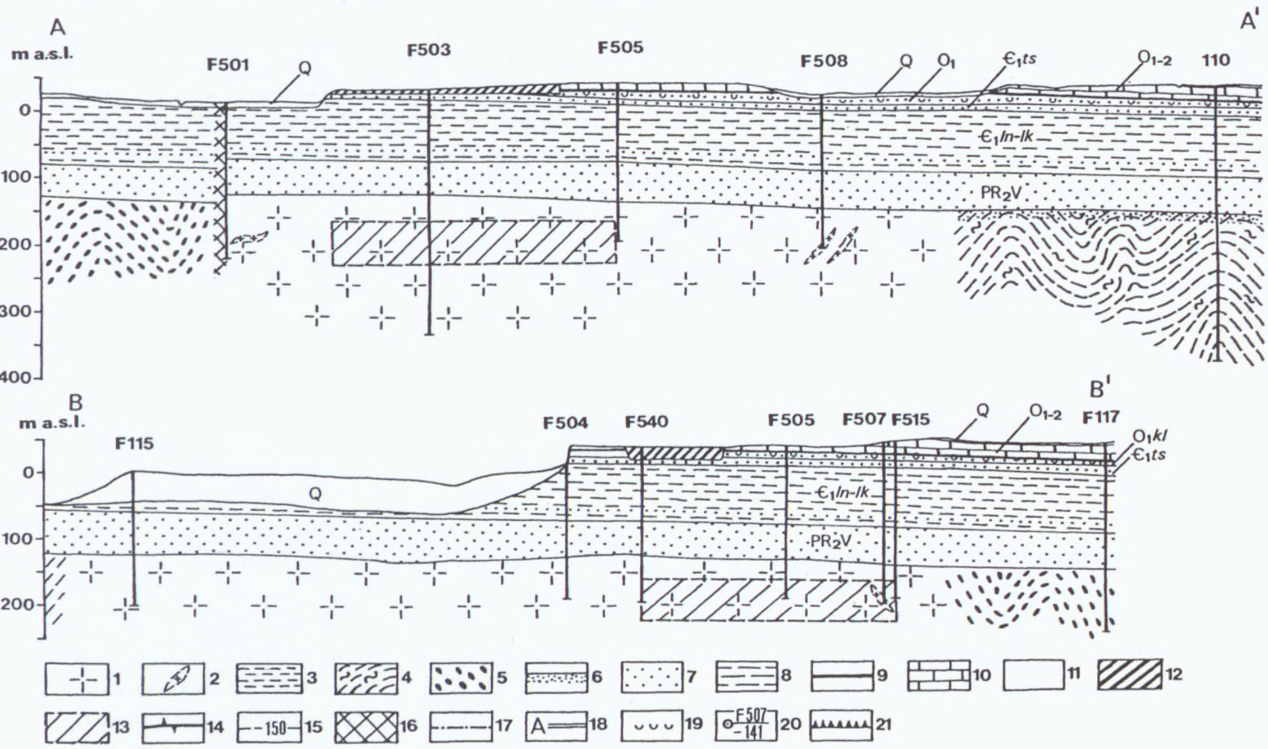


Fig. 230. Geological sections of the Maardu granite deposit: 1 - porphyry-potassium-granite; 2-5 - metamorphic rocks: 2 - amphibolite- and biotite-amphibolegneiss; 3 - quartz-feldspar and granite gneiss; 4 - sillimanite-, cordierite- and garnet-bearing biotitegneiss; 5 - biotite- and amphibole-biotitegneiss; 6 - weathered crystalline rocks; 7-10 - sedimentary rocks: 7 - sandstone and siltstone; 8 - clay; 9 - *Dictyonema* argillite; 10 - limestone; 11 - Quaternary sediments; 12 - waste rocks of the phosphorite quarry; 13 - area of calculated granite reserves; 14 - boundary of the Neeme intrusive; 15 - isohyps of the crystalline surface, m below sea level; 16 - zone of tectonic disturbances; 17 - boundary of the Maardu phosphorite quarry; 18 - line of geological section (Fig. 230); 19 - phosphorite layer; 20 - borehole, its number and the depth of the crystalline basement; 21 - escarpment. Geological indices: PR_2V - Vendian, $€_1ln$-lk - Lontova and Lükati formations, $€_1ts$- Tiskre Formation, O_1kl - Kallavere Formation, O_{1-2} - Lower and Middle Ordovician.

are distinct, of clearly cutting character. The veins of granite porphyry formed evidently during the final phase of the formation of the granite pluton. The granite porphyry occurs in 3 boreholes out of 36 and forms a bit more than 5% of the whole mass of the granites penetrated by boreholes.

Porphyry-potassium-granite is a pinkish-grey rock with a rather stable chemical and mineral composition and association of accessory minerals. The medium- to coarse-grained biotite-quartz-plagioclase groundmass comprises slaty phenocrysts of potassium-plagioclase (latticed perthitic microcline), 1-5 cm in diameter. The rock has a massive or slightly trachytoid (boreholes F-106 and F-115) texture. Below the crust of weathering and outside the fracture zones the rock is fresh, monolithic, with single partition joints (average one joint per 10 m), characteristic of granites. The mineral composition of the porphyry-potassium-granite is the following: quartz 20-30 %, microcline 35-40 %, plagioclase 10-25 % and biotite 2-10 %. The rock also contains some amphibole (hornblende), muscovite, fluorite and apatite. The accessory and ore minerals are represented by titanomagnetite, magnetite, sphene, orthite, monazite and zircon. In the zones of weathering sericite, chlorite, illite, kaolinite, montmorillonite, epidote, leucoxene, calcite and ferric hydroxides are present as secondary minerals.

Granite porphyry is a fine-grained massive reddish-grey biotite-quartz-plagioclase vein rock with a porphyritic texture. The phenocrysts of idiomorphic slaty microcline and plagioclase have a diameter of 2-5 mm. The mineral composition of the rock is as follows: quartz 30-35 %, microcline 35-40%, plagioclase 20-25%, biotite 5-10% and muscovite 1%. The content of accessory, ore and secondary minerals is the same as in the porphyry-like potassium-granite.

The average thickness of the weathered part of the Neeme intrusive is 5 m, while in the areas of fracture zones it reaches 32 m (borehole F-501).

The relatively high (60-100 R/h) radiation level is probably due to the inclusions of accessory zircon, monazite and sphene in biotite. The average content of U in the granites is 5.6 ppm and that of Th 25-30 ppm. These contents are quite characteristic of the rapakivi-granites.

The physical-mechanical characteristics of granites (Table 61) and crushed aggregate are high. The average Aggregate Crushing Value of porphyry-potassium-granite and granite porphyry is 15.2% and 9.6%, respectively. The content of flaky particles does not exceed 7%, the average being 5%.

Table 61. The physical-mechanical characteristics of the granite of the Maardu deposit

Parameter	No of samples	Unit	Test results		
			Min	Max	Mean
Specific gravity	138	g/cm^3	2.62	2.72	2.68
Volume weight (density)	138	g/cm^3	2.58	2.70	2.65
Porosity	138	%	0.1	0.1	0.1
Uniaxial compressive strength:					
- air-dry rock	138	kg/cm^2	1400	2822	1956
- water-saturated rock	138	kg/cm^2	1204	2692	1756
- after 100 cycles of freezing and thawing	138	kg/cm^2	912	2312	1565
- after 150 cycles of freezing and thawing	68	kg/cm^2	112	2060	1513

The reserves (cat. T) of granite of the Maardu deposit, calculated in an area of 11.4 km^2 (Fig. 229) at the thickness of the commercial bed 65 metres, are 741 million cubic metres.

Considering that in the course of mining at least 50 % of the rock remains in the safety pillars, the mineable reserves amount to 370 million cubic metres.

The possibilities of expanding the reserves are practically unrestricted. From a theoretical point of view, the only limiting factor is the geological boundary of the granite massif.

Due to the great thickness of overburden (123 m) and hydrogeological conditions in the area, the only possible way to operate the deposit is by underground mining. The Maardu granite deposit is overlain by the Vendian aquifer system which is the main potable water source for Tallinn and the whole northern Estonia. The deposit can be taken into use only in the case it is quaranteed that the aquifer system would remain undisturbed. Technically, this is feasible but rather expensive. The worked out space has good prospects to be used for underground storages.

CARBONATE ROCKS

Introduction

Carbonate rocks of commercial interest are widespread in Estonia. The continuous outcrop belt of the Ordovician and Silurian carbonate rocks embraces the territory northward from the Pärnu - Lake Peipsi line (Fig. 231). Owing to the rather thin Quaternary cover, these rocks have been accessible for use since 5000-4000 yr BC. The Devonian carbonate rocks, occurring on some levels only and cropping out in southeastern Estonia where the Quaternary sediments are rather thick, are available for use in very restricted areas.

Until the 13th century, flagstone was used in building tumuli, fences, wells, walls of strongholds, *etc.* without any binder (Einasto & Matve 1989). After the lime mortar was taken into use the utilization of various kinds of carbonate rocks increased remarkably and they became the main construction material in medieval castle and town building (Foto 67). Quarries were laid out in the vicinity of buildings and, in all likelihood, their total number exceeded one thousand. Flagstone from coastal quarries and stone blocks for sculptures and decorative details from inland quarries were exported to Russia, Germany and Scandinavia (Jürgenson 1960). Up to the end of the 19th century, the use of carbonate rocks was based only on the skill and experience of craftsmen who had achieved the highest level in using diverse rock types for certain purposes in building and lime burning (Hupel 1777,1782,1789, Fischer 1778). From the end of the 19th century, in accordance with the development of industry, carbonate rocks were widely used beside building and stone masonry also in the cement, metallurgic, sugar, paper and glass industries both in Estonia and abroad. Limestone and dolomite were exported mainly to Russia, Germany, Sweden, Norway, Finland and Latvia (Winkler 1922, Orviku 1933a,

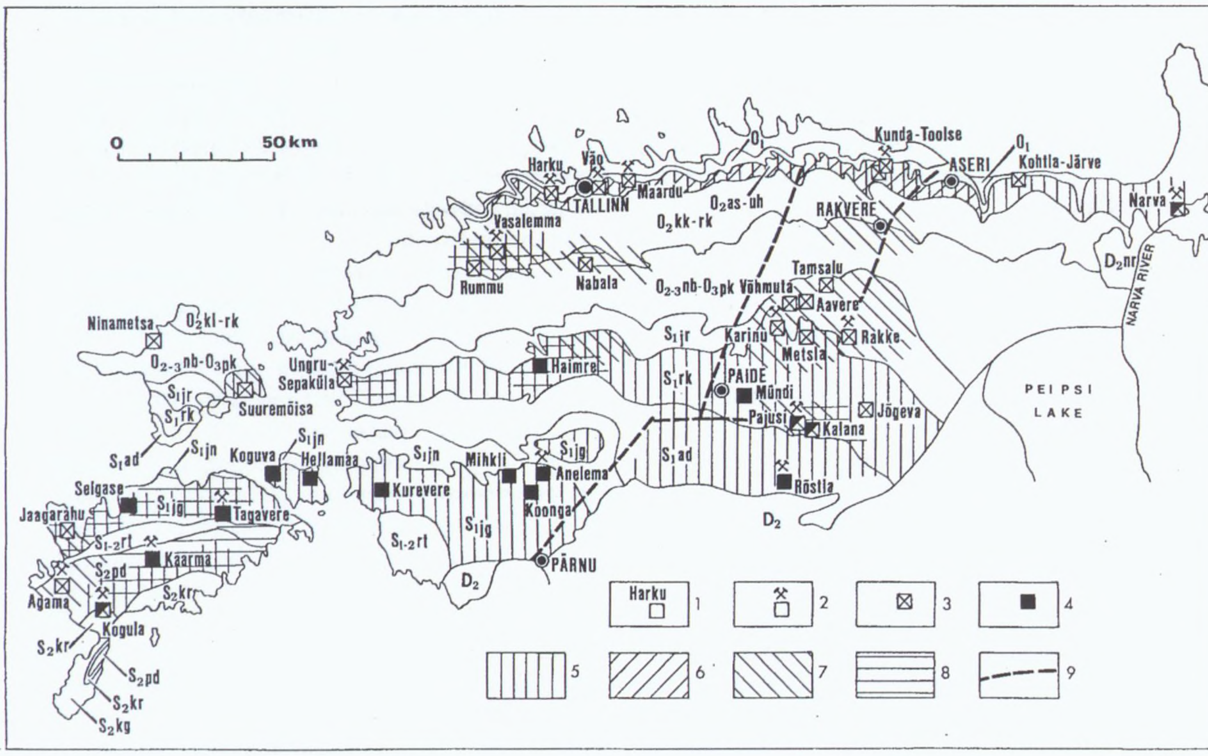


Fig. 231. General distribution of limestone and dolomite resources in Estonia: 1- deposit and its name; 2 - deposit in exploitation; 3 - limestone deposit; 4 - dolomite deposit. Promising areas: 5 - for construction purposes; 6 - for cement production; 7 - for lime production; 8 - for facing and ornamental purposes; 9 - zones of disturbances. Stratigraphical indices of regional stages: O_1 - Pakerort, Varangu, Volkhov, Kunda; O_2 as-uh - Aseri, Lasnamäe, Uhaku; O_2 kk-vr - Kukruse, Idavere, Jõhvi, Keila, Oandu, Rakvere; O_{2-3} nb - O_3pk - Nabala, Vormsi, Pirgu, Porkuni; S_1 jr - Juuru; S_1rk - Raikküla; S_1ad - Adavere; S_1 jn - Jaani; S_1 jg - Jaagarahu; S_{1-2} rt - Rootsiküla; S_2 pd - Paadla; S_2kr - Kuressaare; S_2 kg-oh - Kaugatuma, Ohesaare; D_2 - Middle Devonian.

Luha 1946). In the past, limestone from coastal quarries was loaded directly into ships, but today coastal movement is confined to relatively small quantities of technological limestone from the Vasalemma quarry and cement from the Kunda Works. The traditional fields of use have remained the same up to the present, but the proportions have changed. During the last 20 years, the exploitation of carbonate rocks has been devoted primarily to the production of aggregate, with less than 30% used for other purposes. The amount used as building and facing stone has declined the most.

The testings of properties began earlier (Kupffer 1870), but systematic studies and exploration of limestone and dolomite deposits were initiated not until the end of World War II (Teedumäe 1993). The total amount of mineable resources, explored on the ground of the former USSR instructions, are sufficient to meet all forseeable needs for another 50 or even more years. However, the assessment of the availability of the resources for current exploitation poses serious problems.

Distribution and characteristics

Workable deposits occur at many stages in Estonia and the country is well endowed with limestone resources (Кийпли 1984, Тээдумяэ 1983, 1986, 1988б). However, because of the thickness of the overburden, quality and consistency of the material and the convenient location of the outcrops, more than 75% of exploitation has been concentrated on Ordovician limestones (Lasnamägi, Photo 68; Uhaku, Photo 69;

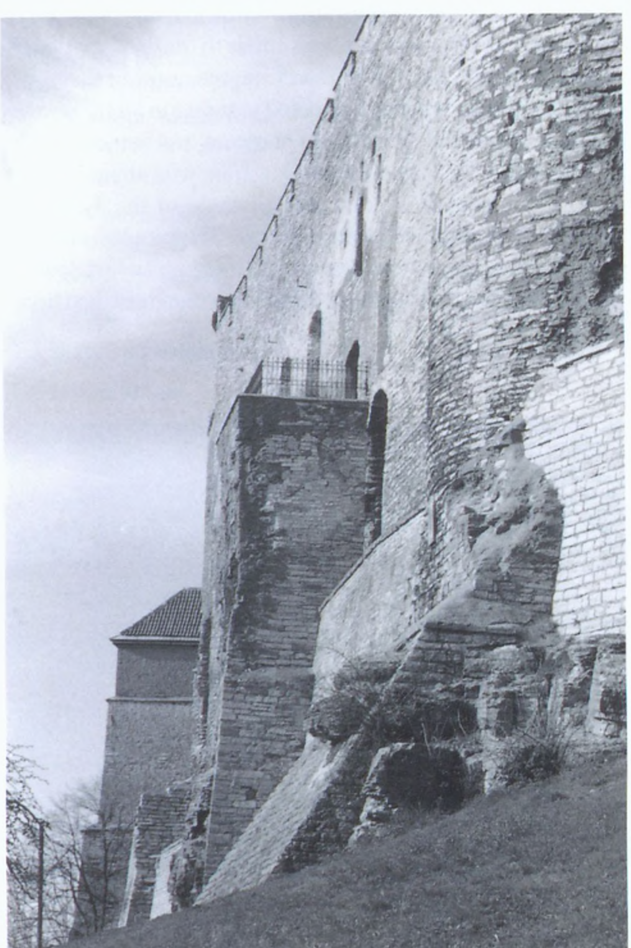


Photo 67. Toompea Castle (13th-14th centuries) in Tallinn, built from the rocks of the Lasnamägi Stage. *Photo by J. Nõlvak.*

Oandu and Keila stages). Other levels of industrial importance include limestones of the Lower Silurian Juuru Stage and dolomites of the Raikküla and Jaagarahu stages (Fig. 231).

In contrast to the rest of the territory, southeastern Estonia possesses few limestone and dolomite resourses. These occur only on some levels of the Devonian, predominantly terrigenous rocks.

In the following, the resources of economic or potential economic importance are described in order of geological age by stratigraphical units - stages, which correspond to the location of their approximately latitudinal belts of north-south trending outcrops (Fig. 231).

The **Ordovician** is represented mainly by limestones and marls. In eastern Estonia limestones are dolomitized. Glauconite-containing limestone and dolomite with intercalations of argillaceous limestone and marl of the Volkhov and Kunda stages have prospects for use as building material only in complex with overlying rocks in deep limestone quarries or in case of opencast mining of phosphorite deposits (Кивимяги и Тээдумяэ 1971, Пуура 1987, Teedumäe 1990).

Goethitic ooids-containing limestones of the **Aseri Stage**, crop out on the klint and in a narrow belt south of the klint. West of Tallinn, the high content of clayey and sandy components eliminates the use of the rocks. In the area between Tallinn and the Aseri tectonic disturbance zone, the composition of limestones is quite stable (Table 62) and in the past they were widely used for building purposes and in the cement industry. Due to the low (*ca.* 1.0) silica ratio, the additional silica-rich constituents are required to meet the specifications for cement. Eastward the dolomitization has limited the use of rocks to the production of building stone and aggregate. At present, the rocks make the bottom of the commercial bed in the Harku, Väo, Kunda and Narva deposits (Fig. 231).

Limestones, dolomitized limestones and dolomites of the **Lasnamägi Stage** make up a 7–10-m-thick complex, which crops out in an area, up to 10 km wide, and runs through the most densely populated and urbanized northern part of Estonia. This has been the essential region of active workings within the last millennium. Limestone has been used for building stone and aggregate, lime burning and stone masonry, afterwards also for cement production. The rocks of the Lasnamägi Stage are the most intensively quarried carbonate rocks in Estonia and, without doubt, one of the most important resources in the country. Today, the major operating quarries producing carbonate rocks for aggregate (Harku, Väo, Maardu) and cement (Narva) are of the main commercial significance in Estonia.

In the western part of the outcrop, three lithologically different complexes of rocks are distinguished (Orviku 1940, Кийпли 1983). In ascending order these are: thin- to medium-bedded argillaceous limestone with intercalations of marl (2.5-3.0m); dolomite (0.5m); medium- to thick-bedded limestone with clayey partings (3-7m). To the east of Kunda, the proportion of dolomite in the section increases rapidly and the easternmost sections are represented by dolomite.

The physical-mechanical properties of rocks in the whole distribution area (Table 62) meet the specifications for good quality building materials in block form and also for aggregate. The chemical composition is highly variable (Table 62), but the commonly high content of insoluble components confines the use of rocks to the field of cement production and agriculture. The limestone in the whole outcrop area west of Kunda may serve for cement in mixture with the Cambrian clay (Тээдумяэ 1983, Teedumäe 1990).

The main part of the succession of the **Uhaku Stage**, which includes highly argillaceous limestone containing interlayers of kukersite and marl is of no economic significance. On the outcrop west of Kunda (Fig. 231), less argillaceous limestone of the Väo Formation with insoluble residue 7-10% are spread in the basal part of the stage (Photo 69). The physical-mechanical properties are similar to those of the underlying Lasnamägi Stage and, therefore, the rocks of these two stages have been actively extracted together in all the abovementioned quarries for building purposes and cement.

Generally, the argillaceous limestone with interlayers of marl, kerogenous limestone and oil shale of the **Kukruse Stage** is of no commercial importance. The problems related with its use are acute because of the inevitability of dealing with several hundred millions tonnes of coarse crushed lime-

Photo 68. Southern part of the Lasnamägi quarry, which has provided main building stone for the Tallinn area. *Photo by A. Rõõmusoks.*

Photo 69. An excellent building stone of the Väo Formation from the Lüganuse quarry. *Photo by A. Rõõmusoks.*

Table 62. Properties of carbonate rocks of the main industrial levels (generalized data of geological explorations)

Stage, rock type	Content,%			Water absorption, %	Compressive strength,MPa	Aggregate Crushing Value,%	Los Angeles Abrasion Value,%
	CaO	MgO	i.r.*				
Aseri Stage							
Oolitic limestone	46-47	1.5-2.5	7-9	1-2	50-70	-	32-28
Dolomitized oolitic limestone	42-44	4-6	10-12	2-3	60-80	-	35-40
Lasnamägi Stage							
Argillaceous thin-bedded limestone	43-45	2-3	10-12	1-2	80-90	11-16	33-38
Dolomitized limestone	34-41	8-18	7-10	2-5	80-100	12-16	30-35
Medium- to thick-bedded limestone	44-48	2-3	6-8	1-2	90-170	10-12	20-25
Keila Stage							
Argillaceous limestone	41-46	1-4	13-16	2-3	80-100	13-24	-
Nodular limestone	48-50	1-4	6-9	1-2	90-120	15-20	26-28
Crystalline bioclastic limestone - "the marble of Vasalemma"	50-54	0.5-1.5	1-3	0.5-1	60-80	15-25	35-40
Aphanitic (micritic) limestone	48-51	1.5-2	4-5	0.5-1.5	140-170	12-18	
Nabala Stage							
Microcrystalline and aphanitic limestone	41-54	1-5	1-10	-	40-120	12-18	20-25
Juuru Stage							
Nodular limestone	46-49	1-5	6-8	1-2	60-100	15-17	20-30
Coquinoid limestone (*Borealis* limestone)	48-52	2-6	1.5-2.5	0.7-1.5	50-70	17-20	25-35
Stromatoporoid limestone	50-53	1.5-3	1.5-3	0.5-1	100-120	-	-
Raikküla Stage							
Micro- and cryptocrystalline limestone	48-50	0.5-1.5	5-6	1-2	100-120	10-15	30-40
Bioclastic limestone	48-53	0.5-1.5	2-5	1-2	90-120	14-15	30-35
Dolomite	27-30	18-21	2-5	2-3	120-130	12-15	35-45
Argillaceous dolomite	28-30	14-18	10-20	3-6	70-90	20-25	40-50
Adavere Stage							
Fine crystalline slightly argillaceous dolomite	27-30	17-21	5-10	1.5-3	100-120	13-15	20-30
Jaagarahu Stage							
Argillaceous dolomite	25-28	17-20	5-15	3-4	100-120	11-14	30-40
Micro- and finecrystalline bedded dolomite and massive riff dolomite	28-30	19-21	1-5	2-3	130-200	8-10	25-30
"Ornamental" dolomite	27-29	19-20	5-8	5-9	60-80	-	35-45
Bioclastic and biohermal limestone	50-53	1-3	2-4	2-3	110-160	15-25	25-30
Paadla Stage							
Stromatopoid-coral limestone	40-55	1-8	2-8	3-4	50-60	-	-
Microlaminated dolomite (Kaarma type)	28-34	16-20	10-18	6-10	15-40	24-29	-

* insoluble residue

stone removed to waste dumps from oil shale preparation plants. The material could be used for fillings, after secondary crushing for low quality aggregate purposes only because of the low (in cases less than 25 cycles) value of frost resistance.

In most of the outcrop of the **Keila Stage**, extending from the northern tip of Hiiumaa Island across mainland Estonia (Fig. 231), the section consists of interbedding argillaceous limestones and marls (Kahula Formation) of no industrial importance. In the western third of the mainland part, they are overlain by a complex of lithologically variable high-quality limestones which have been intensively used.

The lowermost part of the complex is represented by argillaceous limestone with thin intercalations of marl of the Kurtna Member. Both this and the above-lying slightly argillaceous nodular or seminodular limestone with thin wavy intercalations of marl of the Pääsküla Member have properties making them suitable for use for construction purposes (Table 62).

The Vasalemma Formation is distributed in the top of the section in the widest, central part of the outcrop (Fig. 231); its upper part extends into the Oandu Stage (Мяннииль 1960). This, up-to-15-metre-thick layer of irregular composition is spread discontinuously, except the surroundings of Vasalemma - Rummu. The main composing rock type, known as the "marble of Vasalemma", is a light-grey, bedded limestone, rich in skeletal debris. It contains irregularly shaped reef-like mounds of greenish-grey aphanitic (micritic), mainly massive limestone (Fig. 232). The size of the mounds is highly variable. Sporadical inclusions, lenses and intercalations of darker, greenish-grey highly argillaceous limestone and marl are characteristic of the whole section. The exploration of the Vasalemma deposit has shown that the crystalline limestone forms 62%, aphanitic limestone 31% and highly argillaceous limestone with marl 7% of the extent of the formation. The crystalline limestone is easily distingushable from the aphanitic limestone and can be selectively removed during quarring operations.The mining experience has shown that the soft,

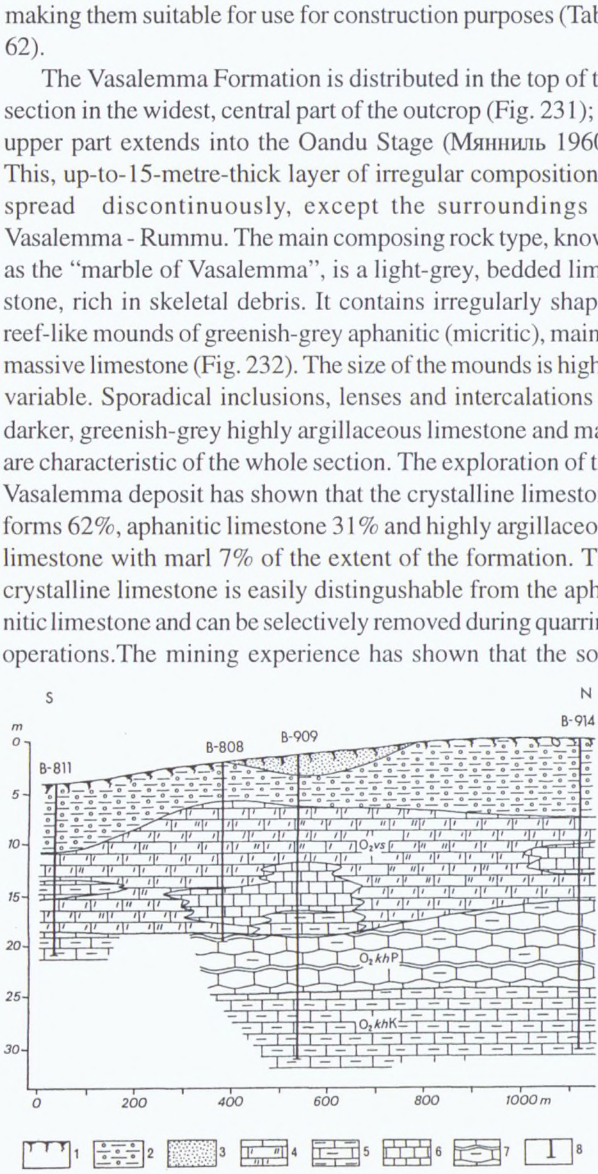


Fig. 232. Geological section of the Vasalemma deposit (after the exploration report by I. Barankina): 1 - soil; 2 - till; 3 - loamy sand; 4 - fine to coarse crystalline bioclastic limestone - "marble of Vasalemma"; 5 - argillaceous limestone; 6 - aphanitic (micritic) limestone; 7 - seminodular limestone with intercalations of marl; 8 - number of borehole. Stratigraphical indices: O_2 vs - Vasalemma Formation, O_2 khP - Pääsküla Member, O_2khK - Kurtna Member.

highly argillaceous limestone would fall off during preliminary treatment (blasting, loading, crushing) and would not affect the quality of the final product.

The physical characteristics (Тээдумяэ 1988в) and outstanding chemical purity (Table 62) account for the extensive use of limestone since the 13th century. Limestone was mined in several quarries and evaluated as a weatherproof good-looking facing stone both in natural and polished fashion, perfect material for stone masonry and building purposes. In accordance with the development of the technology, limestone found use in the pulp and paper, metallurgical and lime industries, in water treatment and purification of fuel gases in both Estonia and abroad. During the course of the Soviet period, most of the pure limestone produced from the Vasalemma and Rummu quarries was used for low-quality aggregate and only about a quarter found the high grade end use in the lime and paper industries. This high-rate, about million tonnes annually, wasteful exploitation has caused remarkable harm to the restricted resources of this valuable high-purity limestone.

The **Nabala Stage** crops out in a wide area extending through the whole of Estonia, but the natural exposures and old quarries are largely confined to a relatively small areas with a thin Quaternary cover on islands and on the mainland south of Tallinn, near the Town of Tapa in central Estonia and Tudulinna in eastern Estonia (Fig. 231). The section of thin-bedded, mainly aphanitic limestone with patterns of fine crystalline pyrite and intercalations of marl contains some thicker layers of dolomitized limestone which in the past was used as a building and decorative stone (Штакеншнейдер 1852), even in the construction of the Ermitage in St. Petersburg. There were no operating industrial quarries. Local people used limestone mainly for limeburning and some layers were recommended (Eichwald 1854) as lithographical stone. The chemical composition of limestones is variable (Table 62). The limestone of the most constant composition with the content of CaO above 50% is located south-west of Tallinn (Fig. 231, Тээдумяэ 1983). The main bulk of thin-bedded (less than 5cm) limestone may serve only for low-quality aggregate and for neutralising acid soils.

On its outcrop the **Silurian** is represented by carbonate and terrigenous carbonate rocks of a great variety (Юргенсон 1988). The primary dolomites are of a restricted distribution, but secondary dolomitization is widespread.

The continuous meridional outcop of the **Juuru Stage**, extending through the central part of Hiiumaa Island and the mainland, turns towards the Pandivere Upland in the south-east and becomes buried under the Devonian cover (Fig. 231). The lower part of the stage - the Varbola Formation, consists of nodular limestone and dolomitic limestone with a variable content of fragments of fossils and argillaceous material. Wavy intercalations of marl occur throughout the section, but their amount grows with depth. Due to the restricted exposures, the limestone has not been intensively used. The quality characteristics (Table 62) meet the requirements of low-quality aggregate. Recently, the problems related to its use, have become topical because of the exhaustion of the resources of covering chemically pure limestone on the Tamsalu and Rakke deposits. The limestone, exposed on the bottom of large quarries, is available for exploitation, whenever needed, without any restriction.

Limestones of the Tamsalu Formation in the upper part of the stage are the main source of raw material for the lime industry in Estonia, founded in 1914 at Tamsalu. White massive coquinoid limestone (Tammiku Member), composed of the valves of the brachiopod *Borealis borealis*, forms the famous *Borealis* Bank which is spread at the whole length of the outcrop and reaches its maximum thickness of 13 metres on the Pandivere Upland (Нестор 1970). Eastward, both the thickness of the bank and the content and size of skeletal particles decrease, while the content of argillaceous material and the number of marl intercalations increase. The composition of the limestone excels of chemical purity (Table 62), the higher contents of magnesia are confined to the irregularely dolomitized disturbance zones.

The discontinuosly developed fine-crystalline up to aphanitic limestones rich in fragments of stromatoporoids and corals (Karinu Member) cover the commercial bed of the *Borealis* limestone on main deposits (Fig. 231) and have the properties acceptable for complex use (Table 62). The limestone has mainly been used for lime production, in the pulp and paper industry, as aggregate for construction purposes and in agriculture.

The limestone area between Rakke, Tamsalu and Väike-Maarja is of great importance in terms of the development of lime industry. It accommodates four well-explored deposits (Fig. 232, Table 63) and several preliminarily studied sites forming a considerable reserve base.

The wide outcrop belt of the **Raikküla Stage** starts from the southern part of Hiiumaa Island and runs through central Estonia. The Quaternary cover is less than 5 metres thick on most of the outcrop. There are numerous small old quarries of which the best-known ones, like Pusku (Ungru -Sepaküla), Orgita (Haimre), Kalana a.o., belong today to the areas of explored deposits (Fig. 232).

The section consists of a great variety of alternating limestones, dolomitized limestones and dolomites, in places with high-quality beds. The content of terrigenous material is higher in the lower and middle parts of the section (Юргенсон 1970). The rocks of the Raikküla Stage have not found intensive and high grade industrial use, probably because of the difficulties in maintaining a consistent product.

According to the composition of the upper, minable part of the sequence, four areas are distinguished (Fig. 231) with different prospects for use (Teedumäe 1992).

In the Western area, the whole section consists of micro- and cryptocrystalline and bioclastic limestones with persistant physical-mechanical properties and high-purity layers (Table 62) which are really prospective for construction purposes, including small decorative elements. In case of selective mining they may also serve for purposes where chemical purity is needed.

The Rapla area, extending from the surroundings of the Haimre deposit as far as the Town of Paide is characterised by a considerable variability of alternating limestones and dolomitized up to dolomite limestones (Table 62). The thickness of different rock types in section varies from 0.2 to 4 m. Both dolomite and limestone can be used for the production of construction materials: aggregate, dimension stone, decorative units. Dolomites, particularly microlaminated ones (Haimre deposit), are promising for sculptural purposes. The possible dimensions of the blocks are restricted because of the small thickness (mainly 20-30cm) of the beds and vertical fractures.

In the Paide area, which extends eastward as far as the Pajusi deposit (Fig. 231), the whole section is composed of dolomites, cavernous in some places. There are no prospects for sizable resources of high purity dolomite as interlayèrs, rich in the terrigenous component, are completely common. The dolomite has prospects for use in the industry of construction material only (Fig. 232, Table 63).

In the Eastern area (east of the Paide area), the section is variable in composition. It is resented by argillaceous dolomite and mainly aphanitic limestone, here and there dolomitic and with interlayers of limestone rich in skeletal component. Occasionally (Kalana deposit), the aphanitic limestone is 3-4m thick. The medium- up to coarse-crystalline high-purity (Table 62) limestone comprises numerous chalcedony nodules (Юргенсон 1958б). When polished, it attains a marble-like appearance, accounting for its popular name - the "marble of Kalana" (Einasto & Perens 1985). The limestone has been quite widely used for several purposes. The hand-mined rock has found application in the lime and glass production and also in small decorative and sculptural units. The restricted thickness and distribution of this unique rock, exploitable in complex with dolomite and aphanitic limestone, suitable (Tables 62, 63) for construction purposes only, make obvious difficulties in achieving high outputs.

The outcrops of the **Adavere Stage** extend as an almost west-east trending belt, widening abruptly east of the Paide - Pärnu zone of disturbances, from Hiiumaa Island through central Estonia (Fig. 231). The lithological composition of rocks is variable. Great changes are due to the depositional environment, *e.g.* eastwards the deep-water sediments are replaced by shelf sediments. Based on the lithological variability, several subdivisions of different rank, volume and stratigraphical position have been distinguished (Юргенсон 1966, Кальо 1970в, Эйнасио и др. 1972, Kaljo 1990b, Perens 1992).

In the western part of the outcrop limestones, except single interlayers, are highly argillaceous and may serve for local needs only (Teedumäe 1996a). The Mõhküla beds are prospective for industrial use. Lithologically, they consist of predominantly fine-crystalline secondary dolomite with intercalations of highly argillaceous dolomite and domerite. Caverns, chalcedony nodules, relicts (in places silicified) of skeletal debris of fossils and accumulations of pyrite are common in the whole section. In the well-known Võhma-Vaki area of sulphide mineralization (Пальмре 1960) the crystals of galenite and sphalerite as well as breccia dolomite occur.

The chemical composition (Table 62) is variable and related to the content of terrigenous component (Кийпли 1984, Teedumäe 1996a). The constantly high (more than 0.2%) content of Fe_2O_3 confines the prospects of the use of dolomites to the categories beyond the colourless glass. The use of dolomite in agriculture has not yet been practiced in Estonia, but in case of need the Mõhküla dolomite is available. In general, the physical-mechanical properties of dolomites (Table 231) meet the requirements for construction purposes. The relatively high porosity, caused by the presence of the closed cavities originating from dolomitization, does not af-

Table 63. Main deposits of carbonaceous rocks

Deposit	Active reserves as of 01.01.97, th. cu m		Rock type and main field of use
	Cat. T	Cat. R	L - limestone, D - dolomite
Aavere	2,995	292	L, lime, pulp and paper industry, building stone, agriculture
Agama (Mõisaküla)	42	-	L, lime production, agriculture
Anelema	2,440	2,650	D, building stone and aggregate
Harku	3,676	-	L, building stone and aggregate
Hellamaa	834	-	D, glass industry, agriculture
Kaarma	1,209	751	D, building and finishing monumental stone
Kalana	312	1,119	L, D, building and finishing stone, aggregate, lime production
Karinu	2,710	936	L, lime, pulp and paper industry, agriculture
Kogula	1,716	5,346	L, D, building aggregate
Koguva	5,552	-	D, building aggregate
Koonga	5,980	-	D, building stone and aggregate, agriculture
Kunda-Toolse	11,274 13,217	37,846 87,946	L, building stone and aggregate L, cement production
Kurevere	3,390	-	D, building aggregate, agriculture
Maardu	1,056	-	L, building stone and aggregate
Metsla	2,162	-	L, lime, pulp and paper industry, agriculture
Nabala	-	167,928	L, building aggregate
Narva	550	3,117	L, building stone and aggregate
Haimre	623	-	D, building, finishing, monumental stone and aggregate
Pajusi	3,837	-	L, D, building stone and aggregate
Rakke	2,081	-	L, building aggregate
Rummu	872 4,338	- -	L, lime, pulp and paper industry, agriculture, finishing and monumental stone L, finishing and monumental stone, aggregate
Rõstla	2,201	-	L, building aggregate
Selgase	716 775	2,902 287	D, finishing stone D, building aggregate
Suuremõisa	1,115	-	L, lime production, building stone and aggregate
Tagavere	1,020	1,163	D, finishing stone, building stone and aggregate
Ungru-Sepaküla	190	168	L, D, building stone and aggregate, small decorative details
Vasalemma (site No. 2)	7,663	-	L, lime, pulp and paper industry, agriculture, building and monumental stone, aggregate
Võhmuta	2,263	-	L, lime, pulp and paper industry, agriculture
Väo	17,961	-	L, building stone and aggregate, smaller decorative articles and finishing slabs

fect frost resistance as is confirmed by the low value of water absorption. The intercalations of highly argillaceous dolomite and domerite do not remarkably reduce the quality of aggregate as they would fall off during mining and processing operations. In case of need, the washing is used (Rõstla quarry). The silica nodules may affect the use of dolomite for concrete aggregate in places they are revealed. In practice, complications due to their adverse impact have not been recorded so far. The dolomite could be polished and used for sculptural details, but the commonly small thickness of beds, seldom more than 25cm, reduces the prospects to small local handycraft works.

The fairly extensive outcrop of the **Jaagarahu stage** extends through the northern part of Saaremaa Island and the southern part of Muhu Island as far as the southwest of mainland Estonia. The thickness of the covering Quaternary sediments is smaller in the northern part of the outcrop where numerous small old quarries and recently explored deposits and quarries in exploitation are located (Fig. 231). These limestones of shallow-water origin abound in coral stromatoporoid bioherms, in places mud mounds occur. The limestones are dolomitized in most of the area, except the western part of Saaremaa Island, and serve as the major source of dolomite in Estonia. The secondary dolomites, attributed to the limestones they originate from, have a relevant composition and distribution (Kiipli 1986). The purest dolomite is connected with reef facies (Figure 233). The spacial distribution and size of reefs is variable. The largest reefs with a diameter of some kilometres and thickness up to 16 metres are known from Saaremaa Island (Аалоэ 1970). Lithologically, the massive cavernous, in places breccia-like, dolomite and the surrounding medium- to thin-bedded fine-crystalline dolomite are highly different, but their chemical composition and even physical-mechanical properties are quite similar. These, the purest and most solid dolomites in Estonia (Table 62), have good prospects for use in agriculture, metallurgy and for construction purposes. The content of terrigenous component is gradually growing with depth. The thickness of high-purity dolomite is varying according to the rate of the post-Silurian denudation, but the whole northernmost part of the outcrop on the mainland and Muhu Island have good prospects for revealing sites ofwith a workable thickness of the chemically pure dolomite. The restrictions to exploitation are mainly connected with the land ownership, the problem being currently acute on Hellamaa, Anelema, Koonga and Kurevere deposits.

The limestones in the westernmost part of Saaremaa Island, represented by biohermal and surrounding aphanitic and fine-crystalline limestone with the fragments of fossils and intercalations of marl, are high in purity (Table 62). Before World War II, biohermal limestone was excavated and exported for the lime, sugar, metallurgical, *etc.* industries from the Jaagarahu quarry, exhausted by now. The prospects of revealing high-purity limestone resources are restricted to some small amounts.

The upper part of the Jaagarahu Stage consists of highly variable rocks. Due to the cyclic alternation of the rock types, characteristic of the section as a whole, they do not form considerable areas of equal qualities. The facing slabs produced in the Tagavere quarry have been used in constructing many outstanding public buildings (Photo 70). The Selgase quarry has been exploited periodically for the same purposes. The small (20-50 cm) thickness of beds and the presence of fractures determines both the size of the end product and the amount of waste. The latter can be used in producing aggregate for construction purposes.

The lithologically multivarious complex of interbedded limestone and dolomite of the **Paadla Stage** extends through the central part of Saaremaa Island. It has been extensively exploited in small quarries, mainly for local building purposes. There are two rock types suitable for wider use - the microlaminated argillaceous dolomite, distributed in the Kaarma deposit and biohermal stromatoporoid-coral limestone, distributed in the western area and forming the Agama and Atla deposits (Fig. 231, Tables 62, 63). The Kaarma dolomite (Эйнасто 1970) is thick-bedded (up to 1 m) weatherproof good-looking grey dolomite, soft when mined, but hardening in the air. It has been the main source for the industry of facing slabs and other decorative details in Estonia during the last 40 years.

The stromatoporoid-coral limestone excels of high chemical purity (Table 62). It is used for limeburning in restored ancient lime kilns at Lümanda (Agama deposit). The lime is also slaked in wood-coated trenches as it was done in the past. The high-quality product is used for the restoration of architectural monuments in Estonia and Finland.

The **Devonian** carbonaceous rocks are of restricted and discontinous distribution and of little economic importance. The limestone and dolomite of the Pļaviņas Stage outcropping in southern Estonia (Fig. 90) may be used for local needs of this remote area. There are three explored deposits and some prospective areas on the sites with a thin Quaternary cover having value for constructional purposes. Although high-quality limestone is present in places, its thicness is small (less

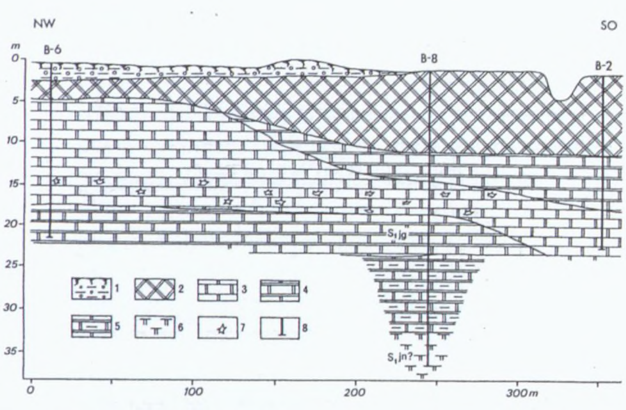


Fig. 233. Geological section of the Kurevere deposit (after the exploration report by S. Korbut): 1 - soil and till; 2 - reef dolomite, 3 - fine to microcrystalline horizontal-bedded dolomite; 4 - dolomite with intercalations of domerite; 5 - argillaceous dolomite with intercalations of domerite; 6 - domerite; 7 - caverns; 8 - number of borehole. Stratigraphical indices: S_1 jg - Jaagarahu Stage, S_1 jn - Jaani Stage.

Photo 70. The building of the National Library in Tallinn (arch. Raine Karp), covered with facing slabs from the Tagavere quarry. *Photo by R. Einasto.*

than 3 metres) and it does not form resources, sufficient for industrial use.

Resources and exploitation

The reserves (cat. T) of carbonate rocks in more than 80 explored deposits exceed 650 million cubic metres of limestone and 50 million cubic metres of dolomite. The amount of the preliminarily estimated (cat. R) reserves is almost as high. About 80% of the reserves was explored for use for construction purposes, 11% for the cement industry and 9% for the lime and pulp and paper industries. The last 20 years under the Soviet Union brought about the concentration, low specialization and extensive growth of the industry, the number of the deposits left in exploitation varied between 10 and 15, with the total annual output of about 4-5 million cubic metres (10 million tonnes). At the same time, the technological progress came to standstill, and in 1991 the majority of the industry was operating at the technological level of the 1970s. Today, fundamental renovation and reorientation to the demands of the new market and ownership relations is urgently needed. The process has just started and the new trends are in the state of formation. The first to be renovated is the cement plant at Kunda. According to the investment programme, the capacity of the plant - 900,000 tonnes will be achieved after 1998.

Today, the annual total output of carbonate rocks is about 12 million cubic metres (3 million tonnes). Of that, 670 thousand cubic metres is used for construction purposes, 350 thousand cubic metres for cement production, 80 thousand cubic metres for the lime industry, water treatment and desulphuration of fuel gases, glass manufacture, facing and ornamental stone. Beside the large industrial quarries (Table 63), small hand-mined quarries have been taken into use during the past years. The production is used to meet the local building needs, in cases, for ornamental stone.

Actually, the problems associated with the availability of deposits to exploitation and their use, having regard to competing demands of land use and other relevant factors, in every particular case need assessment in response to contemporary conditions. The process has started already and the first results have controverted the traditional opinion of unlimited resources of carbonate rocks. Resources available for mining are limited for many different reasons and, as a rule, the most favourable of them are located near towns, harbours and roads. It particularly concerns the industrially developed northern Estonia where the largest resources and consumers of construction materials are located. The problem is acute and waiting for a reasonable solution.

SAND AND GRAVEL

Introduction

Sand and gravel are the most common local building materials in Estonia. Sand is used in different building mixtures, as fine aggregate in concrete, ferroconcrete and asphalt concrete, in the manufacture of silica and silicalcite products, ceramics and glass, in road-making, *etc*. Sands with prevailing quartz and feldspar composition are most widespread (Раукас 1978).

Gravel is composed of magmatic, metamorphic and sedimentary rock particles, however, the variaties with carbonate composition dominate.

Natural sand and gravel consist of several components with different grain-size. Most common are gravel and sand mixtures which often contain pebbles and cobbles mixed with finer material, such as silt and clay. Several grain-size classifications are in use in Estonia (Table 64). The upper boundary of sand fraction varies with different fields of science, *e.g.* it is 1 mm in the decimal system, 2 mm in the soil science and geological mapping, 5 mm in mineral reserves exploration.

In Estonia, a sand or gravel deposit is registered as useful mineral if it meets the established requirements. In case of sand, the modulus of grain size must be at least 1.3, the content of clay and silt particles may not exceed 10% and the percentage of particles larger than 5 mm in diameter must be less than 35. In technological sand, which is a variety of sand acknowledged as useful mineral, the concentration of SiO_2 must be at least 95%, Fe_2O_3 less than 0.6% and Al_2O_3 below 2.5%.

The deposit of gravel is useful mineral when at least 35% of its particles are larger than 5 mm in diameter and its silt and clay content does not exceed 20%. Conventionally, cobbles, pebbles and boulders are also referred to gravel. The deposits of sand and gravel which do not meet the requirements prescribed to useful minerals, are considered as earth material.

Distribution and quality

The composition and properties of sand and gravel are determined by their genesis. Genetical aspects of sediment properties have been dealt with in several works (Кессел и Раукас 1967, Раукас 1978, Калм и Мешин 1983, Калм 1984, Калм и др. 1985).

In Estonia, sand and gravel deposits are mostly glaciofluvial (70-80%), to a lesser extent glaciolacustrine, aeolian and marine in origin. Glaciofluvial deposits occur in eskers, kames, fluvioglacial deltas and outwash plains over a wide area (Fig. 234).

The reserves of sand and gravel and the composition of sediments in eskers are changeable depending on the esker's shape and size (Раукас 1978). The narrow, steep-sloped eskers consist usually of coarse-grained sediments, while in large flat-roofed eskers sand and fine gravel prevail. In a long ridge of eskers, both the shape of eskers and composition of sediments may vary with the different parts of the ridge (Ряхни 1967). The concentration of coarse particles in eskers may reach 60-70%. This kind of material is dominated by carbonate rocks, the content of which may amount to 80-90%. The proportion of magmatic and metamorphic rocks does not usually exceed 20-30%. Generally, the sand fraction is dominated by quartz (more than 60%) and feldspars (up to 25%). Occasionally, the concentration of carbonates exceeds 60% (Раукас 1978).

The deposits of eskers are in natural form suitable for use

Table 64. Classifications of disintegrated deposits on the basis of granulometric composition

Grain size, mm	Decimal system	Geotechnics	Geological mapping	Exploration of mineral resources
10,000	gigantic boulders			
1000	boulders		boulders	boulders
200		boulders		
100	cobbles	cobbles	cobbles	cobbles &
70				pebbles
20		pebbles		
10	pebbles		pebbles	
5				gravel
2		gravel	gravel	
1	gravel			
0.1	sand			
0.05		sand	sand	sand
0.01	silt		silt	silt
<0.002	pelite	silt	clay	clay
<0.001		clay		

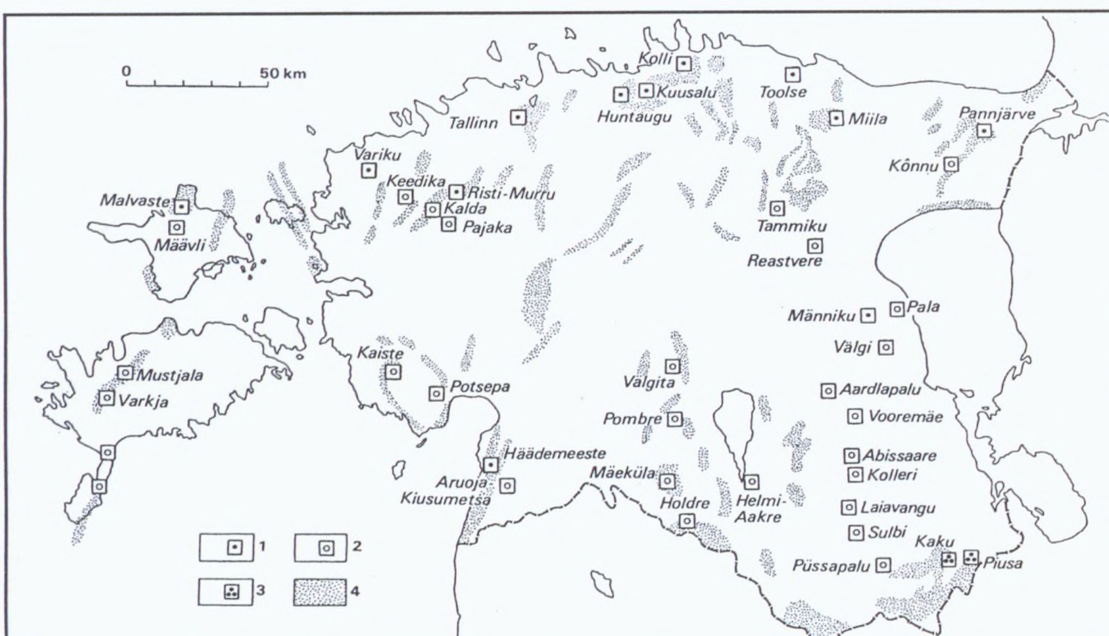


Fig. 234. The most important sand and gravel deposits: 1 - sand; 2 - gravel; 3 - quartzose sandstone; 4 - main perspective areas of sand and gravel (after Karukäpp 1988).

in road construction only. To get material for the manufacture of concrete and other building material, selective mining, crushing and screening are usually needed. The largest deposits of esker sediments are situated at Kaopalu, Keedika Kalda and Pombre (Fig. 234).

In kames, glaciofluvial sediments are more fine grained, uniform and with clearly observable bedding. Sand with layers or lenses of gravel or sand and gravel mixture prevails. Small lenses of silt and clay occur throughout the section, but they are thin and do not affect the properties of the extracted material. Kame sand is richer in quartz (more than 70%) and poorer in feldspars (below 20%) than esker sand. For instance, in the sand excavated on the Pannjärv deposit, the content of quartz exceeds 90%, making it suitable for use in concrete and mortar, but also in the manufacture of silica and silicalcite products. Pannjärve, Miila, Välgi, Välgita, Tääksi, Püssapalu, Vooremäe and Kukemetsa are the major deposits situated in kame sediments (Fig. 234).

The sediments of glaciofluvial deltas and outwash plains consist mostly of sand which is well sorted and relatively poor in silt and clay particles. In the proximal part, the deposits are coarser (often pebble and gravel with cobbles) than in the distal part. Occasionally, lense-shaped layers of gravel are encountered. In the mineral composition quartz (80-95%)

prevails, followed by feldspars. Sands may find use for different building purposes – from road construction up to the manufacture of concrete, silica, silicalcite and ceramics products, in mortar and plaster. Tallinn, Huntaugu, Kuusalu and Kolli are the largest sand deposits (Fig. 234).

Glaciolacustrine sands are generally fine, often clayey and of little promise. Coarse sand suitable for building purposes is of limited distribution and occurs only within beach ridges, where sands alterate with gravel and pebbles.

Aeolian sand forms dunes in northwestern, southwestern and northeastern Estonia, and also in the northern part of Lake Peipsi Depression and in the Võrtsjärv Depression. The largest dune fields relate to the coastal zones of the transgressive phases of the Ancylus Lake and Litorina Sea (Raukas 1968). The sediments consist of fine sand with uniform grain-size, which has been used mainly for filling. Owing to its high SiO_2 (70-92%) content, the sand has been successfully applied to producing cell-silicate construction units, recognized as a good heat-insulating, noise-damping and frost-resistant building material.

The sand and gravel deposits within and under till are not large, but in areas, where better material is lacking, they are in use as a mineral resource. Excavation of sand lenses is common in drumlins, but also in southern Estonia where glaciofluvial sediments form the core of hills, covered by till. This kind of sand and gravel have the content of clay and silt particles usually in excess of the permitted limit, and do not always meet the requirements of useful mineral. Mining is sometimes restricted by the great thickness of the overlying till.

The coastal sediments of the ancient stages of the Baltic Sea are an important source of sand and gravel in western Estonia (Valgevälja, Mustu-Nõmme) and on the islands of the West-Estonian Archipelago (Määvli, Tehumardi, Kõpu). Coastal formations contain often coarse, well-sorted gravel but its resources are limited.

Estonia' bedrock is dissected by deep valleys and valley-like troughs, filled with glacial and post-glacial sediments (Таваст и Раукас 1982) which comprise great quantities of sand and gravel. In terms of the composition and quality, fluvial sediments in buried valleys are similar to glaciolacustrine delta sediments. The suitability of fluvial sediments of buried valleys for building purposes has little been studied so far. In this respect, the Vasavere ancient valley within the Kurtna Kame Field, where the thickness of the Quaternary cover reaches 80 m, has been studied in more detail.

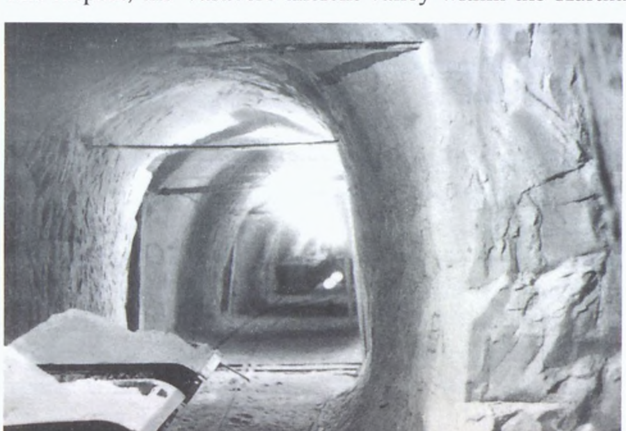


Photo 71. The technological sand deposit at Piusa is related to the Gauja Formation. *Photo by I. Kala.*

Sand and gravel occur also in contemporary coastal formations, on the sea floor and on the bottom of large lakes. To take these reserves into use, large-scale research into coastal dynamics is needed. Excavation of sand in coastal areas is complicated, because beaches are important recreation areas. In the Piirissaare delta, where dredging is needed to keep the fairway open, the removed sand has been used since long at the Construction Material Plant at Tartu.

Beside Quaternary sediments, the Devonian weakly cemented sandstones have also been used in the construction industry and in the manufacture of lower-quality glass (Orviku 1933a). Continuity of glass production has been maintained by the Järvakandi Works which have operated since 1878 and used the sand from the Piusa deposit since 1936.

Piusa and Kaku, the only technological sand deposits of Estonia, are related to the Gauja Formation (Fig. 234). In natural state the fine and very fine quartzose sandstone of the Piusa deposit (Photo 71) is suitable for the manufacture of flat and contained glasses (except colourless) and for foundry patterns (Table 65). To reach the standards of colourless glass, the enrichment of sand is needed, which so far has not been practiced on a large scale.

The quartzose sandstone of the Kaku deposit is finer and richer in iron compounds (Fe_2O_3 content 0.4-0.8%) and suitable for producing coloured container glass and foundry patterns.

The weakly cemented fine to very fine Lower-Cambrian sandstone of the Tiskre Formation outcropping in the lower part of the North-Estonian Klint has a composition (Table 66) which makes it suitable for use as technological sand. In Estonia, it has not been used so far, but in Russia it has been worked for producing glass since long.

Table 65. Chemical composition of the sandstone of the Piusa deposit (after Korbut, manuscript report 1991)

Part of the deposit	Content of components, %					
	SiO_2	Al_2O_3	Fe_2O_3	CaO	MgO	CO_2
Piusa, right bank	96.92-	0.61-	0.13-	0.20-	0.07-	0.21-
(Tuhkavitsa)	98.58	1.66	0.27	0.23	0.12	0.37
Piusa,	97.04-	0.21-	0.09-	0.07-	0.05-	0.30-
left bank	97.83	0.35	3.74	0.42	0.10	0.31

Table 66. Chemical composition of the sandstone of the Tiskre Formation (Чехомский и Пальмре **1960**)

Location	Content of components				
	SiO_2	Al_2O_3	Fe_2O_3	TiO_2	CaO
Saka	96.81-	0.53-	0.05-	0.15-	-
	99.40	1.69	0.09	0.22	0.36
Kunda	96.69-	0.58-	0.13-	0.46	
	98.80	0.81	0.20		
Aseri	98.08	0.36	0.18	-	0.41

Reserves and use

The resources of sand and gravel are of uneven distribution (Fig. 234, Table 67). About 70% of explored reserves belong to sand, half of those occur in the four largest deposits – Pannjärve, Vooremäe, Kuusalu and Tallinn. Currently, sand is found almost everywhere in Estonia, but the reserves of gravel have already been exhausted in several counties.

Sand and gravel form scenic landforms and landscapes which have been taken under protection in wide areas. This complicates estimation of the resources and exploitation of

Table 67. Main sand and gravel deposits in Estonia

County, deposit	Material (s=sand g=gravel)	Active reserves, million m^3		Genetic type
		Proved	Possible	
Hiiumaa				
Malvaste	s	0.2	7.5	glaciofluvial, Litorina Sea
Määvli (Kapastu)	s&g	6.5	13.0	coastal, Litorina Sea
Läänemaa				
Kalda	g	0.8	-	glaciofluvial
Keediku	g	1.3	-	glaciofluvial (esker)
Variku	s	3.6	3.1	aeolian (dune)
Risti-Murru	s	-	2.5	glaciofluvial (kame)
Harjumaa				
Tallinn	s	5.5	1.9	glaciofluvial (delta)
Huntaugu	s	7.4	3.8	glaciofluvial (delta)
Kuusalu	s	18.4	13.3	glaciofluvial (delta)
Kolli (Sillaotsa)	s	0.4	14.7	glaciofluvial (delta)
Lääne-Virumaa				
Toolse	s	0.9	-	glaciolacustrine (limnokame)
Miila	s	0.1	8.3	glaciolacustrine (limnokame)
Tammiku	g	8.0	-	glaciolacustrine (limnokame)
Ida-Virumaa				
Pannjärve	s	71.1	9.1	glaciofluvial (kame)
Kõnnu	g	-	42.6	glaciofluvial (esker)
Saaremaa				
Varkja	s&g	0.1	5.2	glacial + aeolian
Mustjala	s&g	-	3.0	coastal, Ancylus Lake
Tehumardi	s&g	1.9	2.9	coastal, Limnea Sea
Viieristi	s&g	-	3.4	coastal, Litorina Sea
Pärnumaa				
Potsepa	s&g	3.7	3.6	glaciofluvial (delta)
Aruoja- Kiusumetsa	s&g	-	18.1	glaciolacustrine (kame)
Häädemeeste				
Kaiste	s	0.4	-	aeolian (dune)
Raplamaa	g	2.1	-	glaciofluvial (esker)
Pajaka				
Jõgevamaa	g	1.6	8	glaciofluvial (kame)
Reastvere				
Pala	s&g	-	4.3	glaciofluvial (kame)
Männiku	g	-	3.8	glaciofluvial (kame)
Viljandimaa	s	-	9.0	glaciofluvial (kame)
Mäeküla				
Välgita	s&g	2.7	27.7	glaciofluvial (kame)
Pombre	s&g	0.2	-	glaciofluvial (kame)
Valgamaa	s&g	0.4	-	glaciofluvial (esker)
Helmi-Aakre				
Holdre	s&g	1.5	39.6	glaciofluvial (kame)
Tartumaa	s&g	4.0	1.5	glaciofluvial (kame)
Välgi (Selgise)				
Vooremäe	g	0.4	4.2	glaciofl. + glaciolac. (kame)
Aardlapalu	s&g	20.0	1.2	glaciofluvial (kame)
Põlvamaa	s&g	0.5	1.0	alluvial
Kolleri				
Abissaare	g	0.5	1.1	glaciofluvial (kame)
Piusa	g	4.2	-	glaciofluvial (kame)
Laiavangu	s	2.9	-	Devonian
Võrumaa	g	-	11.8	glaciofluvial (kame)
Sulbi				
Püssapalu	s&g	8.4	-	glaciofluvial (kame)
Kaku	g	1.0	7.1	glaciofluvial (kame)
	s	1.6	9.9	Devonian

the explored deposits. Therefore, the data presented in Table 67 do not necessarily reflect the possibilities of the actual use of the explored reserves.

The use of sand and gravel has always been controlled by the intensity of construction. The amount of the sand and gravel used was at its highest in 1975-90 (Raudsep & Räägel 1993) when 12 -15 million m³ was excavated annually from 900 deposits. Of the total used, 50% was gravel. More than 75% of gravel was used in road-making.

In the restored independent Republic of Estonia, the need for large-scale construction activities decreased. This was reflected in exploitation of mineral resources for construction purposes. In 1990, only 0.9 million m³ of sand and 0.6 million m³ of gravel were used (Raudsep & Räägel 1993). These quantities were much the same during 1991-95. According to the data available at the Geological Fund, 55 sand and 74 gravel deposits were under exploitation in 1994. As of January 1, 1995, the active sand reserves of category T were estimated at 140 million m³ and the probable reserves of category R at 171 million m³. For gravel, these figures were 30 and 31 million m³, respectively. The abrupt decrease in the number of deposits under exploitation and in the amount of reserves was due to the circumstance that many open-cast pits were closed down, the system of registration and account of the reserves was changed and the resources of deposits re-estimated according to the new requirements.

In the nearest years, the building activities are expected to stabilize and some 2.5 - 3.0 million m³ of sand and gravel is annually needed.

CLAY

Estonia's natural reserves of clay are practically unlimited, but in terms of quality parameters the clays are rather similar and do not allow a wide range of application. Clays are used in the manufacture of simple construction ceramics - bricks, stove tiles, drainage pipes, ceramic tiles, lightweight aggregate, as raw material in cement making and for isolation purposes in waste depositories.

The quality parameters of clays depend on their genesis. Cambrian and Lower- Ordovician marine clays, the most common type of clay in Estonia, have the illite-chlorite mineral association persistently dominating in their composition. The fairly widespread Quaternary glaciolacustrine clays of the last glaciation are higher in illite and lower in chlorite. Occasionally kaolinite occurs, but its proportion is low (10-12%) and does not exert any remarkably effect on the properties of clay. The granulometric composition of these clays is inconsistent and, accordingly, the content and composition of loose particles (quartz and carbonates) resulting from the denudation of Palaeozoic, Ordovician and Silurian rocks vary in a great deal. The latter circumstance determines the field of application for these clays (stove tiles, keramsite).

Estonia's higher quality commercial clay deposits occur in the Middle-Devonian Burtnieki and Gauja stages outcropping in the southern part of the Republic. These clays are of complicated alluvial (delta) and marine genesis. As a result, several grey varieties of clays contain up to 30-40% kaolinite (Утсал и Утсал 1982, Pirrus 1995a), which raises the fusing point to 1380-1450°C, widens the baking interval and makes the clays suitable for the manufacture of higher quality products. The irregular bedding, small dimensions and complicated lense-like structure of the deposits is a major constraint on the commercial use of the clays (Tallinn & Pirrus 1992).

There are clays still higher in quality, but these deposits are commercially unworkable because of the great bedding depth. In the weathering crust of metamorphic and magmatic rocks of the pre-Cambrian basement the prevailing component is kaolinite, which in places forms pinkish-white clay lenses, some metres in thickness (Дилакторский и Архангельская 1953). Some chamosite (Кууспалу и др. 1971) and hydrothermal sepiolite occurrences have also been recorded in the weathering crust (Кууспалу и др. 1973). In the sedimentary rocks of the Vendian complex of subcontinental genesis resting directly on the weathering crust, the proportion of kaolinite is still significant reaching often 40-60%, occasionally even 80-90% (Voronka Formation) of the clay component. On this level, chamosite has also been found (Пиррус 1973). However, all these occurrences are small in volume and bedded at great depths (100-300 m) which makes them practically unworkable. For the same reason, the 10–40-m-thick deposit of illite clays of the Late-Vendian Kotlin Formation in eastern Estonia is only of theoretical commercial value.

Cambrian clays are known as "blue clays". The term is not quite correct, because the colour is greenish-grey or purplish-grey rather than blue. The most noteworthy feature of these clays is their plasticity which they have maintained despite the venerable geological age of about half a milliard years. Owing to their normal marine genesis and long-term deposition in a relatively deep body of water, they spread as a uniform 60–70-m-thick deposit all over Estonia. Commercially, they are workable only in a restricted outcrop area on the southern coast of the Gulf of Finland. Elsewhere south of the North-Estonian Klint, they are overlain by Ordovician and Silurian sedimentary rocks and their mining is out of the question. The composition and quality of the Cambrian clays are much the same throughout the outcrop area. The contours of the deposits have been established on the basis of nongeologic criteria. Of the five deposits explored (Table 68), clay from Kunda has been used in the cement industry during more than a hundred years, the other deposits have provided material for the ceramics industry. The reserves of clay are sufficient to meet industrial needs in Estonia during more than a hundred years.

In the Lontova Formation, the best-quality blue clays occur in the Kestla Member (third from the bottom), where the evenly distributed multicoloured clays form a 20–25-m-thick layer (Fig. 235). However, the underlying greenish-grey clays of the Mahu Member (*ca.* 20 m) with a higher silt content and the clays of the uppermost, Tammneeme Member (up to 10 m) with occasional interlayers of siltstone are also suitable for use. The blue clay deposit in a wider sense covers also the lowermost, clayey part (6m) of the overlying Lükati Member which belongs to the Dominopol Stage. This clay, earlier mined at Kopli and Kolgaküla (Fig. 236) deposits, is not in

Table 68. Reserves (01.01.1995) of industrial clays in Estonia

No	Deposit	County	Cat. T, m³		Cat. R, m³		Output in 1994, m³
			Active	Passive	Active	Passive	
	A. Cambrian clays						
1.	Kopli	Tallinn	-	663,000	-	-	-
2.	Kallavere	Harju	1,320,000	6,608,000	-	184,533,000	-
3.	Kolgaküla	- " -	1,445,700	857,800	-	722,000	2,000
4.	Kunda	Lääne-Viru	17,528,000	-	11,059,000	-	27,000
5.	Aseri	Ida-Viru	2,609,300	4,044,000	-	6,585,000	50,700
	B. Devonian clays						
6.	Arumetsa	Pärnu	1,340,200	680,400	-	-	7,000
7.	Joosu	Võru	2,290,300	415,000	-	-	-
8.	Küllatova	- " -	2,298,700	-	2,412,000	-	-
	C. Quaternary clays						
9.	Sakla	Saare	610,000	-	652,000	-	-
10.	Türi	Järva	594,300	-	205,300	-	-
11.	Vana-Vigala	Rapla	-	-	4,851,000	-	-
12.	Tohvri	Viljandi	170,500	-	85,900	-	-
13.	Perametsa	Võru	54,300	-	-	-	100
14.	Määsi	- " -	29,200	-	-	-	100
						Total:	86,900
						Among this for cement:	27,000

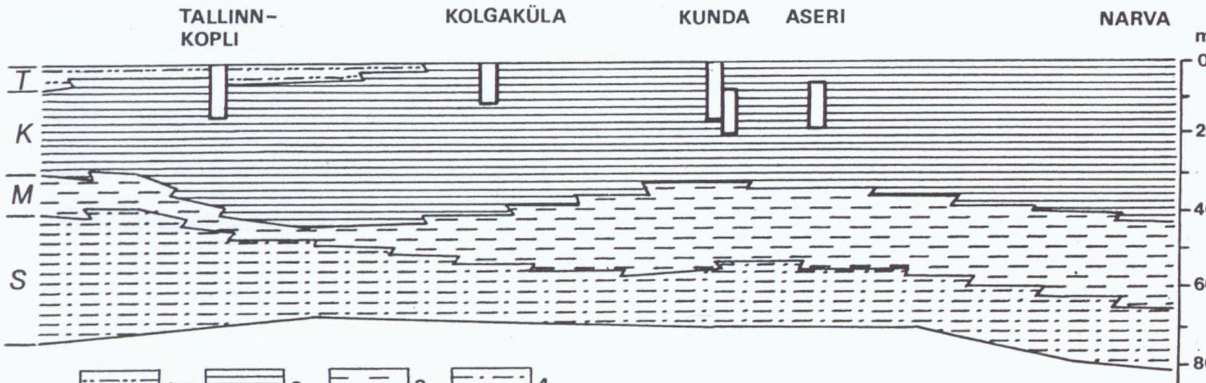


Fig. 235. Mined levels of Cambrian blue clays of the Lontova Formation. The lithologically differing subdivisions (members) of the deposits are marked with letters. 1 - clay and silt; 2 - clay; 3 - silty clay; 4 - sandstone and clay.

use today, because it abounds in lense-shaped interlayers of silt- and sandstone. Actually, the clays of the Lükati Formation have formed of the material redeposited from the underlying Lontova Formation which explains their similar properties and outlook.

Cambrian clays are used in the manufacture of bricks and, to a lesser extent, for the production of drainage pipes at Aseri. Production of roof tiles is under introduction, however, there still remain some technological problems to be addressed. Some 100,000 tonnes of clay is used in the cement industry. Like other fusible clays, Cambrian clay could be used in producing lightweight aggregate for concrete, however, there is no market for such kind of product today.

The low fusing point (1200-1290°C) and the short baking interval (80-120°C) of Cambrian clays causes technological problems. A serious disadvantage is also the relatively low plasticity (8-26) resulting from the unretrievable adhesion of the originally fine-dispersed clay particles. All this complicates preparation of raw mixture for molding and regulation of the firing process.

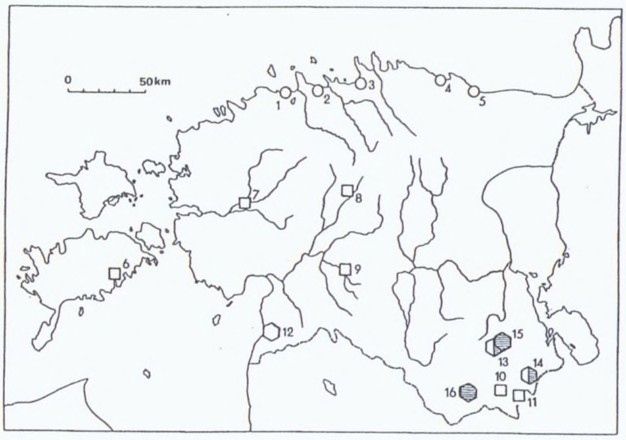


Fig. 236. Clay deposits in Estonia. **1-5 - Cambrian clays** (circles): 1 - Kopli; 2 - Kallavere; 3 - Kolgaküla; 4 - Kunda; 5 - Aseri. **6-11 - Quaternary clays** (squares): 6 - Sakla; 7 - Vana-Vigala; 8 - Türi; 9 - Tohvri; 10 - Perametsa; 11 - Määsi. **12-16 - Devonian clays** (hexagons): 12 - Arumetsa; 13 - Joosu; 14 - Küllatova; 15 - Süvahavva; 16 - Sänna. Shaded symbols denote refractory clays.

The admixture of pyrite (up to 1.8%) and its decomposition compounds lower the quality of clays complicating the firing process and causing white spots on products.

The chemical composition of clays is rather persistant as evidenced by the small amplitude of variation in respect of average values (Table 69). Against the background of the general illite prevalence, the mineral composition of the clay component is rather complicated. The finest fraction abounds in mixed-layered minerals. In illite, generations of different age have been established (Pralow 1938, Викулова 1952, Дилакторский 1956, Gorokhov *et al.* 1994).

Geotechnically, Cambrian clays behave as typical semi-rock surfaces, however, in a long-term contact with water they may turn plastic and their geotechnical properties will significantly deteriorate. This phenomenon is well-known in the upper part of the deposit, immediately below the Ordovician-Cambrian aquifer system (Пиррус и Саарсе 1979) and has become topical in the ever widening construction of harbours on Estonia's north coast.

The clay of the Varangu Formation, occasionally up to 2.5 metres thick, lies immediately below the Ordovician limestone and is recognized as a baking clay. Its application in the ceramics industry would pay off only if the phosphorite overburden, within which it occurs, were subject to complex use. As a deposit, the Varangu clay has not been studied so far.

Devonian sedimentary rocks in the southern part of Estonia occur mainly as lenses of irregular shape, size and composition in sand- and siltstones.

Two types of Devonian clays may be distinguished: low-quality reddish or variegated clays and higher-quality grey clays. The latter are of commercial significance, since they have a low content of iron compounds and high (20-40%) content of kaolinite (Althausen 1938, Öpik & Krusenberg 1929, Дилакторский 1956, Утсал 1968, Утсал и Утсал 1982, Pirrus 1995a). These are refractory clays, with the fusing point at 1350-1480°C and a relatively wide baking interval up to 200-250°C. Grey clays are related to the Burtnieki or Gauja stages and associate with the relatively light sands of coastal plains or near-shore sea floor (Fig. 237). Five deposits have been explored - Joosu, Võlsi, Süvahavva, Sänna and Küllatova (Table 68).

The genesis of grey clays is still open for discussion. According to Latvian geologists, these clays were formed in the near-shore sea under the conditions of subwater sliding processes (Сорокин 1981); Estonian geologists support their alluvial origin (Tallinn *et al.* 1970).

Reddish and variegated clays are lower in quality, rich in iron compounds and suitable for application in brick-making only. Even this use is restricted by the lense-like occurrence and small reserves. These clays can be mined only on the

Table 69. Chemical composition of Estonian industrial clays

Component	Cambrian blue clays 21 samples		Devonian grey clays 9 samples		Quaternary glaciolacustrine clays 57 samples	
SiO_2	55.55-66.97	60.40	48.08-63.85	53.83	47.20-61.86	57.91
TiO_2	0.24-1.54	0.84	0.80-1.72	1.24	0.22-2.11	0.85
Al_2O_3	13.98-22.07	17.98	16.89-24.00	21.65	12.38-19.52	15.71
Fe_2O_3	1.95-8.03	4.14	3.02-5.07	4.25	4.17-8.57	6.64
FeO	0.49-3.49	2.33	0.20-0.72	0.44	1.10-2.38	
CaO	0.14-2.21	0.71	0.34-2.61	1.21	1.02-11.83	4.87
MgO	0.87-3.82	2.31	1.15-2.62	1.89	0.87-4.32	2.45
K_2O	3.40-6.43	5.12	3.16-6.22	4.55	2.30-4.70	2.84
Na_2O	0.04-1.78	0.45	0.17-1.68	0.82	0.60-3.00	1.59
CO_2	0.18-1.64	0.86	0.13-2.28	0.98	0.95-10.48	5.06
SO_3	0.10-0.54	0.37	0.02-0.17	0.09	0.02-0.15	0.11
Loss on ignition	4.08-6.54	5.47	3.56-6.86	6.28	4.64-12.70	7.19

deposits were they are plaiting with grey clays, *e.g.* on the margin of the exhausted Joosu deposit and, in future, most likely also on Küllatova and Sänna deposits. As a rule, the clays have a high iron content (Fe_2O_3 7-10%). The only exception is the deposit of brown clay which fills a valley cutting deep into the sandstones of the Aruküla Stage. Since 1995, the clay of the Arumetsa deposit has been mined for producing a lightweight aggregate for concrete and also as an admixture in the production of ceramic tiles in Tallinn.

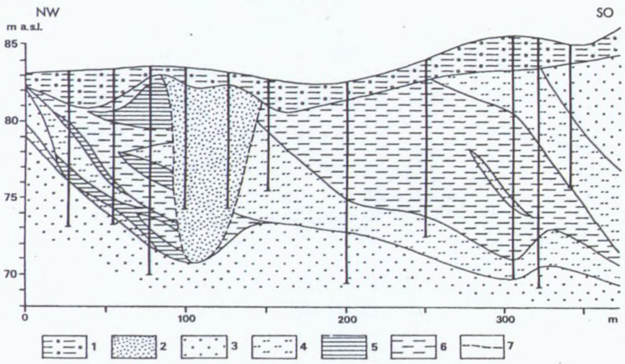


Fig. 237. The complicated mode of occurrence of Devonian clays in the Joosu deposit (after Tallinn *et al.* 1970): 1 - till; 2 - sand; 3 - sandstone; 4 - silt; 5 - red clay; 6 - grey clay; 7 - contour of buried valley.

The granulometric composition (Fig. 238) of Devonian clays varies in a wide range (fraction <0.01mm 45-77%), the plasticity number ranges from 4 to 28. They also differ in the presence of impurities, such as sand- and siltstone interlayers and lenses, carbonaceous and phosphorite concretions. For the above reasons, the Devonian clays are suitable, first of all, for a small-scale production at brickworks or occasional local brick yards, known from the last century.

The prospects to find new clay deposits in southern Estonia where the Quaternary cover is thick, are rather scarce. However, the noteworthy content of kaolinite in the area where the Gauja River flows into the Gulf of Riga (Лутт и Калласте 1992) shows some promise.

Quaternary glaciolacustrine clays are widespread in Estonia, forming in places extensive lense-like deposits easy to quarry. The distribution area covers the Fore-Klint Lowland with the ancient valleys, lower parts ot the North-Estonian limestone plateau (particularly in western Estonia), ancient valleys of southern Estonia and some hilly regions (Fig. 239). The conditions of clay formation differed with regions, which is well revealed in the texture (Pirrus & Saarse 1979, Saarse & Pirrus 1988) and composition of glaciolacustrine

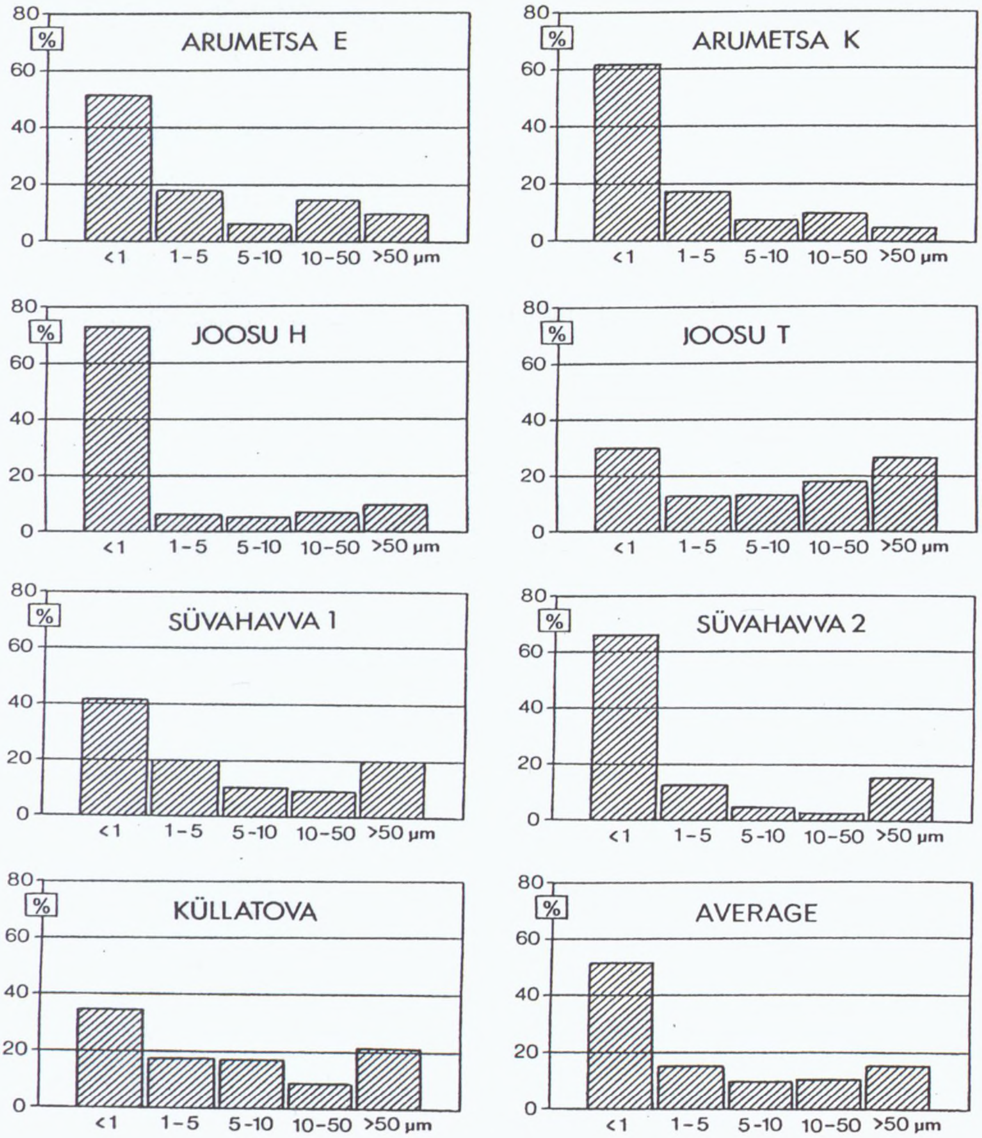


Fig. 238. Granulometric composition of Devonian clays in some deposits.

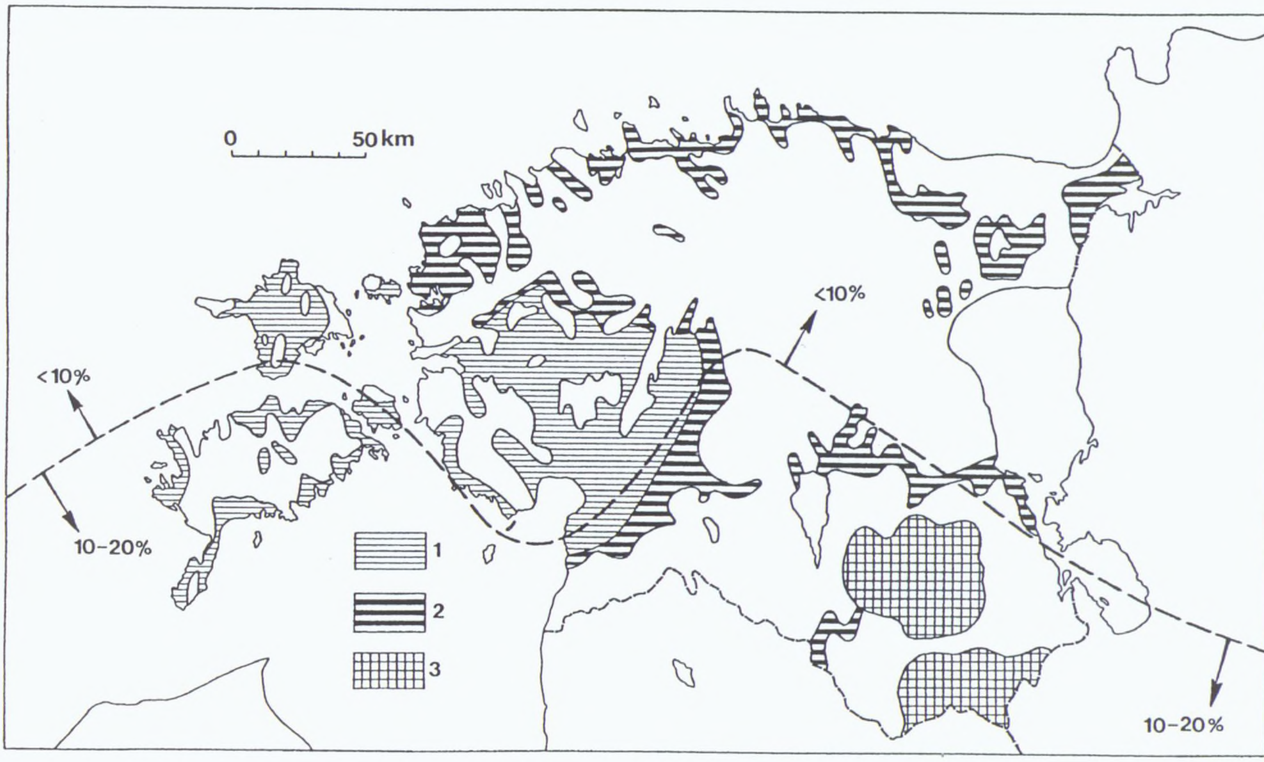


Fig. 239. Distribution of glaciolacustrine clays with different texture (after Pirrus & Saarse 1979): 1 - clays of large proglacial lakes with weakly differentiated summer and winter varves; 2 - clays of small glacial lakes with well-differentiated varves; 3 - clays, mostly unstratified or with unclear deposition of annual layers. Dotted line marks the boundary of marked increase in the carbonate admixture (after Пиррус 1964).

clays. The clays belonging to this type represent the most fine-grained component washed out of till during the recession of the continental ice. During the glaciation, very little clayey weathering product was produced and the initial rocks in Fennoscandia did not have any noteworthy weathering crust either. Therefore, the clay component of the Palaeozoic bedrock forms the major part of glaciolacustrine clays (Пиррус и Раукас 1963, Пиррус 1968). Important sources were the Cambrian blue clays which crop out at the foot of the klint, the Ordovician-Silurian limestones and especially the marls with a total thickness of more than 300 m. The contribution of Vendian and Devonian sediments seems to have been less. As a result, the glaciolacustrine clays are dominated by minerals of illite group with an admixture of degraded chlorite and occasional kaolinite. The illite component is rich in mixed-layered varieties which essentially contribute to the plasticity and workability of these clays. However, they also provide the clays with a higher water-binding ability which accounts for the high natural humidity of these clays. But what's even more important, the coarse component contains, besides quartzose sand, also fine-dispersed carbonaceous matter which is the weathering product of carbonate rocks (Table 69). The latter manifests itself clearly in central Estonia (Fig. 239), first as an admixture of calcite and towards southern Estonia and Latvia as a persistent admixture of dolomite (Пиррус 1964). The carbonaceous component controls the baking properties of clays and determines the colour of the final product which is often light; high porosity and water-absorption ability of clays are maintained. Glaciolacustrine clays are characterised by the rhythmic alternation of summer and winter layers. The phenomenon is absent only in shallow-water basins in the hilly topography of southern Estonia where clays, often with a high sand content, are encountered (Саарсе 1976б). Among the glaciolacustrine clays there are also varieties, suitable for the production of expanded clay aggregate – keramsite.

The granulometric composition (Пиррус 1963) of clays varies greatly even within the vertical section of one and the same deposit (<0.01 mm 42-99%, <0.001 mm 13-74%). The number of plasticity varies from 8 to 29 (more frequently from 14 to 18). Often the properties of clays need to be improved with an admixture of sand. Glaciolacustrine clays have a low fusion point at 1200°C, the baking interval is 40-100°C.

The use of glaciolacustrine clays is restricted by their high natural humidity (40-70%), particularly in the area of high-quality clays in western Estonia. For this reason, clay has to be dried before its application, but this would raise production costs. Another limiting factor is the occurrence of lime concretion, formed as a result of redistribution of carbonaceous material, which lowers the quality of products and needs to be ground.

Because of the above-mentioned technological complications and high natural humidity, the glaciolacustrine clays are not currently used, although at the beginning of the 20th century there were more than twenty brickyards active in Estonia. Of all Estonia's clay deposits, six are prospective (Fig. 236, Table 68). In consideration of local needs and increased transportation costs, glaciolacustrine clays may be taken again in use in the near future.

The resources of explored T and R category glaciolacustrine clays are estimated at 1.4 million cu m and 5.8 million cu m, respectively. However, there are good prospects for the explored reserves to be increased.

LAKE AND SEA MUDS

The Estonian offshore sea, coastal and inland lakes and mires hold large reserves of fine-grained sediments high in organic matter. The mud from offshore sea has been used for curative purposes for several hundred years by local population and during almost two hundred years in special mud-baths. The use of lake mud has been very limited.

Lake mud is an organo-mineral sediment, formed under freshwater conditions. In the basis of the Estonian lake mud studies lies the genetic classification worked out in Russia (Пидопличко 1975, Ильина и др. 1983). The main criterion underlying this classification is the origin of the material (from the lake itself or outside the lake) prevailing during the formation of the lake mud deposit. In this classification, lake muds are divided into biogenic, clastic and mixed. The latter falls into subtypes depending on the ratio of organic and mineral parts and their composition. The lowest content of organic matter in a deposit considered as lake mud, is 15% of the dry mass (Ramst 1992). In the list of the Estonian mineral resources the lake muds with a high content of carbonaceous material (the dry solid matter contains at least 40% of CaO) are registered as a separate mineral resource - lacustrine lime.

Sea mud is a sediment, formed under brackishwater conditions. The content of organic matter is usually less than 10% of dry matter mass. In the coastal lakes, which isolated from the sea in the late geological past (*e.g.* Mullutu-Suurlaht on Saaremaa Island, Fig. 240), sea mud is sometimes covered with a layer of lake mud.

The composition of lake mud is controlled by the climate of the region, the shape of the lake depression, the hydrological regime of the lake and the lithological composition of the rocks in the recharge area. The main constituents are organic matter, the particles of sand, silt and clay which have been transported into the body of water, and carbonaceous material of terrigenous or biochemical origin. Usually there is no clear boundary between the lake mud and the underlying clayey lacustrine or glaciolacustrine deposits. Depending on the composition, the colour of the lake muds varies between light beige and black, and its consistence between jelly-like and plastic. The average volume weight of lake mud is 1.05 t/m^3 and the average water content is 90%. As to the grain-size distribution, the 0.05-0.01 mm fraction is most common accounting for 30-50%. The content of pelite fraction increases towards the floor of the layer. As a rule, the content of sand fraction is low; the increased sand content is characteristic of the muds, deposited in the nearshore zone of big lakes (abrasive influence of wave action). Lake muds have almost the same mineral composition as the rocks around the lake depression. The main minerals in lake mud are quartz, calcite and pyrite. The chemical composition of lake muds is variable depending on the grain-size composition and the content of carbonaceous matter in sediments. The organic part of lake mud consists of carbohydrates, humic acids, bitumens and nonhydrolyzed residue. Several bioactive components (vitamins, hormones, antibiotics) and microelements (Mn, Zn, Cu,

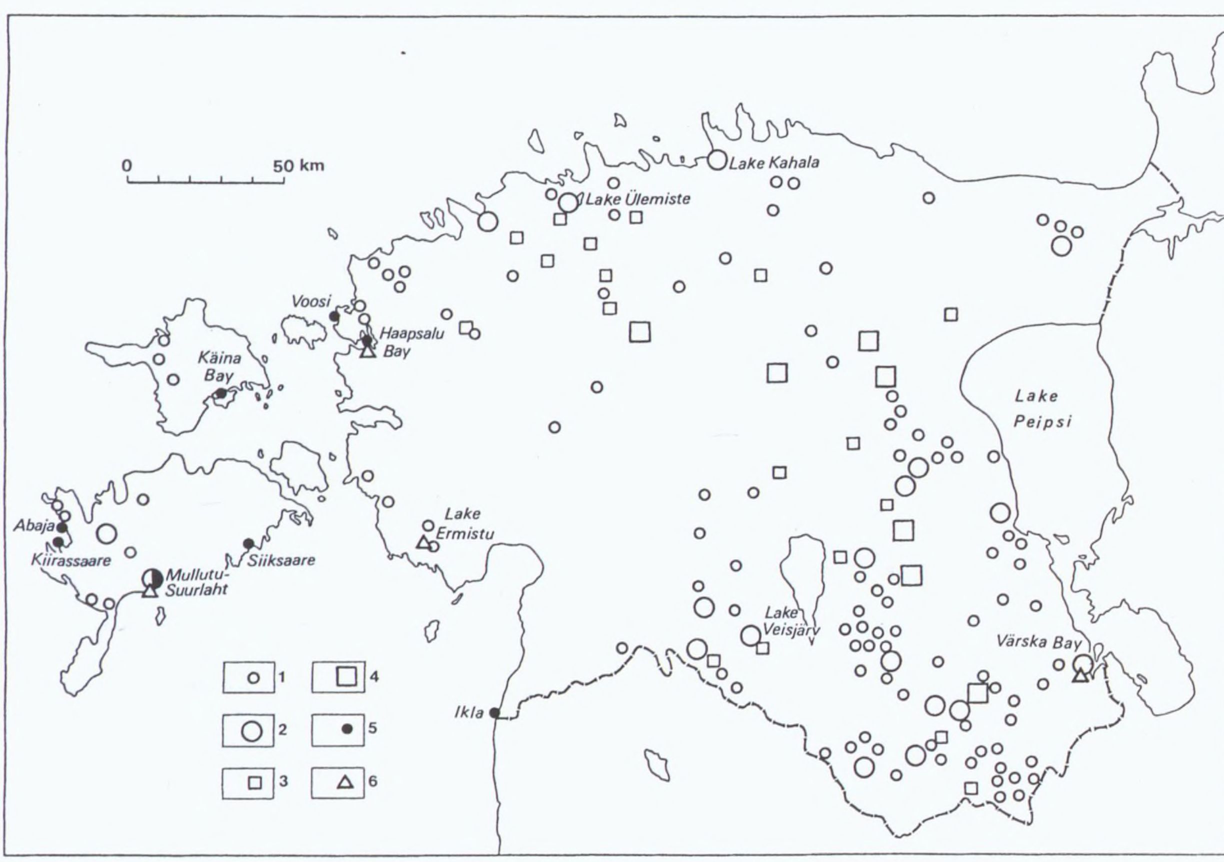


Fig. 240. Investigated mud reserves in Estonian lakes, mires and offshore sea: 1 - deposits in lakes with reserves under 1 million tonnes; 2 - deposits in lakes with reserves over 1 million tonnes; 3 - deposits in mires with lake mud reserves under 1 million tonnes; 4 - deposits in mires with lake mud reserves over 1 million tonnes; 5 - sea mud deposits; 6 - deposits in exploitation.

Mo, Co) are also present. The composition of lake muds has been studied by Veber (1973), Paap (Паап и др. 1981), Saarse (Саарсе 1976a, Саарсе и Каск 1981, Саарсе и др. 1981, Saarse & Arbeiter 1979, Саарсе 1994) and others.

In Estonia, the deposition of lake mud began approximately 10,000 years ago during the early Pre-Boreal, when the number of species of flora and fauna remarkably increased in the water basins, formed after the melting of the continental ice. The deposition of lake mud was especially intensive during the Atlantic period. The last decades have witnessed ever growing eutrophication of lakes and accumulation of deposits in the bodies of water due to human activity. Nowadays, the lake mud occurs almost in all Estonian lakes, except the young coastal lakes in the western part of the mainland and in the West-Estonian Archipelago. The thickest lake mud deposit (18 m) has been established in Lake Väimela Alajärv in the Võru County, southeastern Estonia (Veber 1964). The greater part of Estonian mires have formed as a result of filling up of shallow bodies of water. In almost half of the area covered by mires the peat layer is underlain by lake mud. The distribution and thicknesss of lake muds are discussed in the works by Veber (1964), Paap (Паап и др. 1981) and Saarse (Саарсе 1994).

Research into lake muds was initiated at Tartu University at the beginning of the 20th century. The more outstanding scientists included L. Zur Mühlen, H. Riikoja and H. Habermann. At present, the studies are basically carried out by the researchers of the Institute of Geology and the Geological Survey of Estonia. The latter institution is mostly involved in estimating, calculating and confirming the reserves. Only the lake mud beds, exceeding 0.5 m in thickness, are registered as mud resources. As a result, lake mud reserves in 127 lakes and 240 mires have been registered. According to the state of study, proved reserves, possible reserves and prognostic reserves are distinguished. Table 70 presents summarized data on the active lake mud reserves in the lakes studied. As in possible and prognostic reserves the lake mud and lacustrine lime reserves have not always been separated, this data is somewhat inexact. The reserves in tonnes are given at 60% water content.

The mud reserves studied are of uneven distribution (Fig.240). The reserves are remarkable in the lakes of northern Estonia, Vooremaa and southeastern Estonia, and insignificant in the West-Estonian Lowland, West-Estonian Archipelago and northeastern Estonia. The most important deposits are located in lakes Ülemiste (22.8 million m^3) and Kahala (11.5 million m^3) in northern Estonia, and in Lake Veisjärv (18.9 million m^3) in southern Estonia.

The reserves of lake mud suitable for producing fertilizers (Veber 1973, Саарсе и Каск 1981, Саарсе и др. 1981) and forage, and usable as curative mud (Kask *et al.* 1991) are large, however, they have been little used so far. Several attempts to excavate mud from lakes Lahepera, Maardu, Väimela and Harku have failed. The hundred thousands cubic metres of mud pumped out from Lake Ülemiste, the potable water reservoir of Tallinn, haven't found use either. Only the mud from lakes Peipsi (Värska Bay), Suurlaht and Ermistu is used for healing purposes.

Sea mud as a natural organo-mineral substance with certain plastic properties and high heat capacity has been traditionally used as curative mud. It comprises therapeutically active compounds (salts, gases, biostimulants) and living microorganisms (Lüttig 1990).

In the middle of the 19th century, the deposits of Rootsiküla, Suurlaht and Haapsalu were the only known sea mud deposits in Estonia, but during the first half of the 20th century additional 57 deposits were discovered (Maide 1940). In 1935, the Municipality of Pärnu ordered exploration of sea mud reserves between Virtsu, Pärnu and Ikla. In 1937, deposits and approximate reserves of sea mud, usable for curative purposes, were documented among other mineral resources by the State Parks Management (later the Nature Conservation and Tourist Institute). Most of the deposits were small and never taken into use, many of them are not existing any more. However, the larger deposits (Haapsalu, Mullutu-Suurlaht) are supplying Estonian mud-baths with curative mud even today. The deposits of sea mud at Haapsalu, Käina and in Mullutu-Suurlaht, as well as the lake mud deposit at Värska (Fig. 240) were registered as the deposits of state importance by the Commission of Estonian Mineral Resources.

After World War II, mud deposits were discovered and studied at many locations in western Estonia. As a result, several manuscripts were compiled by the researchers of the Geological Survey of Estonia and the Institute of Geology of the Estonian Academy of Sciences. The data on the reserves of investigated sea mud deposits is presented in Table 71. Of the eight sea mud deposits explored, only two (Haapsalu and Mullutu-Suurlaht) are under exploitation. Table 72 presents for comparison some properties of sea mud from the Haapsalu and Mullutu-Suurlaht deposits and lake mud from the Värska deposit (Ilomets *et al.* 1993, Kask 1993).

The mud of Haapsalu deposit has been used for medical

Table 70. The explored active lake mud reserves of the Estonia (01.01.1996)

	Cat. T			Cat. R			Cat. P		
	Area ha	Reserves mil. m^3	Reserves mil. t	Area ha	Reserves mil. m^3	Reserves mil. t	Area ha	Reserves mil. m^3	Reserves mil. t
In lakes	1329	42.8	11.3	6248	163.1	63.5	655	7.8	4.9
In mires	238	2.7	0.8	1158	13.4	4.1	7600	82.6	30.0
Total	1567	45.5	12.1	7406	176.5	67.6	8255	90.4	34.9

Table 71. Active reserves of the explored sea mud deposits of Estonia

Deposit	Reserves, th m^3		
	Cat. T	Cat. R	Cat. P
Haapsalu	210.9	36.1	-
Voosi	206.5	45.5	-
Ikla	22.5	2.5	-
Käina	-	768.6	-
Mullutu-Suurlaht	984.0	-	702.2
Siiksaare	-	-	800.0
Kiirassaare	-	-	90.0
Abaja	-	-	80.0

purposes since the beginning of the 19th century, in folk medicine even earlier. The deposit is characterised by an average thickness of 0.6 m (max. 1.3 m), low density (1.28 kg/m³), high water content (60-70%) and good plastic properties (shear strength 3000-4000 dyn/cm²).

The mud in the southern part of the coastal lake Mullutu-Suurlaht on Saaremaa Island is a fine-grained sediment with an organic matter content 15-40%. It has been used in local spars since the middle of the 19th century. The average thickness of the deposit is 1.0 m (max. more than 2m). The mud is characterised by a low density (1.10 kg/m³), high water content (over 80%) and shear strength 1000- 2500 dyn/cm².

The mud at Värska is a fine-grained lake sediment, with the content of organic matter reaching 50%. The average thickness of the mud layer is 4.0 m. The mud has a high H_2S (12-70 mg/100 g wet mud) content and high heat capacity.

The annual output for curative purposes from the Haapsalu deposit is 1000-2000 m³, from Mullutu-Suurlaht 500-1000 m³ and from Värska Bay 600-800 m³. The present rate of output meets the demands of small local spars. In Estonia, the deposits of state importance can be taken into use only with the permission of the Ministry of the Enviroment, the permission for exploitation of small deposits is given by District Councils. Protection of sea and lake mud deposits is ensured with the establishment of sanitary regions and zones around them.

Table 72. Comparison of some properties of muds from Haapsalu, Mullutu-Suurlaht and Värska deposits

Characteristic feature (average)	Deposits		
	Haapsalu	Mullutu-Suurlaht	Värska
Volume weight, kg/ m³	1.26	1.10	1.05
Dry matter, %	18.12	7.27	8.92
Weight loss at 600°C relative to dry matter, %	16.09	30.54	51.83
Bitumic matter, %	1.67	1.32	1.95
Elemental content, %			
C	5.51	13.77	22.40
H	1.22	2.26	3.09
O	8.29	15.49	22.33
N	0.64	1.35	2.06
H_2S in natural mud, %	0.018	0.012	0.096
Free humic acids in dry matter,%			
Humic acids (Ha)	0.56	0.41	1.74
Fulvic acids (Fa)	1.16	1.75	1.87
Hymatomelanic acids (Hy)	0.07	0.12	1.38

LACUSTRINE LIME

Lacustrine lime is an unconsolidated carbonate rich sediment, formed in shallow hardwater lakes. Highly calcareous and rather consolidated lime is termed lacustrine chalk. In the regions with the calcareous bedrock and limy glacial sediments the runoff, soil and ground water are enriched with Ca^{2+} and HCO_3^- from weathering reaction between water, carbon dioxide and calcium (Håkanson & Jansson 1978): $CaCO_3 + H_2O + CO_2 = Ca^{2+} + 2HCO_3^-$. Changes in the chemical or physical (first of all CO_2 and temperature) conditions may cause precipitation of calcite which accumulates on the lake bottom as lacustrine lime. In the littoral zone of hardwater lakes the formation of lacustrine lime is promoted by carbon dioxide consumption by algae or macrophytes. The precipitation of carbonates is restricted to water depths less than 6 m (Мяннилъ 1964, Саарсе 1994). In deeper lakes, the calcite settling through the thermocline, comes into contact with colder water in which the CO_2 content is higher, and may dissolve again.

Mollusc shells play an important role in the accumulation of carbonates. In Estonian deposits of lacustrine lime, 36 species of molluscs have been identified (Мяннилъ 1964).

Research into Estonian lacustrine lime goes back to the end of the last century when the ancient Lake Kunda was subject to studies (Grewingk 1882). Up to the middle of the 20th century, the studies of this kind were not particularly intensive. The first overview of the Estonian lacustrine lime deposits described 54 localities from southern Estonia (Hallik 1948). The most comprehensive studies on lacustrine lime in Estonia were made by Reet Männil (Männil 1961, Мяннилъ Р. П. 1963a, 1963b, 1964, 1967). In the candidate thesis (Мяннилъ 1964), she summed up the results obtained through the study of 167 localities and more detailed records on the stratigraphy, chemical composition and malacofauna of 43 localities.

In the 1980s, the researchers of the Institute of Geology undertook to study the isotopic composition of lacustrine lime from several localities, including Tapa, Turvaste Valgjärv, Äntu and Võrtsjärv (Fig. 184). The results obtained were partly published (Punning *et al.* 1983). Saarse (Саарсе 1994) summarized the results of the lacustrine lime investigations in the contemporary lakes and synthesized the data of their composition.

Researchers of the Geological Survey of Estonia carried out geological exploration with the calculation of resources on the Varangu deposit and preliminary estimation of possible resources in many other localities. The results of the investigations have not been published.

The lacustrine lime beds are located unevenly. Most of them are concentrated in Upper Estonia and northwestern Estonia in the regions with a favourable geomorphological situation, carbonate bedrock and carbonate-rich Quaternary deposits. Five main regions (Fig. 2) are differentiated (Männil 1961, Мяннилъ Р. П. 1963а, б): Pandivere Upland, Vooremaa Drumlin Field, Võrtsjärv Lowland, slopes of the Sakala Upland and Haanja Heights, northwestern Estonia.

The thickest lacustrine lime beds, richest in $CaCO_3$, are located on the Pandivere Upland where they were formed mainly in narrow valleys. Most of them are small in area, rather thick (5-6 m), and with a complicated structure. The lacustrine lime beds in northwestern Estonia and Võrtsjärv Lowland are large in area, but thin (less than 2 m), commonly underlain by clay or sand. In the Vooremaa Drumlin Field the lacustrine lime has been deposited mainly in the littoral zone of deep lakes. Many of those beds, like in Soitsjärve, Kivijärve and Kaiavere, are thick (2-3 m) and large in area. In the Sakala Upland and Haanja Heights, the lacustrine lime beds are thin and very small in area.

Lake sediments, including calcareous ones, can be classified according to several principles and parameters. Unfortunately, there is no generally accepted classification for lake deposits. The interpretation of the lowest content of $CaCO_3$ in a sediment considered as lacustrine lime varies from 40-50% (Hallik 1948, Саарсе 1994, Мянниль 1964) to 90% (Merkt *et al.* 1971) of the dry mass. The average $CaCO_3$ content in Estonian lacustrine limes is 70-95%, being higher in northern (80-95%) and lower (70-90%) in southern Estonia. The highest $CaCO_3$ content has been fixed in drained lakes (Tapa, Varangu, Siplase) which due to the decrease in water level occurs above the ground-water table, favouring the oxidation of organic substances.

Lacustrine lime is a white- or beige-coloured sediment, due to the clay content it is grey in the bottommost part of the section. Its water content is usually 40-60%, bulk density about 1.40-1.50 t/m^3. Lacustrine lime consists mainly of particles with a diameter of about 0.05 mm or of aggregates with a diameter of about 0.15 mm (Мянниль 1964).The upper and lower limits on the contact with the other lacustrine or telmatic units are commonly diffuse. Lacustrine lime consists of calcite with some addition of dolomite, quartz, illite, kaolinite and pyrite (Льюкене и Утсал 1971, Саарсе 1994). In the pure lacustrine lime the calcite/dolomite/quartz ratio is 92/1/7 (Саарсе 1994). According to the unpublished results of investigations, carried out in the Geological Survey of Estonia, the lacustrine lime from the Varangu deposit contains in average 90.7% $CaCO_3$, 2.6% $MgCO_3$, 0.5% P_2O_5, 0.3% SiO_2, 0.2% FeO, 0.2% Al_2O_3, 0.1% Fe_2O_3, 3.8% organic substance and 1.6% other components in the dry matter mass.

The results of light isotope variation studies of the lacustrine lime from the Tapa and Võrtsjärv localities, are in good agreement with palaeoclimatic evidence (Punning *et al.* 1983) and show the promise of the lacustrine lime isotope composition for the purposes of palaeoclimatic reconstructions. Unfortunately, the obtained records are too fragmentary to allow any profound conclusion to be drawn.

The main components of lacustrine lime are carbonates, derived from the bedrock or calcareous till, complicating the determination of the real age of lacustrine lime by radiocarbon dating. According to pollen evidence, lacustrine lime started to accumulate in Estonia after a considerable climatic warming in the beginning of the Pre-Boreal, about 10,000 years ago (Мянниль 1964). The accumulation rate of lacustrine lime was rather high (about 0.5 mm per year) in the Pre-Boreal. The same accelerated trend continued in the Boreal. During the Atlantic period the accumulation decreased in most of North-Estonian lakes (0.4 mm per year in the first half and 0.2 mm per year in the second half of this period in average), but remained the same in the lakes of southern Estonia. In the Sub-Boreal and Sub-Atlantic it decreased considerably everywhere in Estonia due to the decrease in water level and the lakes being filled with sediments. Most of Estonian lacustrine lime beds are covered with a peat layer today. According to Männil (Мянниль Р. П. 1967), 25% of the total lime resources in Estonia have formed in the Pre-Boreal, 35% in the Boreal, 30% in the Atlantic, 10% in the Sub-Boreal and less than 1% in the Sub-Atlantic.

Evidence about the resources of the Estonian lacustrine lime deposits are scattered in different sources. A lot of data on the lacustrine lime sites and bedding conditions are contained in the manuscript reports of peat deposit explorations. As the localities have been investigated at different times and with different methods, the assessment of the total resources of the lacustrine lime is a complicated task. In Estonia, the lacustrine lime beds with a thickness of more than 0.5 m and with the CaO content above 40% (70% $CaCO_3$) of the dry mass are registered as lacustrine lime resources. The total amount of the Estonian lacustrine lime resources are estimated roughly at 250 million m^3. At Varangu – the only explored lacustrine lime deposit in Estonia which has been included into the list of the Estonian mineral resources – the reserves (cat.T) amount to 0.6 million m^3 (Table 73). The preliminarily estimated (cat. R) reserves of 3 deposits form 1.9 million m^3. The other 180 registered localities comprise prognostic resources (cat. P).

Table 73. Reserves of lacustrine lime

Category of active reserves	Area, ha	Reserves, million m^3
T	32	0.6
R	250	1.9
P	15,000	250

Almost all lacustrine lime deposits are located in the bogs (Fig. 224). The largest and richest are the Endla Mire System (*ca.* 40 million m^3), Tapa Mire (*ca.* 24 million m^3), Ulila Mire (*ca.* 20 million m^3), Epu-Kakerdi Mire System (*ca.* 19 million m^3) and Väinjärve Bog (*ca.* 15 million m^3). Only a few lakes, such as Porkuni, Äntu Sinijärv, Väinjärv, Turvaste Valgejärv (Fig. 184) contain rather considerable prognostic resources (Саарсе 1994, Saarse & Liiva 1995).

Lacustrine lime is suitable for producing lime, chalk, additional feed for the cattle and poultry. It has also been used in farming to neutralize the excessive acidity of soils and silage. In 1994, 4000 m^3 of lacustrine lime excavated on the Varangu deposit was used for additional feed for the cattle. The distribution and limited dimensions of deposits determine their use for local needs only.

MINERAL OCCURRENCES

Iron ore

During magnetometrical investigations conducted by the Topogr. Dept. of the Estonian Army in 1931-37, several magnetic anomalies were discovered, of which the Jõhvi anomaly was the biggest (Fig. 241). It was supposed that the anomaly was caused by a body of iron ore, located in the crystalline basement. To get more information about that ore body, the company "Magna Ltd." was established (Luha 1946) which carried out a detailed magnetometrical survey of the anomalous area. As a result of the survey, two apexes of the disturbance area were recorded, the distance between them being 3 km.

In the middle of the westward area, two deep boreholes were made, both reaching the crystalline basement. The team, headed by the mining engineers A. Linari and A. Differt and the geophysicist J. Kark, consisted of the best specialists of that time. The first borehole stopped at 505 m, the other reached a depth of 721 m, but neither of them fully penetrated the ore body. Nevertheless, both drillcores were described (A. Linari, J. Kark), sampled and analyzed during and after drilling. Mineralization occurs in garnet-amphibole-pyroxene, garnet-amphibole or, more rarely, in garnet quartzite. The mineralized formation with a thickness of 200-400 m is divided into two 60–100-m-thick parts by a layer of clay gneiss. The crystalline basement lies at a depth of about 240 m from the surface, the major mineralization starts at a depth of 100 m from the surface of the crystalline basement. The iron content of the ore varies between 25 and 34.4%. As an average, the iron ore presented for analyzing by Linari contained 31.15% Fe, 1.62% Mn, 41.83% SiO_2, 4.17% Al_2O_3, 2.41% CaO, 3.88% MgO, 0.42% K_2O+Na_2O, 0.08% P, 0.20% S, 0.02% Cu, 0.27% Ti, traces of Zn, 11.75% O, the water content was 0.45% and ignition loss 2.26% (Luha 1946). The investigations carried out by the Geological Survey of Estonia after World War II registered Mn, Cu and Zn anomalies in the same place: in iron ore the contents of Cu and Zn were 0.7 % and 0.42 %, respectively

The calculations of iron ore reserves are rather approximate because until now the body of ore has not been penetrated to the full extent. The diameter of the area is about 8 km. The possible reserves to the depth of 500 m are estimated at about 355 million tonnes and to the depth of 700 m at about 629 million tonnes (Петерсеплль и др. 1981).

The beneficiation experiments with the ore showed that special separation would give good results (Luha 1946). The concentrate obtained contained 65.1 % iron and the yield was about 85% (phosphorous separated in the course of beneficiation). To apply this method, the ore was milled to a 60 μm fraction.

Despite the convenient location in an economically well-developed oil-shale district only 5 km away from the sea, the deposit is of no commercial interest today on account of the depth at which it occurs and the existence of better or comparable deposits elsewhere making the establishment of a profitable and competitive mining company on the basis of the Jõhvi deposit impossible today (Raudsep & Räägel 1993).

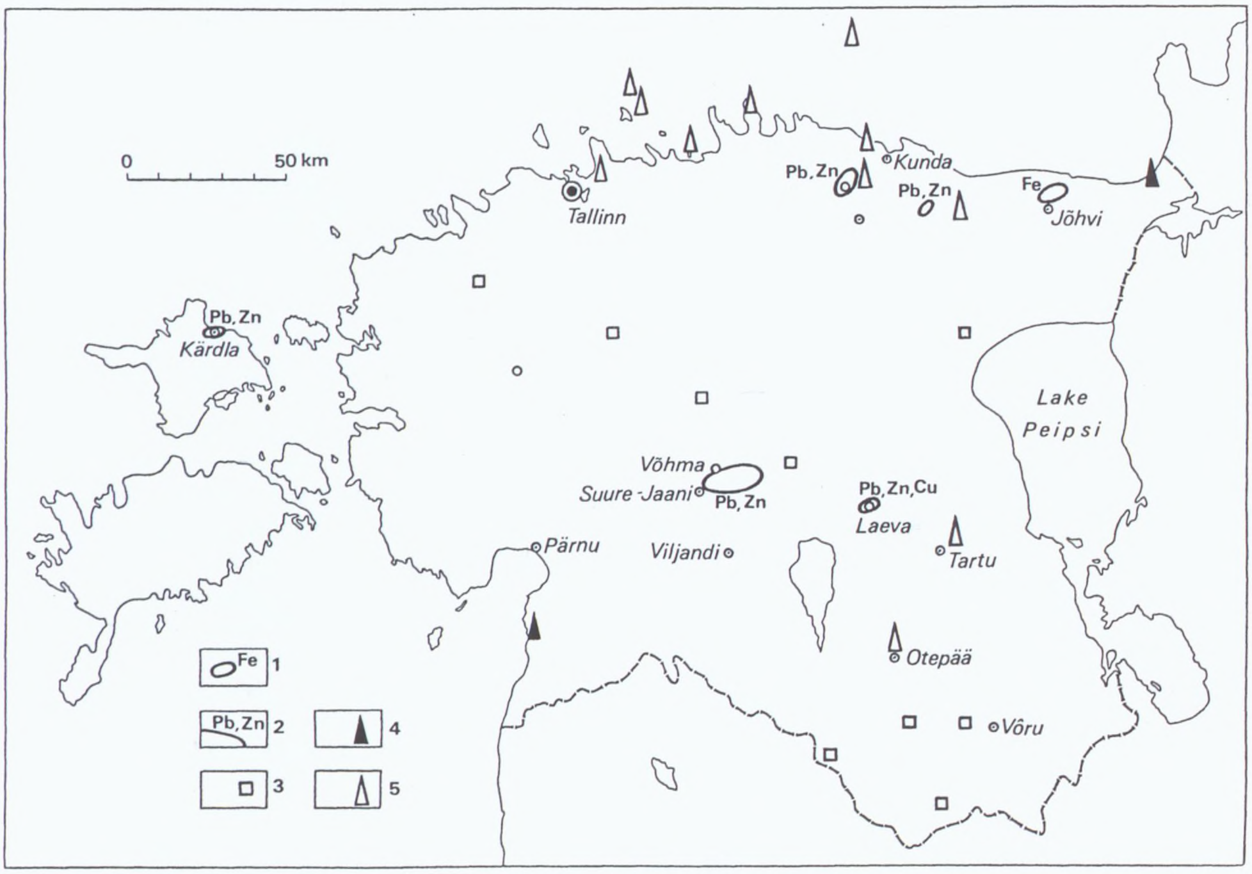


Fig. 241. Mineral occurrences: 1 - iron ore; 2 - polymetallic ore; 3 - ochre; 4 - diatomite; 5 - gas.

By now, several other iron ore occurrences have been recorded in Estonia; however, they are not comparable with the Jõhvi deposit in extent and ore quality.

Polymetallic ores (lead and zinc)

According to Grewingk (1861), in the present-day Viljandi County galenite was used for melting already several hundred years ago, during the so-called Swedish time (Luha 1946). In 1803 and 1853-55, it was studied in the vicinity of Võhma, Kõo, Arusaare, Paaksima and Maalasti. In 1803, trenches were dug and several hundred kilogrammes of samples were taken. In both cases, the investigations were finished because no positive results were gained. Afterwards, some boreholes were made in this region. In 1931, the Mining Board committed this task to I. Reinvald, during 1965-66 the drilling was performed by the Board of Geology and Protection of Mineral Wealth. However, the attempts to discover essential deposits always failed. The occurrences registered the presence of crystals and nodules of galenite (PbS) of hydrothermal origin in the dolomite of the Adavere Stage. Beside galenite, pyrite (Fe_2S) and sphalerite (ZnS) have also been found. The chemically weathered nests of galenite are covered by a white crust of cerussite ($PbCO_3$), while pyrite has oxidized and turned into ferric hydroxide.

According to Petersell (Петерселль и др. 1981), polymetallic mineralization has been recorded also in black shales in the crystalline basement of northern and northeastern Estonia (Uljaste, Haljala). In the Uljaste area the maximum content of lead and zinc is even as high as 5 %.

Recently, polymetallic mineralization has been recorded in almost all sedimentary rocks from the Vendian to Devonian. Intensive polymetallic mineralization is characteristic of *Dictyonema* argillite. In western Estonia the mineralization occurs by layers. The content of sphalerite is mainly 0.3-1.0, occasionally 3-4 %.

The galenite-sphalerite mineralization was also detected in the surroundings of the Kärdla meteorite crater, in the Haljala-Vanamõisa area (in Vendian sandstones), at Laeva (Pb, Zn and Cu anomalies from the Lower Ordovician to the Middle Devonian) and at Oostriku (Fig. 241).

Later, the regions of polymetallic mineralization in central Estonia were investigated. In the vicinity of Suure-Jaani Town, B. Sudov distinguished a polymetallic mineralization area (Fig. 241) in the Silurian and Devonian carbonaceous rocks where polymetallic mineralization has been registered in the rocks of several stages (Table 74).

Table 74. Distribution of lead and zinc in the Suure-Jaani area (according to V. Petersell in Raudsep & Räägel 1993)

Stage, rock	Number of samples	Average content, ppm Lead	Average content, ppm Zinc
Narva Stage (D_2nr), dolomite and domerite	8	161	84
Raikküla Stage (S_1rk), dolomite	91	24	38
Raikküla Stage (S_1rk), limestone	12	16	24
Juuru Stage (S_1jr), dolomite	5	15	1152
Juuru Stage (S_1jr), limestone	19	15	24

The reserves of polymetallic ores have never been calculated. The best ore bodies have already been exhausted, and neither at present nor in the future the utilization of polymetallic ores is expected.

Uranium

During the Soviet period, uranium was explored at Sillamäe in the course of some twenty years but, unfortunately, the results of these investigations are not available today. There are numerous uranium occurrences in Estonia's crystalline bedrock (Raudsepp & Räägel 1993).

At 40 points in the crystalline basement, the uranium content is several times in excess of its average content in the Earth's crust. In the granites of northern Estonia the maximum uranium concentration reaches 928 ppm. The maximum thorium concentration is 3215 ppm due to the presence of uraninite in the rock.

In the *Obolus* sandstone (phosphorite) the uranium content is mainly 13-47 ppm exceeding the corresponding clarke for sedimentary rocks. Uranium occurs in the shells of brachiopods and, consequently, its content depends on the content of P_2O_5.

In *Dictyonema* argillite the highest uranium contents have been recorded in eastern and western Estonia in the places where the argillite is rich in organic matter. In western Estonia, the uranium content is particularly high (up to 304 ppm) in an one-m-thick bed of shale resembling greatly the alum shale distributed in Sweden and used earlier for producing uranium. The organic-rich layers of the Toolse phosphorite deposit are also high in uranium (Пуура 1987), the average content being 212 ppm in seam A and 205 ppm in seam B. In seams C ja D these values are 10 ppm and 70 ppm, respectively.

Within the limits of the Toolse deposit the uranium content in the *Dictyonema* argillite varies between 3 and 850 ppm (average 192 ppm). The corresponding figures for thorium are 25-500 ppm and 13 ppm, respectively. The uranium mainly occurs in absorbed state and is related to clay minerals and organic matter.

In the Quaternary deposits, a rather high uranium content has been recorded in a peat layer on the Mustjõgi River in southern Estonia.

The uranium reserves in the crystalline basement have not been calculated. They have been better explored in the Kabala mining field of the Rakvere deposit where the possible reserves are estimated at 10,588 tonnes (average U content in the phosphorite 21.1 ppm).

Dictyonema argillite, with its total reserves around 60 milliard tonnes, is a remarkable depository of uranium. At Toolse the probable reserves of uranium have been estimated at 27,149 tonnes, but in the Kabala mining field *Dictyonema* argillite is practically missing.

Dictyonema argillite has been experimentally used for processing uranium. With the complex development of phosphorite deposits in future, processing of uranium from phosphorite and particularly from *Dictyonema* argillite may become possible. If *Dictyonema* argillite is treated with sulfuric or nitric acids, the yield of uranium will reach 90-100 %. With the annual production of *Dictyonema* argillite some two million tonnes, about 250 tonnes of uranium would be obtained.

Rare earth elements (REE)

The occurrences of REE (lanthanoitids and yttrium) have been recorded in the rapakivi of the crystalline basement in the Ereda, Neeme, Märjamaa, Naissaare and Taebla boreholes. In the above-mentioned places, the content of REE exceeds the corresponding clarkes 4-10 times, at Märjamaa even up to 100 times. In a sample from Märjamaa, the content of La and Ce was 1.4% and the concentration of Y 350 ppm (Raudsep & Räägel 1993).

In *Obolus* sandstone the content of REE is 350-840 ppm which is 4 times as high as the clarke for sedimentary rocks. The content of REE depends on the content of P_2O_5 and is highest in the phosphorite of the Toolse deposit.

In the Kabala mining field the possible reserves of the oxides of REE are estimated at 237,400 tonnes (at the average content of REE 0.0474% in the rock). Their utilization depends on the complex development of the phosphorite deposits.

Natural pigments

The main pigments are ochre (bog ore) and glauconite sandstone (Kivimägi 1974, Raudsep & Räägel 1993).

In terms of chemical composition, **ochre** is a ferric hydroxide which usually is mixed with sand. The colour of ochre depends on its iron content (it may reach 50%): the higher the content of iron, the darker the colour. Ochre occurs also in different Quaternary deposits as loose and porous forms (Luha 1946).

There are no precise data about ochre reserves. Several ochre deposits and occurrences are known (Fig. 241), but the level of the investigations performed is different; the data about the average thickness of a layer and the ore quality are either entirely missing or obtained by questioning local inhabitants (Raudsep & Räägel 1993).

Ochre can be used for producing yellow and reddish-brown paints (coloured earth). According to Luha (1946), several ochre deposits have been used, such as Laura (near the present border with Russia) and Võisiku (on the Põltsamaa - Võhma road). There is a rather detailed information available on the latter deposit. During 1937-40, at Võsiku 60.5 tonnes of ochre was mined and used mainly in producing yellow and red-coloured earth. However, there is no information as to the further use of the pigment. According to some data, ochre was mined at Pedeli and Mõniste deposits during the fifties and early sixties. In both cases, mining was performed by the Paint Producing Department of the Valga Integrated Plant of Local Industry.

With the re-establisment of small farms and private business, ochre may be taken into use as a cheap raw material for producing natural weather-proof paints. Therefore, all-Estonian revision of ochre deposits should be arranged to find out the deposits suitable for mining (Raudsep & Räägel 1993).

Natural and artificial outcrops of **glauconite sandstone** occur along the whole North-Estonian Klint. The glauconite content in sandstone is 60-70% and it can be used in the manufacture of coloured building materials. Electromagnetically separated glauconite may find use in producing water softeners (Vilbok 1949) and colouring pigments. Since glauconite tends to change its colour when heated, the colour spectrum of possible pigments is wide - from brown to green.

The resources of glauconite sandstone in the Toolse phosphorite deposit and Kabala mining field (Fig. 219) are estimated at 60.6 and 67.9 million tonnes, respectively.

Diatomite

In present-day Estonia, diatomite is found only in the western part of the Leekova Mire (Luha 1946) and in small quantities also at Rannametsa in the Pärnu County (Fig. 241).

Diatomite is a grey or dark-grey porous deposit, in which the proportion of diatom shells reaches 60%. Usually it contains also sand, clay or organic matter (Raudsep & Räägel 1993). Diatomite can be used as filling, isolation or absorbing material in several branches of industry.

The Leekova deposit is located 7-8 km northwest of the Town of Narva on the Fore-Klint Lowland. Of the whole deposit, 55 hectares have been investigated. The diatomite forms a layer with a thickness of 0.2-3.0 m. The productive layer, which is overlain by a 3-m-thick peat layer and underlain by peat and sand, consists of two parts: the upper one is 0.4 m and the lower one 2.1 m thick. The actual proved reserves are about 8 million m^3, or about 4 million tonnes (diatomite's volume weight is 0.39-0.79 g/cm^3).

Gas

The first natural gas occurrences were recorded in 1902 on Keri Island where during 1905-12 gas was used for lighting the local beacon. The daily amount of gas produced was about 560 m^3.

On Prangli Island, gas occurrences were studied in 1946-47 and in 1959. An emission of gas was observed in Quaternary deposits at a depth of 25-123 m. Gas emitted in the form of small bubbles, but occasionally intense eruptions with sand and gravel were recorded. The daily amount varied from some tens to 200-300 m^3, and the pressure from 1-2 to 8-9 atmospheres. The gas consisted mainly of methane, in places it contained also nitrogen (31-63 %). The reserves of methane were rather small – less than 17 million m^3 (Kattai & Pihlak 1993, Raudsep & Räägel 1993).

The occurrences of natural gas are also known from the sea bottom in the surroundings of Äksi and Mohni islands, near Viinistu, Kolga and Ihasalu, but also from the mainland near Püssi railway station, at Tartu, Pangodi, Otepää, *etc.* (Kattai & Pihlak 1993, Fig. 241).

In most cases, natural gas is stored in small local lens-like sand layers which are located in till serving as a non-conductor for gas. If only gas succeeds in penetrating this kind of reservoir, a considerably intensive emission will take place, but it ceases soon enough. Considering the above-described geological conditions, there is not much hope to detect any extensive gas deposit.

In some places (Pirita, Kopli Peninsula, Koliotsa), where boreholes were made to reach the Cambrian sandstones, emissions of inert nitrogen gas were registered. The gas spouted out together with water.

Oil

The first records on natural bitumens in Estonia date from the middle of the 19th century (Helmersen 1856). In 1905, during hydrogeological drillings near the Vaemla manor on Hiiumaa Island, the occurrence of oil-like liquid was registered. Oil prospecting was continued in the area during 1912-24, but the results have not been recorded (Raudsep & Räägel 1993).

By now, some 150 small oil-natural bitumen appearings are known. Conventionally, two major districts of the distribution of naphtides are distinguished (Каттай 1990а, Kattai *et al.* 1994) with different stratigraphical position, form of occurrence, degree of concentration in the rock, consistence and composition. In northeastern Estonia, small lenses of solid bitumens (asphaltite) have been discovered in Lower Cambrian and Lower and Middle Ordovician rocks.

Western Estonia (particularly Hiiumaa Island) is another district of bitumen appearances (Каттай и др. 1990б,1992) which occur in the Middle and Upper Ordovician and Silurian rocks. The depth of occurrences is variable – from the Earth's surface to a depth of 360 m.

RATIONAL USE OF MINERAL RESOURCES

Mineral resources, occurring in the Earth's crust in limited amounts, belong to the category of exhaustible and irretrievable natural resources. The ratio of their extraction in immense degree has been exceeding the ratio of natural increment, practically nonexistent according to the human time sense. For the present, the deposits, most favourable for mining, have been exhausted already. This fact as well as the inevitability to meet the requirements of the environmental quality underlie as a cause for the permanent and universal growth of expences of minerals industry. In this respect, the principles of rational use aquire a very special significance (Viiding 1976a, Viiding & Raukas 1984, Тээдумяэ и Вийдинг 1990).

Rational use of mineral resources means the possibly complete and complex mining and the most effective utilization of explored resources with minimum possible adverse impact to the Earth's crust and to the environment in general (Photo 72). In the centre of attention there have always been the explored and at the moment economically effective resources. Nevertheless, for future mineral policy options the knowledge of minerals beyond immediate commercial interest (prognostic and hypothetic resourses) is to aquire a determinating position indicating to the inevitability of their systematic monitoring and protection as the only replenishment for the resources in exploitation (Тээдумяэ 1988а, Teedumäe 1996b).

In most cases, the mineral development projects are of long duration and there is always some uncertainty component inherent. The interval between the moment of the discovery of the deposit and its introduction into industry may last for years, sometimes even tens of years. During that time there may occur some changes in technology, economy, legislation, *etc.* transforming the primary workability estimations of resources.

Finding of the most balanced and acceptable decision for mining action, considered through three general perspectives: technical feasibility, economic viability and environmental soundness, is a complicated task and in several cases difficult economic choises and environmental compromises are inescapable. The mineral policy in the former USSR, and in the Estonian SSR respectively, was based on the planned permanent growth of both explored reserves and production. The aspects of complete and complex utilization of mined reserves and the related environmental impact were considered either in a slight degree or ignored entirely (Кивимяги и Тээдумяэ 1971, Пуура 1987). On the initiative of Estonian scientists and specialists the problem was raised on governmental level, and in 1981 the relevant regulations enforced initiated the systematic studies on this matter. The programme "Optimal Use and Reproduction of Natural Resources and Environmental Protection in North-East Estonia " (Раукас 1984) was the first attempt to outline the complex approaches of mining activities and environmental protection for the optimization of mining industry. In the mid-1980s, the well-founded position of Estonian scientists on the decision of the former Soviet Union government addressing the development of large-scale phosphorite mining in the Rakvere area enabled to resist realization of this hazardous project.

With the restoration of the independent Republic of Estonia in 1991, possibilities were created for rearranging the Estonian mining industy according to the principles of rational utilization of mineral resources. Elaboration of the relevant legislation and criteria, based on the concepts acknowledged and currently in use in rest of Europe, is under way

The framework for environmental legislation in general is provided by the Law on the Protection of Nature in Estonia, approved on February 23,1990, for mineral resources in particular by the Law on the Earth Crust, adopted on November 9, 1994. A number of more specific acts and regulations have either come into force or are under development. On February 13, 1990, the Government of Estonia issued the regulation "On the System of Charges and Taxes in Nature Management " which established the system of economic measures aimed at improving the management of the environment and natural resources. The regulation is continuously being improved according to the experience gained through

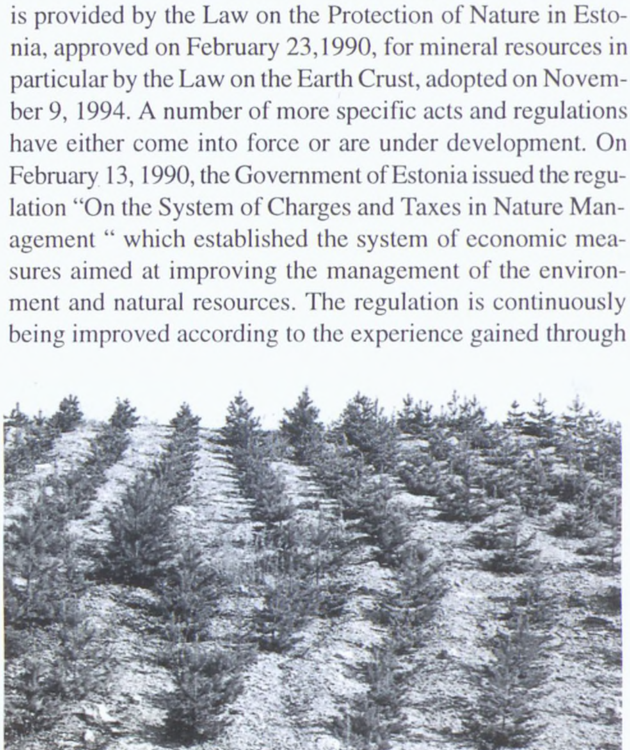


Photo 72. Young pine forest on the recultivated area of the Aidu quarry. *Photo by J. Lasman.*

the implementation practice. The Order of the Estonian Government No. 314 of November 13, 1992 on the Environmental Impact Assessment (EIA) establishes that for all the projects on extraction and processing of mineral resources EIA is required.

The responsibility for the development policies and strategy, coordination of management and related applied geological research of mineral resources as well as legislation and protection activities lies on the Ministry of the Environment of Estonia. All reports of geological exploration of the deposits to be taken into use must be presented to the Estonian Commission on Mineral Resources for approval.

Despite some progress in the field of legislation and management, there is still a number of shortcomings in exploitation of mineral resources, different in reasons and possibilities of their avoidance, minimization, compensation or resonable alternation (Teedumäe 1990, National... 1992, Пязок 1990, Reinsalu *et al.* 1992, Toomik & Kaljuvee 1994, Raukas 1996). For instance, the mining methods in use do not ensure the complete extraction of oil shale bed (seams A-F) or the stability of surface topography. The associated minerals of overburden (limestone, peat, sand) and processing wastes (ash, limestone) have found use in small amounts, in cases not at all. Waste depositories of oil shale industry, covering thousands of hectares, contain a number of hazardous compounds, easily washed out by precipitation and polluting the environment. The resources of limestone, sand and gravel have not found the most efficient end use because of the low-standard mining and processing technology.

The speed of reaching the higher standards to a great degree is determined by economic and technological feasibilities, though reconstruction of the out-of-date industrial enterprises does not solve environmental problems. This especially concerns the activities aimed at reducing the damage made in the period of Soviet occupation. The impact of mining industry on the environment has already essentially disturbed the balance of the natural processes and structures. The experience obtained from north-eastern Estonia indicates at the ongoing unpredictable changes of topography and water regime causing remarkable damage to agriculture, forestry, waterworks and other fields of human activities concerned. To find out the limits of impact guaranteeing the reversability of processes after its cessation, *i.e.* to determine the nature stability level, is today the problem of the utmost responsibility. Though it makes the development of the mining industry greatly dependent upon the efficiency of measures available to environmental quality conservation, it gives the only chance to prevent unpredictable and irreversable damages. Today the main goal of minerals policy of Estonia is providing necessary mineral resources for long-term and balanced development of the national economy. The priority objective in this sense is the sparing and environmentally acceptable utilization of available resources for the most efficient end use concerning particularly the power, chemical and construction material industries. The means of managing the rational utilization of mineral wealth besides substantial activities of institutions in response, include promoting of environmental awareness, improvement of scientific and educational activities and stimulating of public participation. In these fields Estonia has long traditions and experience owing to the Estonian Society of Naturalists, the Estonian Nature Conservation Society, the Estonian Academy of Sciences, the Tartu University, *etc.*

XI GEOLOGICAL MONUMENTS

BOOK OF PRIMEVAL NATURE

Protection of nature monuments has long traditions in Estonia. In 1937, besides rare plant and animal species also several rarities of inanimate nature were taken under protection. During the course of past decades, nature monuments and landscape complexes have been increasingly endangered by ever expanding human activities, including the mining of mineral resources, construction works, *etc.* In earlier years, the nature conservation management in Estonia was unsystematic and in many respects insufficient, particularly as far as large landscape forms and groundwater were concerned. With this in view, an idea was suggested to carry out a large-scale revision of unique inanimate nature objects. The aim was to document these objects, to fix their present state and to assess to what extent their preservation is endangered. On the one hand, these plans addressed elaboration of scientifically grounded recommendations needed in arranging the protection of valuable geological monuments at state level. On the other hand, the aim was to notify both specialists and public about the actual number and value of natural rarities in need of protection.

The idea was advanced and unrestingly propagated by Herbert Viiding (1929-88). With the support of the Nature Conservation Commission of the Estonian Academy of Sciences, he reached the goal in 1982 when the order of the Estonian Government validated compilation of the Book of Primeval Nature. In the strict sense of word, this is not an ordinary book, although the final aim of the researchers is to publish a generalizing survey of the results obtained during the labour- and time-consuming work. First of all, this is a data bank registering in Estonia all noteworthy monuments of inanimate nature, each of these objects being provided with a special identification card, filled in by hand. Thereafter the data will be inserted into the computer. It was decided to register the rarities by counties. The work is coordinated by the Institute of Geology and financed by the Ministry of the Environment of Estonia. After H. Viiding's death, the work was continued under the leadership of Ülo Heinsalu (1928-94). Since 1994, Enn Pirrus has been at the head of the undertaking. Preliminary work was carried out in 1982-88. During 1990-95, 6000 pages assembled in 21 volumes and dealing with the inanimate nature monuments of 9 counties, were completed (Fig. 242). Considering the possibilities and amount of the work to be done, the book will be finally completed during 2002-2004. Thereafter the compilers will start data

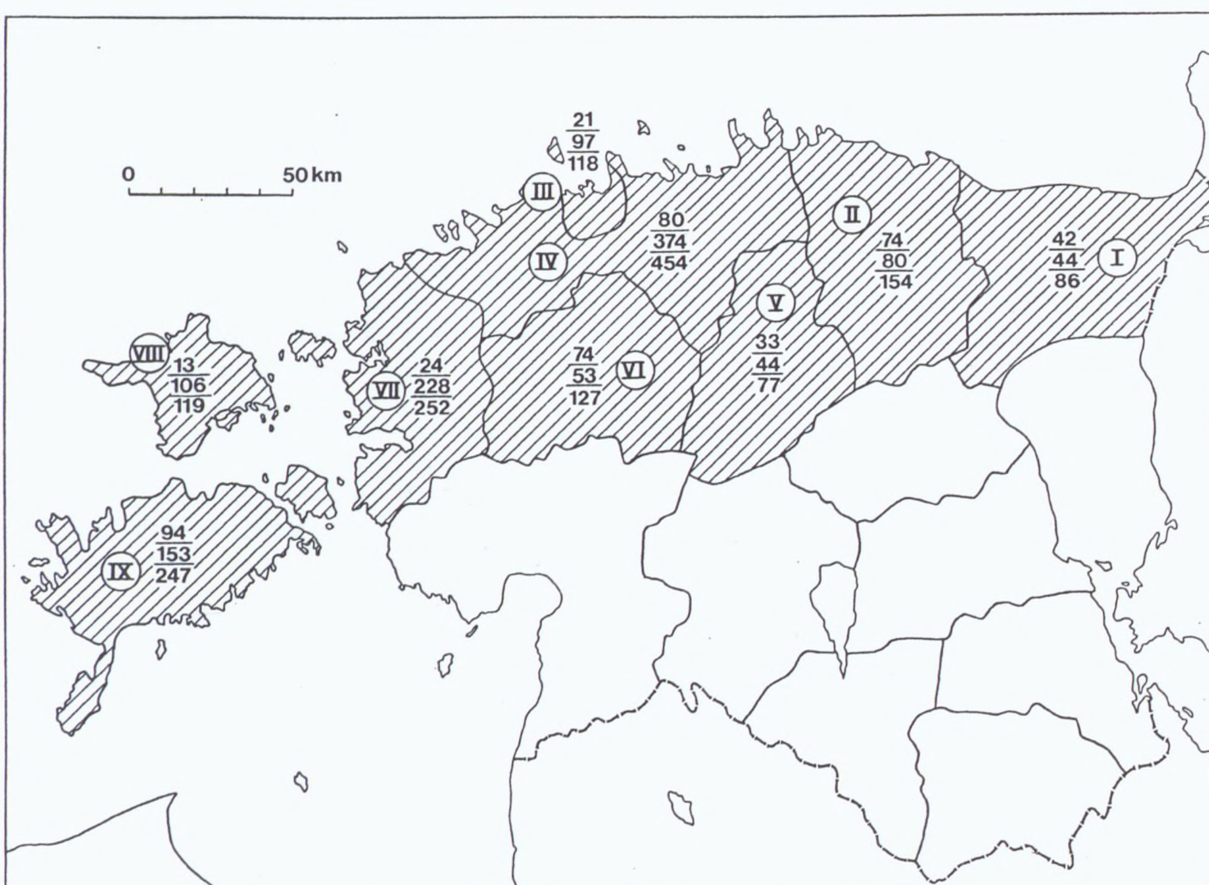


Figure 242. Compilation of the Book of Primeval Nature as of January 1, 1997. In the shaded areas the studies of the first stage have been completed. Figures mark: total number of recorded inanimate nature monuments (lower row), the number of big erratic boulders and boulder fields (middle row), the number of other nature monuments (upper row): I - Ida-Viru County; II - Lääne-Viru County; III - Tallinn; IV - Harju County; V - Järva County; VI - Rapla County; VII - Lääne County; VIII - Hiiu County; IX - Saare County.

processing, generalization, supplementation and revision with an aim of finding out the drawbacks to be liquidated. The results will be summarized and published. With the compiled data bank on protected inanimate nature monuments, Estonia will undoubtedly set a good example in the world.

On the basis of the initial classification (Viiding 1985), 20 groups of inanimate nature rarities were distinguished: meteorite craters, sections of the North-Estonian Klint, rafts, big erratic boulders, boulder fields, drumlins, eskers, kames, morainic hills, coastal formations, dunes, river valleys, waterfalls, karst phenomena, caves, springs, stratotypes, outcrops with rare fossils or minerals, tectonic disturbances and alvars.

However, during the compilation of the first volumes it became evident that this classification needs supplementation or revision. Under certain circumstances, several groups, *e.g.* karst and caves, karst and springs, valleys and waterfalls, may combine and their differentiation as single elements won't be purposeful. Additional rubrics should be introduced for several nature rarities in need of protection (curative mud deposits, geologically peculiar mire systems, wetlands around springs, some lake types, *etc.*). Differentiation of alvars aroused suspicion since these are geobotanical landscape elements rather than geological objects. Too few objects have been differentiated in the group of tectonic disturbances. Thus, the classification of objects is preliminary, disputable and will be finalized not until the final stage of data bank compilation. However, even in its present form it shows how wide is the spectrum of inanimate nature rarities resulting from the geological structure of Estonia.

An important criterion underlying the selection of objects for registration is their uniqueness. A decision was passed to include all meteorite craters, big erratic boulders and type sections in the list. From other groups, the objects are selected on the basis of expert opinions given by single specialists or teams. However, the criteria underlying the assessment of nature monuments may differ with groups and the results won't be always objective enough as reflected by the material collected at different times.

In compiling the data bank, it was decided to divide the objects, according to their value, into four categories and mark each category with a distinctive colour (sheets or writings). According to this decision, the red colour is used for extraordinarily valuable or unique rarities of inanimate nature, the destruction of which would be an irretrievable loss, *e.g.* exposed meteorite craters. For the preservation of such rarities finances are needed. The yellow colour marks more common nature monuments, endangered by human activities and in need of protection through special regimes which would be established considering the nature of these objects. The green colour is applied to the objects not in the immediate peril of destruction, like the buried valleys, the buried meteorite craters at Kärdla and Neugrund, most of tectonic disturbances, *etc.* The nature monuments, heavily damaged or ruined by human activities, are marked in black. This kind of subdivision (Viiding 1985) is interesting and expressive, but it can be realised only in the final stage of the work or even in publishing the generalised results, because the availability of the entire data bank is a prerequisite to the compilation of such classification.

Alongside the criteria of classification, the regulations for registration and study of inanimate nature monuments were elaborated. However, in view of the unique character of objects, it was possible to unify only the measurement of big erratic boulders. The corresponding instructions were published (Viiding 1987).

The general order of the registration of nature monuments on identification cards was established. It was agreed that the first page would contain the name of the nature monument (also synonyms), the group of classification, the description of its location and short characteristics. The second page would fix its value and state of preservation, the recommendations for protection, the references to the hitherto studies, the literature (also photos) or archival material available on the object. The page would end up with the name of the person responsible for the last revision, and the date of revision. The third page would present the scheme of the site, which is usually an excerpt from a large-scale topographic map showing the location of the object in the landscape. On the bottom of the page there are figures or other graphic presentations characterising the object. Thus, the identification cards present minimum data on the objects. These data are easy to compare and computerize. Other supplementary materials are stored in the working archives of compilers (currently in the Institute of Geology).

Table 75 gives a survey of the inanimate nature monuments by counties covered by the hitherto completed volumes. Some 35% of Estonia's territory, including the Pärnu, Viljandi, Valga, Võru, Põlva, Jõgeva and Tartu counties, still remains to be studied. Approximate estimation shows that there are some 550, together with big erratic boulders about 1800, inanimate nature rarities in need of protection in Estonia.

The Book of Primeval Nature would provide an excellent basis for nature protection management at state and regional level. The resultant measures would assure the preservation of Estonia's varied and in many respects unique natural environment and, naturally, a field of work for scientists. As is seen, the distribution of nature rarities differs with counties, reflecting the diversity and peculiar character of Estonia's nature. These differences are due to several factors. The substitution of the limestone plateau with the sandstone bedrock towards the south causes changes not only in the distribution of karst and springs, but also in almost all groups of nature monuments (waterfalls, glaciogenic landforms, *etc*). An important factor is the sea with its coastal forms and dune ridges and, as far as big erratic boulders are concerned, also the distance of the territory from the basement source area.

Sometimes, the Book of Primeval Nature has been erroneously classified as a register of protected inanimate nature monuments. Actually, this large-scale preliminary work is aimed at providing a basis for the development of argumented protection regime, the implementation of which would be decided on governmental level. Above all, it is a survey of the country's greatest natural treasures. The compilers of the data bank are all agreed that better than prohibitive measures or signs the growth of awareness and responsible cognition of the surrounding environment will lead to nature sparing activities which is the main goal of the Book of Primeval Nature.

Table 75. Main types of inanimate nature monuments in the Book of Primeval Nature

County	Landforms													Complex forms	Outcrops		Erratics		Other specific objects
	Glaciogenic					Coastal forms		Fluvial forms			Groundwater forms								
	End moraines	Eskers	Drumlins	Kames	Elements of hilly topography	Beach ridges and scarps	Dunes	Valleys	Waterfalls	Lakes	Karst forms	Caves	Springs		Cliffs and outcrops	Stratotypes	Big boulders	Boulder fields	
Ida-Viru	2	1 (3)*	-	(1)	-	(1)	3 (4)	1	3(4)	(3)	2 (4)	(1)	1 (4)	7	22	15	43	1	(2)
Lääne-Viru	2	4 (10)	(3)	1 (3)	-	-	-	3	2	(4)	12	1	31	7	11	11	79	1	-
Tallinn	-	-	-	-	-	2	-	1	-	-	-	-	5	-	13	3	91	6	-
Harju	-	2 (5)	-	2 (5)	-	4 (7)	1 (3)	2	9	1 (4)	15	5	11	6	22	10	372	2	(1)
Järva	-	1	2	-	-	-	-	-	-	-	7	-	21	-	2	1	44	-	-
Rapla	-	2 (3)	1	1 (2)	2	2	-	-	-	2 (4)	21	-	16	3	11	4	51	2	13 (15)
Lääne	-	1	-	-	-	3	1	-	-	2	2	-	4	-	9	3	219	9	2
Hiiu	-	-	-	-	-	2	1	1	-	-	4	-	3	-	1	-	104	2	1
Saare	-	-	-	-	-	5 (7)	2 (1)	-	-	-	23	-	18	2	43	21	152	1	1

* Figures in brackets show single landforms in complex units.

METEORITE CRATERS

The meteorite craters, registered in Estonia, include the Kaali and Ilumetsa crater groups and three single craters making up a total of 14 depressions (Fig.243) caused by cosmic objects which, maintaining their high residual energy, penetrated into the ground. These are mostly small impact craters from 9.0 (Simuna) to 110 m (Kaali main crater) in diameter (Fig. 244). As an exception serves the KärdlaCrater which is nearly 4 km long (Puura & Suuroja 1984). The crater groups at Kaali and Ilumetsa and the crater at Tsõõrikmäe are under state protection.

The group of craters at Kaali on Saaremaa Island is the largest and studied in particular detail mainly by Ivan Reinwald and Ago Aaloe. It has also attracted a great number of foreign scientists. Particular interest lies in geophysical studies, especially in the destruction zones of the bedrock in the area of the Kaali craters. The data obtained have been summarised in several booklets (Aaloe 1965, 1968, Tiirmaa 1984) and in a monograph (Tiirmaa 1994). The Tsõõrikmägi Crater with a diameter of 40 m near the Town of Räpina is reasonably well studied (Pirrus & Tiirmaa 1984). The group of craters at Ilumetsa has been studied to lesser extent (Aaloe 1979). The application of up-to-date study methods is complicated due to the presence of soft Quaternary sediments.

In 1984, a meteorite crater was discovered at Simuna (Pirrus & Tiirmaa 1984). It was evidently caused by pieces of a bright bolite which in 1937 disintegrated in the atmosphere above Virumaa (Kipper 1937). The study of the crater is under way.

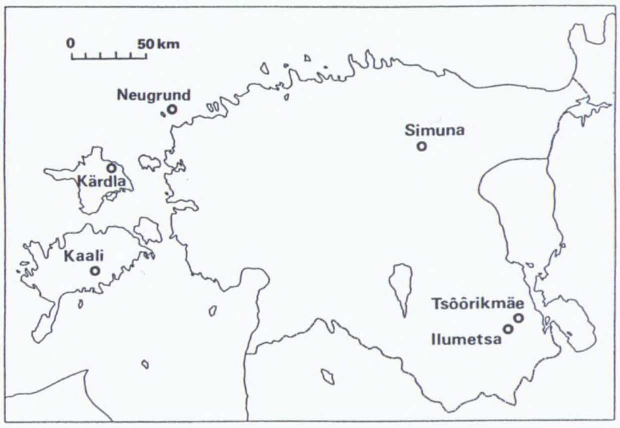


Fig. 243. The scheme of the distribution of meteorite craters in Estonia.

Kaali craters

Until the 1960s, the craters at Kaali on Saaremaa Island were the only known meteorite craters in Europe which had attracted scientists since the first half of the 19th century. The first description of the Kaali main crater appeared in 1827 in J.W.von Luce's book dealing with the nature and history of Saaremaa (Luce 1927).

Different hypotheses about the origin of the craters were advanced between 1827-1928, some of which suggested also their volcanic origin or the emption of gas and steam (Hofman 1837, Teichert 1927b). Eichwald (1843), Schmidt (1858) and Kraus (Kraus *et al.* 1928), who were well acquainted with the geology of Estonia, considered the dislocation zones and karst phenomena most important. In 1854, Eichwald suggested that there had been an ancient stronghold, in which a natural karst lake with man-made walls served as a well.

In 1927, I. Reinwald carried out geological investigations in the craters. In 1937, he collected 30 fragments of meteoritic iron from craters 2 and 5. The chemical analysis showed, that in these pieces Fe and Ni made up 91.5 and 8.3%, respectively. Minerals, characteristic of iron meteorites, were also found (Spencer 1938). According to more recent determinations, the Kaali meteorite belongs to the class of coarse octahedrite (Buchwald 1975).

Up to now, 3.5 kg of meteoritic material has been collected from the craters; the largest piece weighed 28.4 g. In the late 1970s and early 1980s, the types and distribution of the pulverized meteoritic matter were investigated and their preliminary classification compiled (Аалоэ и Тийрмаа 1981, 1982, Тийрмаа 1988, Shymanovich *et al.* 1993).

In 1955, Ago Aaloe (1927-81) proceeded with the studies started by Reinwald. During the course of the succeeding 25 years, he devoted to this work, the geological structure of the craters was studied in particular detail. In 1959, a geological protection area of craters was founded at Kaali and an exhibition pavilion was built near the main crater. In 1984, a memorial stone was opened to Ivan Reinwald and Ago Aaloe.

The Kaali metorite craters, 9 in all, are located within one square kilometre. On the bottom of the main crater there is a natural body of water known as Lake Kaali (Photo 73). The diameter of the lake depends on the water level and ranges from 30 to 60 m. The depth of the lake is 1...6 m, the maximum thickness of lake sediments is 5.8 m. The smaller craters, locally known as dry lakes, are shallow hollows bordered in places with the remains of a low wall. The craters have formed in the clayey basal till and underlying thick microbedded Upper Silurian dolomites (Fig. 245, Aaloe

Photo 73. Kaali main crater. *Photo by R. Tiirmaa.*

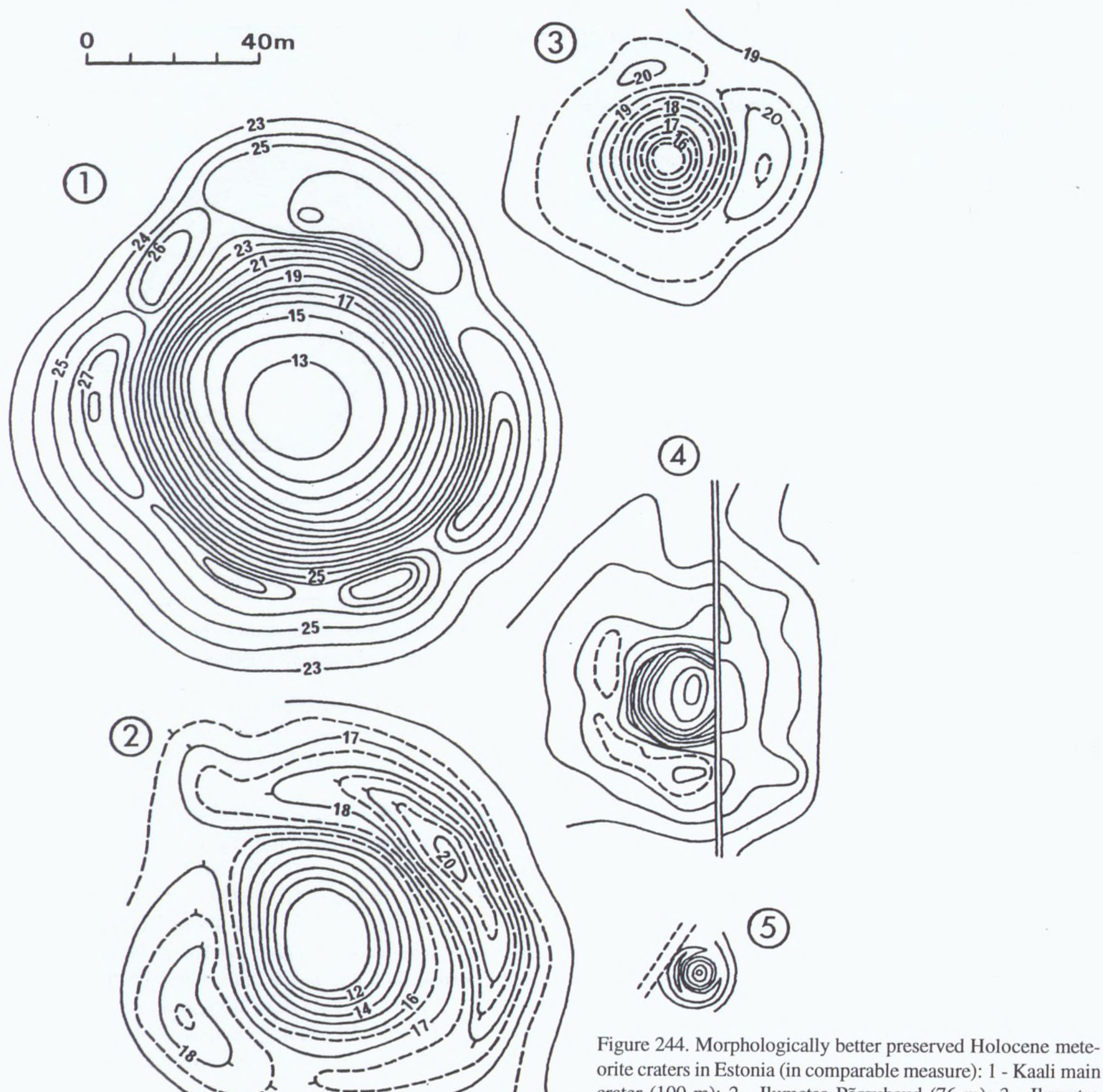


Figure 244. Morphologically better preserved Holocene meteorite craters in Estonia (in comparable measure): 1 - Kaali main crater (100 m); 2 - Ilumetsa Põrguhaud (76 m); 3 - Ilumetsa Sügavhaud (47 m); 4 - Tsõõrikmägi (40 m); 5 - Simuna (9 m, after Pirrus 1995).

1965). The main crater measures 105...110 m in diameter at the top of the mound, and is at least 22 m deep. The upper part of the mound consists of the material ejected from the crater during the explosion and of partly overhanging dolomite layers tilted at an angle of 25-90°. The uplifted bedrock complex with an average thickness of 10 m (Photo 74) has been split into nine shifted blocks, each up to 50 m wide (Fig. 246, Аалоэ 1963a).

The diameters of the secondary craters range from 12...40 m, and they are 1...4 m deep. On the bottom of craters 4 and 5 meteorite impact traces have been discovered.

The energy at the formation of the Kaali craters has been estimated at $4x10^{19}$ ergs for the main crater (Бронштейн и Станюкович 1963). Based on the main crater's energy of formation and the supposed angle of incidence of 45°, the following ranges of values were obtained: initial mass of meteorite 400 to 10,000 tonnes, mass at impact 20 to 80 tonnes, initial velocity upon entering the atmosphere 15 to 45 km/sec, velocity at impact 10 to 20 km/sec. The meteorite pieces causing the small secondary craters separated at an altitude of approximately 5...10 km, and their combined mass did not exceed 18 to 20% of the total mass.

Opinions differ as to the direction and angle of incidence of the Kaali meteorite. The sizes of the craters led the first investigators to believe that the direction of movement was from the southeast to the northwest (Reinwaldt 1937). Aaloe (Аалоэ 1958), basing primarily on the study of impact traces at the bottom of craters 4 and 5, maintained that the probable angle of incidence had been 35-40° relative to the horizon. The morphology of the wall of the main crater, the geophysical data available on the destruction zones of the main and secondary craters 1 and 6 (Аалоэ и др. 1982) and the distribution of dispersed material in the craters and outside the crater field (Аалоэ и Тийрмаа 1982, Тийрмаа 1988) suggest that the meteorite fell from the east-northeast.

Various methods have been used to determine the age of

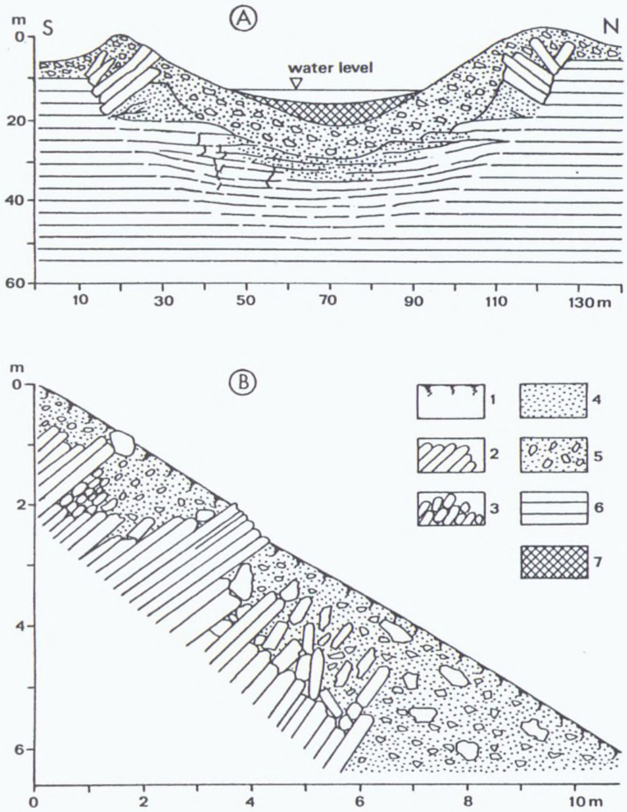


Figure 245. The geological cross-section of the Kaali main crater (A) and detailed geological cross-section of its south wall (B) after Aaloe (1963): 1 - soil; 2 - uplifted dolomite; 3 - shattered dolomite; 4 - dolomitic powder; 5 - filling-breccia; 6 - dolomite; 7 - gyttia and peat.

the craters. As neither deformed remnants from the explosion nor more recent sediments of marine origin had been found in the crater, Linstow (1919) estimated the age of the crater at about 4000 to 8000 years. Since the craters or their embankments did not reveal any traces of marine erosion or accumulation, Reinwald (Reinvaldt 1937), basing on the data available on the history of the Baltic Sea at that time, considered the craters some 4000-5000 years in age. In his first papers, Aaloe (Аалоэ 1958, 1963a) expressed the same opinion, but some years later he maintained that the age of the craters could not be more than 3000 to 4000 years. ^{14}C dating of charcoal discovered in 1961, yielded an age of 2660±200 years (Aaloe *et al.* 1963); the later dates 2530±130 and 2920±40 years allowed Aaloe to place the age of the craters erroneously at about 2800±100 years (Aaloe *et al.* 1975).

As it was not excluded that the dated charcoal was much younger than the craters themselves, great hopes were placed on drilling and dating of lake sediments in the main crater. Palynological analysis by Kessel (1981) showed that the bottommost sediments are Sub-Boreal in age and the craters are more than 3500 years old. Simultaneous ^{14}C and palynological investigations initiated by L. Saarse, placed the craters' age at approximately 3500...4000 years (Saarse *et al.* 1990) or 4000 years BP (Saarse *et al.* 1992).

Recent investigations have shown that the Kaali area was freed from the waters of the Baltic Sea already some 8000 yr BP. In 1994, a high concentration of microimpactites was detected in the peat of the Piila Bog, about 10 km to the northwest from the Kaali craters. The age of the layer with

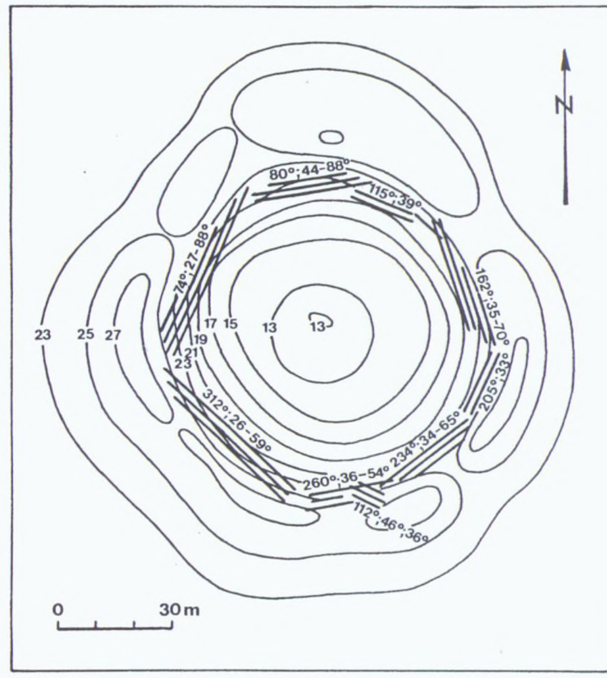


Figure 246. The scheme of dolomite layers crushed by the explosion in the rim of the Kaali main crater.

microimpactites was established by means of palynological and radiocarbon methods. The studies suggest that the Kaali craters were formed probably close to 7500 BP (Raukas *et al.* 1995a).

The mid-1970s witnessed an ever growing interest of historians in the Kaali Crater. Impetus was given by Lennart Meri's books "Hõbevalge" (1976) and "Hõbevalgem" (1984) and by the first archaeological finds in the east wall of the main crater in 1976. In 1978, excavations were begun on the discovered fortification which is located on the outside slope of the northeastern wall of the main crater. From the side of the lake it is protected by a steep slope and from outside by a semi-circular wall. Archaeological finds in the fortress area are limited, few earthenware fragments are dated from the

Photo 74. Dolomites uplifted and destroyed during the impact in the wall of the Kaali main crater. *Photo by R. Tiirmaa.*

7th century BC, most of the pottery dates from the Iron Age, the beginning of which is considered to be 600 years BC locally (Lõugas 1978,1980, 1995).

Ilumetsa craters

The meteorite craters south of the Ilumetsa station on the Tartu - Petseri railway (Fig. 247) were discovered in 1938 in the course of geological mapping. Artur Luha suggested their meteoritic origin. In 1956, A. Aaloe continued investigations. Of the five depressions registered at Ilumetsa (Aaloe 1979), only two – Põrguhaud and Sügavhaud – are definitely of meteoritic origin. Since 1964, they are under state protection.

In the crater area, the Middle Devonian bedrock consists of reddish or light-yellow weakly cemented silt- and sandstones of the Burtnieki Stage. These are overlain by a 1-2-m-thick layer of brown loamy till, which is in places covered by peat and glaciofluvial sands, up to some metres in thickness. The crater bottom reaches through loose Quaternary sediments into the Devonian sandstones (Aaloe 1979).

The diameter of Põrguhaud is 75...80 m at the top of the mound, its depth is 12.5 m (Photo 75). In the central part, the bottom is covered by an up-to-2.5-m-thick peat lense, the palynological and radiocarbon analysis (Liiva *et al.* 1979) of which gave some 6000 years for the age of the crater. The height of the surrounding mound is about 1 m in the west and up to 4.5 m high in the east. The Devonian sandstones below the crater bottom show the traces of the impact to a depth of about 30 m. In the main part of the mound, the bedrock layers have been lifted to plane fold. The crater is filled with a 10-m-thick mixture of sandstone and basal till, known as allothigenic breccia (Fig. 247).

Sügavhaud is located about 900 m to the south from Põrguhaud. The average diameter of the crater at the top of the mound is 50 m (Fig. 244), and the depth is 4.5 m. At the western margin of the crater, the mound is hardly traceable but in the east its height reaches 1.5 m. The destruction zone of the bedrock is about 20 m. The high eastern part of the mound in both craters suggests that the meteorite fell from the east.

Tsõõrikmägi Crater

The oldest meteorite crater of Quaternary age in Estonia is located at Tsõõrikmägi (*ring hill* in Estonian) in the south-

Photo 75. The meteorite crater Põrguhaud at Ilumetsa reaches through loose Quaternary sediments into the Devonian sandstones. *Photo by R. Tiirmaa.*

eastern part of the republic (Fig. 248, Pirrus & Tiirmaa 1984). Its mound is flattened but the structure, as a whole, is well preserved. The diameter of the crater at the top of the mound is 38...40 m, its depth from the highest point of the mound to the mineral-filled bottom is 5.5 m. The crater is located in the reddish-brown basal till, not reaching the bedrock. At the western margin of the crater, the till is covered with a thin (0.5 m) layer of varved clays, the texture of which shows foldings near the crater. The entire crater area is covered with a 0.3...1.2-m- thick layer of aeolian fine sand. The peat layer in the middle of the depression is 4.5 m thick and, according to the palynological and ^{14}C data, its formation started 9500...10,000 years ago (Pirrus & Tiirmaa 1984). Meteoritic material has not been detected, but the geological data available refer to the meteoritic origin of the crater. In 1984-85, Ülo Kestlane discovered and described the characteristic impact deformation in the varved clays.

Simuna Crater

There are data showing that on June 1, 1937, a cosmic body entered the atmosphere of the Earth. It moved from the east-northeast to the southwest-west under an azimuth of 259° and fell to the Earth under an angle of about 60°. Having reached thick air layers, it exploded at an height of 28 km somewhat to the east from the Viru-Roela Settlement (Kipper 1937).

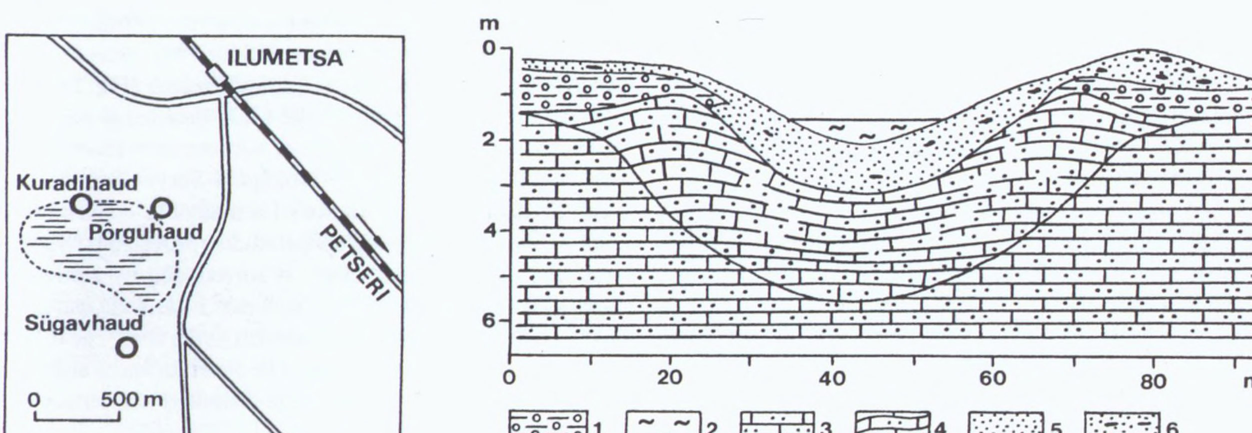


Figure 247. Location of Ilumetsa craters and the cross-section of the Põrguhaud Crater after Aaloe (1979): 1 - till; 2 - peat and sapropel; 3 - Devonian sandstone; 4 - disturbed sandstone; 5 - glaciofluvial sand; 6 - the same with till lenses.

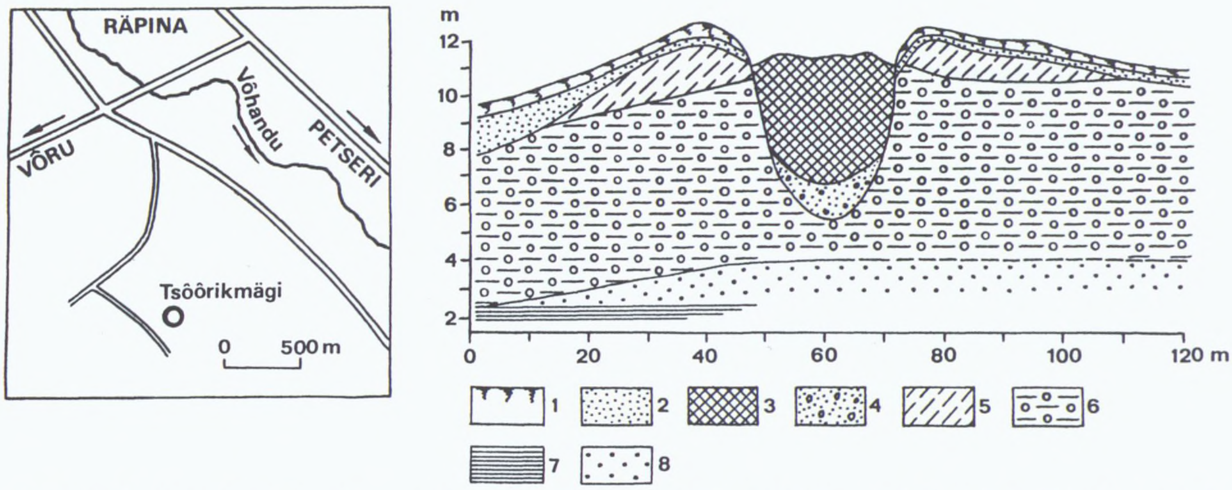


Figure 248. Location and geological cross-section of the Tsõõrikmägi Crater (Pirrus & Tiirmaa 1984): 1 - soil; 2 - sand; 3 - peat; 4 - silty filling material; 5 - till-like crater-wall sediments; 6 - till; 7 - Devonian clay; 8 - Devonian sandstone.

In 1984, the local resident H. Ross informed the Meteoritic Commission of the Estonian Academy of Sciences about a peculiar depression in the vicinity of Orguse Village near Simuna. Although the neighbourhood of Simuna is rich in karst forms, Ü. Heinsalu maintained that it was not a karst funnel. The excavations in 1986 showed that the depression was definitely not a trace of the last war, but probably a trace of impact. The highest part of the mound is in the west which suggests the fall of the crater-forming body from the east.

The diameter of the crater at the top of the mound is 8.5 m and its depth is 1.9 m. The mound around the depression is low (20...25 cm), inconspicuous, but clearly ring-shaped and continuous (Fig.244). It is underlain by a pre-event non-decomposed soil layer, covered by the material ejected as a result of the impact (Pirrus & Tiirmaa 1991). The depression was formed at the site with two distinct layers of sediments – 1.1 m of loose sand on top and clayey till below. In the crater, the sediments occur in reverse order: the bigger and heavier particles which were ejected into the air and are lying now on the soil, fell to the Earth's surface more rapidly than the fine material (Pirrus 1995c).

The crater has been used as a waste depository. For this reason it is not possible to use a magnet or mine detector in searching for meteoritic material, which has not been found so far.

Kärdla Crater

A buried impact crater is located near the Town of Kärdla on Hiiumaa Island (Fig. 243). This crater-shaped structure with a diameter of approximately 4 km and a depth of 540 m was discovered in 1967 (Viiding *et al.* 1969). The subsurface structure of the crater has been studied by gravity anomaly measurements, aeromagnetics and by means of more than 300 boreholes drilled in the crater, on the rim and in the surrounding area. The astrobleme origin of the Kärdla structure was established later (Кала и др. 1984б, Puura & Suuroja 1984, Пуура и др. 1989, Puura *et al.* 1994). The crater has a 110-m-high central peak (Puura & Suuroja 1992). Since 1983, mineral water has been derived by means of a well in the central part of the crater and bottled (trademark "Kärdla") for drinking.

The crater is only partly visible in the present topography. It is filled with Palaeozoic and Quaternary deposits and surrounded by a low ridge along the ring wall (Fig. 249). The crater was formed approximately 455 million years ago in composite stratigraphy: Middle and Lower Ordovician (20 m) and Cambrian (120 m) sedimentary rocks overlying the crystalline basement on the bottom of a shallow nearshore sea. The subsurface structure of the crater is well preserved. The vertical section comprises from top downwards (Puura & Suuroja 1992):

1) post-impact cover of Ordovician sedimentary rocks, 15-100 m thick;
2) allochthonous breccias (filling the lower part of the crater) and beds of fall-out breccias and conglomerates, sandstones and sandy limestones consisting of debris of reworked fall-out breccia and wall rocks surrounding the crater;
3) a body of autochthonous and subautochthonous breccias forming the bottom and the central peak of the crater and also remnants of its rim. Shocked rocks and minerals from autochthonous and allochthonous breccias have been revealed.

The crater became a rather good trap for different kinds of epizonal mineralization. In the outer slopes of the rim, lead and zinc sulphide (Жуков и др. 1987) and bitumen (Каттай и др. 1990б) occurrences have been found. The crater has not been active during the post-impact history (Puura *et al.* 1994).

Neugrund Crater

The Neugrund Bank, a shoal with a very peculiar shape, is situated between Osmussaar and Suur-Pakri islands to the south from the entrance to the Gulf of Finland (Fig. 243). In 1996, during the integrated large-scale geological mapping of the coastal and offshore area of northwestern Estonia, the researchers of the Estonian Geological Survey presented a hypothesis about the existence of a meteorite crater in this region. The following geophysical studies proved this hypothesis. The partially filled crater is surrounded by a circular rimwall which is about 150 m high and 1 km wide and consists of disturbed crystalline basement at the foot. The diameter of the crater is about 7 km. The Vendian and Cambrian target rocks behind the rimwall are disturbed to a certain degree within some kilometres.

According to the preliminary estimations, the event probably took place in the beginning of the Middle Ordovician Kunda Age 474 million years ago.

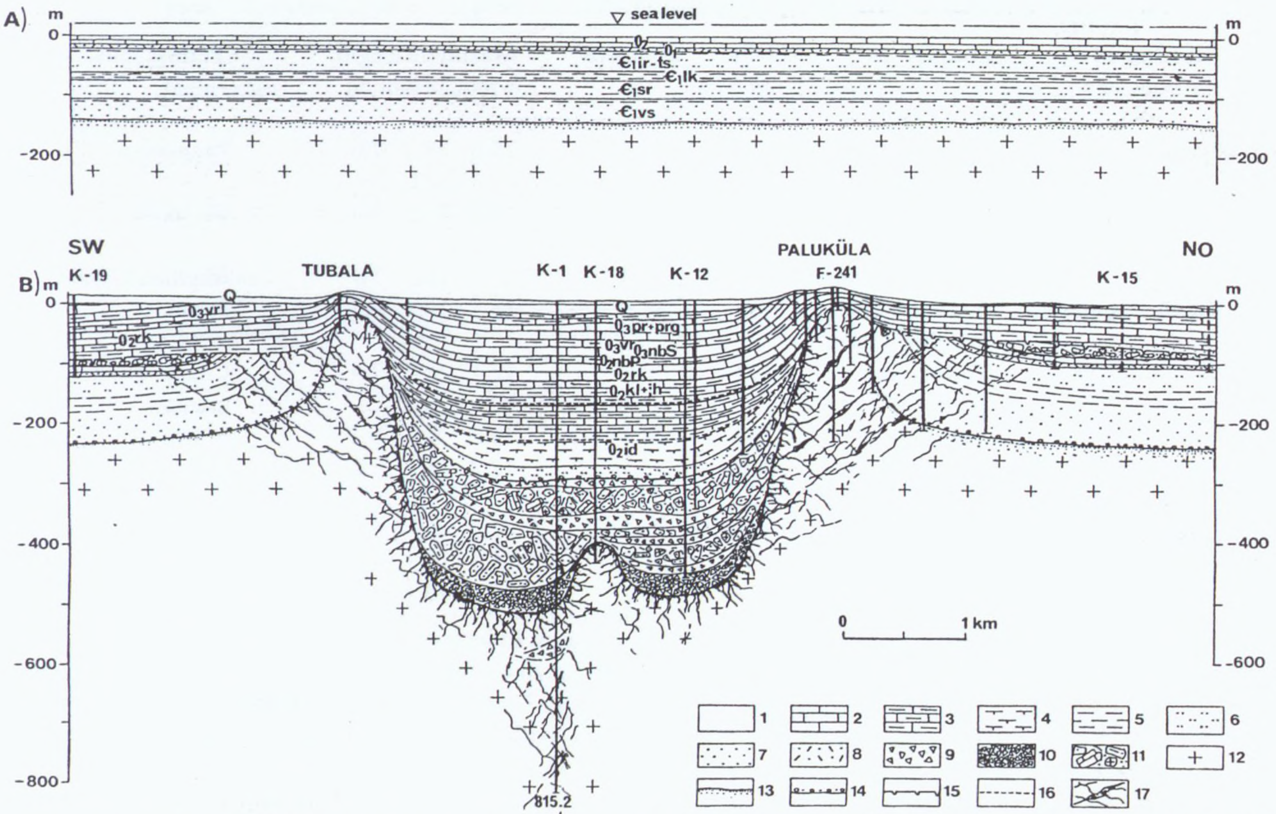


Figure 249. Cross-section of the Kärdla Crater: (a) reconstruction of the pre-impact stratigraphy; (b) present configuration (after Puura & Suuroja 1992): 1 - Quaternary deposits (Q); 2 - limestone; 3 - clayey limestone; 4 - marl; 5 - clay; 6 - silt and siltstone; 7 - sand and sandstone; 8 - sandstone with phosphate debris; 9 - 11 - breccias: 9 - containing mostly crystalline rocks, 10 - crater bottom breccia containing crystalline rocks, 11 - consisting mostly of sedimentary rocks with admixture of crystalline rocks; 12 - crystalline rocks; 13 - weathered crystalline rocks; 14 - basal conglomerate; 15 - main discontinuity surface; 16 - metabentonite layer; 17 - fractured rocks with breccia dikes.

ERRATIC BOULDERS

Owing to the closeness of the Fennoscandian Shield, Estonia's territory is scattered with numerous huge granite blocks of great height, and thousands of erratic boulders (Photo 76). Gigantic boulders inspired scientists in the last century with the idea of continental glaciation and served as the main evidence for the motivation of the corresponding theory (Raukas 1995a). Academician G. Helmersen (1869) being, on the one hand, impressed by the beauty and the majesty of gigantic boulders and, on the other hand, shocked by the ruthless destruction of several erratics, appealed to local landlords for the protection of boulders. His report in the Society of Naturalists at Tartu on February 17, 1879, was the first attempt in the former tsarist Russia to protect objects of inanimate nature.

Most of the gigantic boulders with a circumference above 30 metres or length more than 10 m in North Europe occur within Estonia where their number reaches 62. The largest boulders are listed in Table 76.

Currently, 192 big boulders, 18 boulder accumulations and 10 boulder fields are under the protection of the government (Fig. 250). Many boulders as historic monuments are protected by local authorities. Of the 2150 boulders registered in Estonia, some 1900 have a length of more than 3 m (Viiding 1987). However, every year in Estonia tens of new big boulders are documented, mainly during the studies conducted in the frames of the compilation of the overview "Book of Primeval Nature" ("Ürglooduse Raamat") under the leadership of Enn Pirrus (1995b).

Erratic boulders in Estonia were first used in the Mesolithic (7500-3000 BC) when prehistoric men learned to make stone implements. Initially, diabases, gabro-diabases and uralite-porphyrites were used as tools. Later they were applied to building of strongholds and churches. Foundations, walls and stairs of buildings made of erratics, stone fences around fields, and cobble-stone pavements in old towns are an inseparable

Photo 76. A lot of big boulders are scattered all over Estonia. Piretikivi on the Island of Saaremaa. *Photo by Ago Aaloe.*

Table 76. The biggest boulders in Estonia after Kumari (1979)

No.	The name of the boulder	Locality	Volume, m^3	Perimeter, m	Maximum length, width, height, m	Petrographic type
1.	Ehalkivi	Letipea beach, near Kunda	930	49.6	16.6 14.3 7.6	Pegmatite
2.	Muuga Kabelikivi	Muuga, near Tallinn	728	58.0	19.3 14.9 6.4	Rapakivi
3.	Majakivi	Juminda Peninsula	584	32.0	15.1 11.0 7.0	Migmatite-granite
4.	Vaindloo hiidrahn	In the sea east of Vaindloo Island	480	38.6	15.3 10.1 7.7	Rapakivi
5.	Maisiniidi kivi	Viimsi, near Tallinn	397	30.0	12.1 9.7 7.4	Coarse-grained granite
6.	Hellamaa kivi	Raasiku	360	34.4	14.2 8.4 6.2	Rapakivi
7.	Painuva kivi	In the sea near Turbuneeme	340	34.1	12.2 11.1 5.2	Rapakivi
8.	Ellandvahe kivi	Near Jõelähtme	332	31.3	12.0 8.9 5.9	Rapakivi
9.	Kukkakivi	Hiiumaa Island	324	41.5	16.0 11.3 3.9	Granite
10.	Laulumäe kivi	Near Jõelähtme	317	43.2	15.9 13.1 5.0	Rapakivi

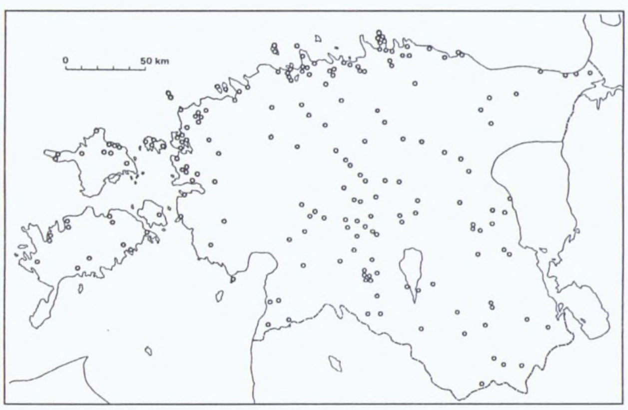


Fig. 250. Big boulders, boulder accumulations and boulder fields currently under the state nature protection (after Viiding 1960 with some modifications).

part of the present-day Estonia, and a source of inspiration for painters, sculptors and visitors coming from the countries where this kind of raw material is rare or absent. To our forfathers erratics were enemies, the frequent occurrence of which was a hindrance to cultivation, and against whom they kept a lifelong and, in some respects, hopeless fight.

There are many old legends and folk-tales connected with boulders. The origin of many erratics has been related with the heroic deeds of the Estonian national hero Kalevipoeg (Viiding 1976b). Several boulders were held sacred by ancient Estonians, they served as an object of worship and place of sacrifice. Often there are numerous hollows on the surface of such stones. A lot of sacred stones are under the state protection, some of those being rather big as, for example, an erratic at Ruila (circumference 34.5 m) in the Harju County and another at Mäkaste (25.3 m) in the Tartu County. As of January 1, 1996, there were 1402 small-creviced cult stones, 177 sacrificial stones and 15 stones associated with folk stories in Estonia.

The accumulations of sharp-edged gigantic boulders of the same type suggest that extremely large blocks were dragged along by the glacier which during the drift or after the melting of the ice crushed into smaller blocks (Viiding 1981). One of such boulder accumulations, identified on the Island of Hiiumaa, was termed the "Boulders of Helmersen" after the famous geologist, who was the first to describe them (Helmersen 1869).

On the basis of mineral composition and texture, more than fifty types of boulders can be distinguished in Estonia. Granites (including rapakivi) in which big boulders account for about 90 % prevail. These are followed by gneisses (6%) and pegmatites (3%). The basic groups of boulders, more than 1 m in length, are listed in Table 77. Some of those have been studied in the area of initial location in clearly marked places. These are indicator (index) boulders which show precisely the direction of the ice movement and promote resolution of several stratigraphical and palaeogeographical problems (Raukas 1995b).

The Vyborg rapakivi (Fig. 251) and the Hoglandian

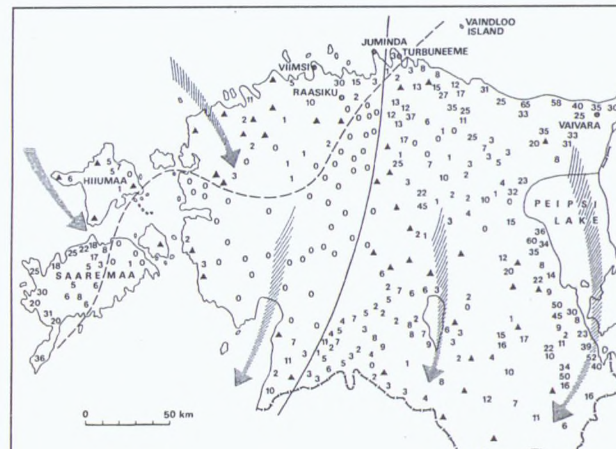


Fig. 251. Distribution of rapakivi (marked with triangles or in % of all counted boulders) after unpublished data of H. Viiding with author's complements. The continuous line denotes the westernmost boundary of the Vyborg rapakivi; the dashed line marks the glacial advance of the Palivere Stadial and the distribution of rapakivi from the Åland islands and from the massifs of southwestern Finland. The big arrows show the main direction of the ice movement (Raukas 1995b).

Table 77. Petrographic composition of Estonian boulders after Viiding (1955)

Main types	Average content, %			
	Estonia as a whole	The West-Estonian Archipelago	Estonian mainland Western part	Eastern part
Granitoids	82.2	75.0	78.5	86.7
Mafic rocks	3.5	4.0	4.0	2.5
Metamorphic rocks	12.9	18.9	16.0	9.7
Quartzites and sandstones	1.3	2.1	1.4	1.0
Carbonate rocks	0.1	-	0.1	0.1
Among granitoids: granites		49.2	60.2	57.5
porphyric granites		6.0	6.0	6.4
rapakivi		10.0	5.0	15.8
migmatites		5.8	5.3	3.7
pegmatites and aplites		4.0	2.0	3.2

(Suursaari) and Ålandian quartz-porphyries are the best indicator boulders in Estonia. However, the rapakivi from south-western Finland (Laitila and Vehmaa massifs), the Ålandian granites and rapakivi as well as the red and brown quartz-porphyries from the bottom of the Baltic Sea, the Satakunta olivine diabase, the Bothnian porphyries and the uralite-porphyrites of Tammela and Pellinki and several other types of rock can also serve for that purpose (Раукас 1963б). Each group of erratics includes several varietes. For instance, in the group of Vyborg rapakivi Viiding (1955) differentiated eight subtypes. Altogether some one hundred different indicator boulders have been described in the Baltic States (Гуделис 1971).

The distribution of indicator boulders in till beds and orientation of clasts in tills suggest that the movement of glaciers during the Sangaste (Oka, Elster 2) and Early Ugandi (Dnieper, Drenthe) glaciations in Estonia was mainly from north to south, but during the Late Ugandi (Moscow, Warthe) and Early-Middle Weichselian from the north-west to the south-east (Таваст и Раукас 1982). There were naturally different local movements depending upon the bedrock topography.

Of lithological methods, the research into crystalline indicator boulders shows the highest promise in solving the problems of Pleistocene stratigraphy and palaeogeography since the content of the boulders in deposits has been only slightly affected by differences in the composition of the local bedrock, remaining more or less stable over vast areas (Раукас 1978).

The resources of erratic boulders as a building material have rapidly decreased from *ca.*16.5 million cu m in 1945 (Luha 1946) to less than 2 million cu m in the early 1990s (Raudsep & Räägel 1993).

GLACIAL TOPOGRAPHY

The most outstanding features of the nowadays landscapes result from the geological structure (morphostructure) of the southern slope of the Fennoscandian Shield and the glacial modelling (morphosculpture) of recurrent glaciations.

As part of the East-European Plain, Estonia is characterised by a considerably flat surface with small relative and absolute heights. The uplands and plateau-like areas alternate with lowlands, depressions and large valley-like hollows. About half of Estonia's territory does not rise higher than 50 m above sea level. During the Pleistocene glaciations, masses of ice moved across the region scouring the bedrock and removing a layer of rocks tens of metres thick, levelling the topography and forming the complicated accumulative complexes. All varieties of glacial topography are represented in Estonia. The unique esker systems (Photo 4) and classic drumlin fields are the most attractive glacial landforms in northern and central Estonia. A picturesque hummocky glacial accumulative topography is characteristic of southern Estonia (Photo 46).

From the scientific point of view, glacial forms, such as

Photo 77. Both, Lake Uljaste and the picturesque esker are under protection. *Photo by Ago Aaloe.*

eskers, end moraines (Foto 45) and kame fields, ought to be protected to preserve our unique topography. Besides, topography is part of any protected landscape, an important prerequisite to its existence. With the intensification of nature use, the latter circumstance is gaining in importance.

In Estonia, glacial streams with different speeds and energies alternated regularly with ice-divide areas, forming lobe depressions, uplands in ice divide zones, interlobate and marginal complexes. Northern Estonia belongs to the glacial erosion zone and is characterised by levelled topography of till plains and glaciofluvial deltas (Photo 47) left behind by the Late Weichselian glaciation. Uplands have a bedrock core and a thin cover of Quaternary deposits. The most remarkable glacial zones are related to the ice-shed uplands, accumulative heights and interlobate and marginal formations. Lobe depressions and interlobate formations are mostly protected through conservation of small landforms, however, the protection efficiency is not so high as in the case of complex forms (Photo 77) or landscape reserves. Unfortunately, we have missed the opportunity to protect several glacial forms which have been destroyed by today.

The type of the glacial dynamics and subglacial relief determined the structure and morphology of glacial landforms. The most remarkable forms of the active glacier are drumlins, usually associated with the lobe depressions of Gotiglacial time. Radial eskers are typical to the later stages of the Gotiglacial and Finiglacial. The lobe depressions are usually bordered by end moraines, glaciofluvial deltas and interlobate complexes.

The circumstance that aqueoglacial and glacial landforms consist mainly of sand has proved fatal for their existence: most of the ice-marginal formations exists now only on historical maps. Even protected landforms (Pähnimäe esker near Rakvere, Uljaste esker in the Ida-Viru County and the Vooremägi hill in the Tartu County) have been partly destroyed by excavation. The protection of landforms consisting of till (drumlins, end-moraines) has been more effective.

The morphogenesis of the **ice-divide areas** is accumulative in the case of the convergent movement of ice and erosional in the case of its divergent movement. **Accumulative insular heights** formed mostly in the beginning of Gotiglacial time. The Haanja and Otepää heights are typical representatives of this category. They have a thick (160-180m) Quaternary cover, complicated structure, high absolute marks and distinct slopes. Due to the island-like position in the topography, they rise high above the surrounding lowlands. The foot elevation of the Otepää Heights is 90 m, the area about 1200 sq km and the highest point rises 217 m above sea level (Hang & Karukäpp 1979). The altitude of the Haanja Heights ranges from 140 to 210 m, the area within Estonia is 950 sq km (Arold 1993) and the highest point is 317.6 m.

Morphometrical studies and the analysis of hypsographic differences of the heights have revealed the vertical and concentric differentiation of the relief (Hang 1979, Hang & Karukäpp 1979). The highest level of glacial accumulation in the central part of the heights is characterised by large (higher than 25 m) hummocks with a top of glaciolacustrine clays. The central high relief is surrounded by a belt of medium-sized (10-25 m) and small moraine hummocks and kame fields. The lowest level of glacial accumulation in the marginal belt of the heights is mainly represented by small landforms, less than 10 m in height. The landform complexes, distinguished by morphology and genesis, show a correlation with the absolute height.

The most typical areas of high topography in the center of the Haanja and Otepää heights are protected. The Haanja Nature Park (9162 ha) includes landforms under protection (Suur Munamägi, Meegomägi and Vällamägi). The Otepää Landscape Reserve (23,031 ha) includes the highest typical part (20%) of the heights (Photo 46).

The Karula National Park (10,318 ha) occupies the hummocky area of the Karula Upland which is genetically connected with the Otepää Heights and has a unique topograhy of rounded cupolas and esker-kames.

Pandivere is a typical bedrock upland of the ice-divide area. The relative height of the bedrock core is about 60 m, the absolute height is 120-130 m. The foot level is 70-80 m above sea level and the area is 3180 sq km (Карукяпп 1978, 1987). The highest point Emumägi in the southeastern part of the upland is 166 m above sea level The glacial relief was studied in particular detail by Rähni (1961, Ряхни 1967 a.o.).

The glacial relief of the Pandivere Upland is represented by till plains, alvars, glaciolacustrine plains, glaciofluvial (delta) plains, hummocky morainic relief, undulating till plains, kame fields, valleys, eskers and drumlins. Flat and slightly undulating plains are prevailing. The hummocky landscape occupies only 13% of the upland (Карукяпп 1979). Eskers are the most remarkable landforms of Pandivere. The total length of the esker systems on the upland is 230 km. The typical esker systems (Neeruti, Mõdriku-Roela) are under protection as they are situated within the Pandivere Water Protection Area. Remarkable glacial complexes of interlobate formations belong to the Kõrvemaa Landscape Reserve (21,720 ha) and Lahemaa National Park. The biggest kame fields are located on the slopes of the upland.

The **local interlobate formations** are represented by hummocky morainic topography, kame fields and eskers. The complexes were formed during the final stage of the glacier activity between small-scale glacier lobes and tongues. Local interlobate and marginal complexes are well developed in the Pandivere stadial belt. In zones between the differently moving ice portions till was concentrated and systems of crevasses were formed. The latter served as the initial stages for the interlobate formations.

The concentration of glacial accumulation was realised on accumulative insular heights. The divergent ice movement on the bedrock uplands gave rise to the ice-shed socle uplands of glacial erosion. Intensive movement of glacial streams and lobes developed lobe depressions with more or less levelled surfaces. The high differentiation of the ice movement generated a great number of small glacial tongues during the later stages of deglaciation and created conditions for the formation of interlobate complexes. The same regularities of glacial morphogenesis can also be followed on the Estonian shelf (Карукяпп и Васильев 1992).

The number of different types of glacial morphogenesis has been recorded in the diversity of the Estonian landscapes. The most typical and unique of them must be effectively protected.

VALLEYS

In 1957, the Government of the former Estonian SSR adopted several laws and regulations addressing the organization of nature conservation in Estonia. As a result, nature reserves were founded in the valleys of the Ahja, Pirita, Tilleoru and Valgejõe rivers (Kumari 1960). The Pühajõgi Valley was taken under protection as part of the Saka - Ontika - Toila Landscape Reserve. Somewhat later the ancient valleys of the Võhandu (1964) and Piusa (1965) rivers were also designated as landscape reserves. In the protected valleys any kind of human activity, such as dredging and damming of rivers, hunting, clear cutting of forest, *etc.* which could have caused changes in natural landscapes, was prohibited. In accordance with the nature protection system valid at that time, the Selja, Pada, Viljandi and several other river valleys were taken under the protection of local authorities. At present, reorganization of the nature protection is in progress. In 1994, the Law on Protecting Nature Objects and some other laws and regulations related to nature protection, were adopted.

The main principle underlying the establishment of the river valley protection regime has been its landscape value. To assess it, the valley as a specific landscape element must be studied with all its components, including geology, geomorphology, flora, fauna, *etc.* The river valleys in Estonia, though young and rather small, differ in their geological structure and origin (see River Activity). Detailed research into Estonian river valleys is of major significance in a number of respects. First of all, it will impart information on the geological evolution of Estonia's territory in the Late Quaternary and Holocene, which provides an excellent opportunity to study the beginning and development of fluvial processes in an area of sedimentary rocks freed from the continental ice. On the other hand, the knowledge of protected river valleys acquired at schools, higher educational establishments or through club activities will essentially contribute to the development of man's attitude to nature, evoking his respect for this heritage and promoting his understanding of the necessity of nature protection. Peculiarities of fluvial processes in lithologically different parent rocks, or in the conditions of changing base level or tectonic regime are better understood during field trips or while hiking along nature study trails. Many valleys, particularly in southern Estonia, are of great cultural value. The Ahja River Valley with the towering outcrops of Devonian red sandstones is a symbol of Estonian nature (Photo 25). The geologically and geomorphologically peculiar nature of Ahja, Võhandu and Piusa valleys has inspired a great many legends and folk stories (Varep 1965, Kumari 1972, Ivask & Hillep 1973, Heinsalu 1987). Deep valleys offered refuge in wartime. There were also picturesque sites where rituals could be performed or festivities arranged. And finally, one must not forget how powerful is the water in rivers with a great gradient. Many old milldams stand as witnesses to the progress of the technical thought in Estonia.

Below we shall deal in brief with the geological and geomorphological peculiarities of the protected river valleys with due regard for their evolution.

In northern Estonia, the valleys on the lower reaches of the Pirita and Valgejõgi rivers have been designated as protected areas. The valley reserve on the Pirita River has an area of 550 hectares and is a most typical North-Estonian valley. This up-to- 20-m-deep valley cuts in Ordovician and Cambrian rocks through the thin Quaternary cover which consists mostly of the sediments of the Baltic Sea and deposits of the last ice age. The stratotype of the Cambrian Lükati Formation is situated on the left steep bank of the valley (Photo 16). The river valley, V-shaped or with asymmetric cross-section has a channel abounding in rapids and boulders, and in cobble of magmatic and metamorphic rocks. In some stretches, the river gradient exceeds 9m/km (Künnapuu 1957, 1971). The stretch of the river valley within the reserve came into being during the Ancylus regression. The abundant terraces formed somewhat later - partly during the Litorina, but mostly during the Limnea regression (Künnapuu 1957, 1971, Мийдел 1967). The location of the Pirita Valley Landscape Reserve within the administrative borders of Tallinn has caused difficulties in following the regime established for nature reserves.

The Valgejõgi Valley Landscape Reserve, some 60 kilometres east of Tallinn, is the largest of the kind in Estonia. The reserve has an area of 665 ha The valley's greatest depth is 36 metres and width 650 metres. As far as the North-Estonian, Klint the river flows in a relatively narrow V-shaped valley which turns gradually deeper until breaking through the klint as a canyon, 20 m deep and half a kilometre long. The towering walls of the canyon expose Ordovician rocks from the Aseri to the Pakerort Stage. On the bottom of the canyon there is a 1.2-m-high Nõmmeveski Waterfall. Downstream of the klint, the river flows into the Valgejõgi Klint Bay filled with glaciofluvial and glaciolacustrine sediments. The wide river valley abounds in terraces which have formed as a result of the drop of the Baltic Sea level since Ancylus time (Linkrus 1963, Hang *et al.* 1964, Мийдел 1967). The Vasaristi Karst Field and the Vasaristi Brook (Photo 78) are situated within the reserve (Miidel 1962, Heinsalu 1977).

The valleys in the lower courses of the Selja and Pada rivers (area 1110 and 37 ha, respectively), so far protected by local authorities, deserve state protection. In the Selja River Valley the stretch from the Varangu Village up to the river

Photo 78. Breaking through the klint, the Vasaristi Brook forms a nice 3.6-m-high waterfall. *Photo by A. Miidel.*

mouth should be taken under protection. The Selja River breaks as a short V-shaped canyon through the escarpment and falls into the klint bay infilled with delta sediments of the Baltic Ice Lake. The stratotype site of the Varangu Stage is situated 300 metres downstream of an old milldam on the left bank of the valley From the klint the river continues seaward along a 200–300-m-wide and 8–10-m-deep valley, the left bank of which is cut by numerous gullies. The 4–5-km-long stretch of the valley near the river mouth is deep (occasionally more than 20 metres), V-shaped and full of rapids. The narrow bottom, less than 50 metres in width, has cut in ice-dammed lake sediments and till. Evidently, this is the only valley with ingrown meanders and asymmetrical cross-section in Estonia. The extremely high concentration of subfossil shells in river sediments is another characteristic feature of the valley in the lower course of the Selja River. The shells evidently originate from lake sediments spread in the upper and middle courses. Discontinuous terraces date from the post-Ancylus period (Мийдел 1967). The valley in the lower course of the Selja River (Photo 53) has a particular visual charm and is undoubtedly among the most picturesque valleys in northern Estonia.

The valley on the middle and upper reaches of the Pada River between the Samma Village and Kongla Brook runs straight, evidently due to the NNE oriented disturbance zones (Мийдел 1971). The extensive Aseri zone of disturbances, in which the valley cuts, has not affected its course. In the vicinity of the Samma Village, the river flows for a short distance underground, appears then to the surface and continues along the narrow (bottom depth often less than 100m), V-shaped valley which turns quickly deeper and attains its greatest depth (more than 20 m) in the vicinity of the Pada Village. The Linnamäe Brook flows into the valley from the right. Up to the Koila Village the valley is in Ordovician rocks. Downstream of the village it is in Quaternary sediments, however, near the lower terrace it again exposes Ordovician and terrigenous Cambrian rocks.

The above valleys were formed mainly in the Holocene, and represent thus the youngest link in the hydrographic network of Estonia. They developed when the area emerged from the waters of the Baltic Sea. The process was also controlled by the steadily decreasing rate of the uplift of the Earth's crust. Both these factors account for the steep gradient and deep incut in the lower courses of the rivers.

The rivers of southern Estonia basically have very old valleys the formation of which is related to the recession of the glacier in response to the lowering of water level in ice-dammed lakes in front of the ice margin. The valleys have cut into Devonian terrigenous rocks. Evidence is derived from the morphology of valleys, although also in southern Estonia V-shaped or flood-plain valleys dominate.

The Ahja River Valley is the best-known valley in southern Estonia. The reserve stretches at a length of some 18 kilometres and has an area of 1101 hectares. Due to the rather steep average stream gradient (up to 1.5 m/km) the river is full of rapids. The best-known rapid in the vicinity of Saesaare remained under water when a hydropower station was founded there in 1952. The valley cuts in the slope of the Otepää Heights to a depth of 32 metres. In this 200–300-m-wide valley terraces occur at eight levels.

The Ahja Valley owes its particular visual charm to numerous (altogether 32) towering multicoloured outcrops of Devonian sandstones of the Burtnieki Formation (Photo 25). The local people have their own names for each of these outcrops, known as "sky houses". There are many legends and folk tales connected with these formations (Kumari 1972). Well-known are Suur (height 24 m) and Väike Taevaskoda (Big and Small Sky House). At the points where water gushes out at the surface as a spring, caves have formed. Each cave has its own name and a legend to explain its origin (Heinsalu 1987).

In general lines, the valleys of the Võhandu and Piusa rivers are similar in structure. The Võhandu Landscape Reserve with an area of 750 hectares is situated on the margin of the South-East Estonian Plateau where the depth of the valley reaches 15–30 metres. The rather narrow (150-200 m) valley has terraces at three levels. Each of the 38 "walls", *i.e.* outcrops of sandstones of the Burtnieki Formation (Varep 1965), has its name given by the local people. Like in the Ahja Valley, the sandstones hold many caves some of which were formed by springs, while others came into being as a result of erosion (Heinsalu 1987).

The primeval valley of the Piusa River on the slope of the Haanja Heights exposes mostly sandstones of the Gauja Formation. The fall of the river within some 12 kilometres is more than 3 m. Due to the great fall, the river has cut deep in the Quaternary cover and the bedrock. Its greatest depth is 43 metres. The initially rather narrow (200-500 m) valley widens quickly and is soon a kilometre wide. Terraces occur at five levels (Liblik 1966). The steep slopes of the valley expose towering walls of Devonian sandstones and contain caves

(Heinsalu 1987). Härma, the most famous of the 14 big outcrops on the Piusa River, has a height of 30 metres (Ivask & Hillep 1973). The terraces, found in all three valleys, formed in response to the lowering of water level in the ice-dammed lake in the Peipsi Depression (Hang *et al.* 1964, Liblik 1966,).

KARST AND SPRINGS

Natural monuments associated with karst forms and springs (Fig. 252) are of great scientific and educational significance, but they are also important in terms of ground water development. In the karst areas, groundwater has no natural protection and can easily become polluted. The quality of water in a river depends on the abundance and quality of water in the spring where it starts from. Therefore, it is purposeful to restrict the use of water in the so-called water preservation areas which are actually reserves where man's economic activities with potential adverse impact on the water are limited (Raukas 1993).

The major karst phenomena and springs with scientific, cultural, educational and aesthetic value have been included in the Book of Primeval Nature and some of them have been taken under protection. For instance, of the 17 karst fields and 40 springs in Virumaa, 9 karst fields and 21 springs are under conservation.

Karst forms develop in easily soluble fissured and porous rocks under the effect of surface and groundwater. They most commonly occur on the outcrops of carbonate rocks in northern and central Estonia and, to a lesser extent, in the southeasternmost part of the republic. In the areas with a thin Quaternary cover, both surficial and subterranean karst phenomena are encountered. Karren, open cracks, sink holes, hollows and valleys (Photo 79) are observable on the surface The subterranean karst forms, sometimes reaching a depth of 50 m from the surface (Heinsalu 1977), include caves, cracks and underground rivers. Karst phenomena are most numerous on bedrock elevations with intensive water exchange (Jõhvi, Pandivere, Lääne-Saaremaa), on plateus in the areas of tectonic disturbances, bedrock elevations and draining landforms (klint, ancient valleys). In some places one may find peculiar temporary karst springs which function only during the spring high-water period.

There are six regions with prominent karst phenomena in Estonia (Fig. 252).

1. Within the Jõhvi Upland, 40 major karst phenomena are encountered. These are associated with the Ahtme fault zone and oil-shale mines (artificial karst). The largest karst field in the region is situated at Kalina on the southeastern boundary of the upland. The Ratva and Kiikla brooks disappear underground in the western part of the upland (15-20 l/sec). The water pumped out from the mines is absorbed in the Kohtla-Ojamaa ditch (Kink 1996).

2. The Pandivere Water Protection Area on the Pandivere Upland holds 352 karst forms or fields (Karst... 1994), of which 15 have been included in the Book of Primeval Nature. 14 karst regions serve as water preservation areas. The largest karst fields are Savalduma, Avispea-Triigi, Pandivere-Raeküla, Assamalla, Karitsa-Jupri, Liisingi, Sääsküla ja Kuksema. The largest groups of temporary karst lakelets occur at Võhmetu, Saksi, Aavere and Annisti. The largest underground streams flow in the Liisingu Karst Field and on the northern boundary of the Rakvere Town.

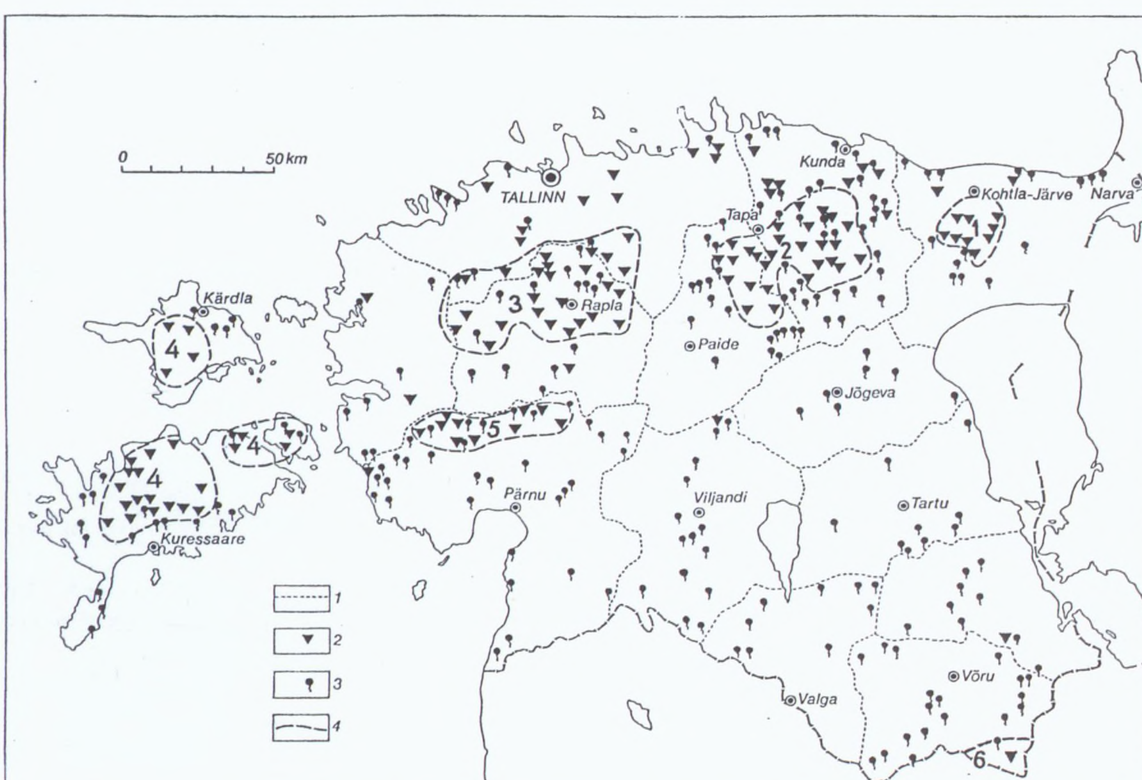


Figure 252. Karst forms and springs in Estonia: 1 - boundary of the county; 2 - karst; 3 - spring; 4 - karst region and its number.

Photo 79. Karst valley in the Kostivere Karst Field. *Photo by Ago Aaloe.*

3. The Kohila Karst Region in the North-Estonian watershed area comprises the karst fields which are situated in the southern part of Harjumaa and in the northern part of Raplamaa. 24 karst fields have been included into the Book of Primeval Nature. The area between Lelle, Kose and Märjamaa holds several hundred karst funnels and hollows on low bedrock elevations (Heinsalu & Vallner 1995). The largest karst fields are Kuivajõe, Kata, Pae and Palamulla.

4. On the islands of the West-Estonian Archipelago, three subregions have been differentiated. On Hiiumaa, karst forms are concentrated in the southern part of the island where the Quaternary cover is thicker. The established typical karst funnels and caves with the exposed bedrock are observable only at Kurisuu and Käina. Of the 38 karst fields on Saaremaa, 23 have been listed in the Book of Primeval Nature. Karst phenomena are encountered mostly in the center of the island on the West-Saaremaa bedrock elevation between Mustjala - Kuressaare - Valjala. On the western slope of the elevation, dunes and beach ridges obstruct the surface water flow. Deep (5-6 m) sink holes and canyons have formed above the subterranen tunnels at Kiidema, Poka, Lepakõrve, Selgase and in several other places. On the eastern slope, the karst forms are not so deep but larger in area, *e.g.* at Ohtja, Kuumi, Kärla, Hakjala, *etc*. The only group of temporary karst lakes is situated in the Kaarmise-Jõempa Karst Field.

Karst is related to local limestone hillocks in eastern Saaremaa and at Viira-Saikla, Parasmetsa and Muhu-Liiva on Muhu Island. Peculiar karst fields occur on Vilsandi Island and on Vaika islets (Photo 80).

5. In the northern part of Pärnumaa, karst phenomena are concentrated in the area of Koonga, Mihkli and Kibura. It is mostly surficial karst represented by cracks opening on the surface and in ditches, which shows that the karst forms are young and in the initial stage of development. About 10 karst fields of the area will be included in the Book of Primeval Nature.

6. In southeastern Estonia, major karst phenomena include karst hollows occupied by lakelets at Tsiitre, and the Meremäe sink hole.

On the North- and West-Estonian plateaus, karst phenomena occur within single but spacious karst fields, like Uhaku in Virumaa (Photo 5), Kostivere in Harjumaa (Photo 79) and Salajõe in Läänemaa.

Springs are the sites where groundwater appears on the surface. On the basis of groundwater outflow, descending and ascending springs are distinguished. In descending springs, the water flows out from the cracks of carbonate rocks or from between the layers. In ascending springs, hydrostatic pressure makes the water move upwards along the cracks. According to the capacity, very large (> 100 l/sec), large (100-10 l/sec), medium (10-1 l/sec) and small (< 1 l/sec)springs are distinguished (Heinsalu 1977). The outflowing groundwater originates from the Quaternary cover and the bedrock.

At the foot of the North-Estonian limestone escarpment the flow from the Cambrian-Ordovician aquifers forms single medium descending springs (at Toila, near the lighthouse on the Viimsi Peninsula) or a belt of small springs.

In the Ordovician-Silurian outcrop area, the springs are related to bedrock elevations, uplands or Quaternary relief forms (kames, eskers, beach ridges). The slopes of bedrock elevations may be buried or exposed. On buried slopes, small ascending springs and on exposed slopes medium descending springs are formed. Ascending springs may also open on lake bottom like, for instance, in the lakes of Äntu in the Pandivere Upland. The largest springs occur at the foot of the Jõhvi and Pandivere uplands at Simuna, Mõdriku, Lavi, Norra and Prandi. On the plateau, karst rivers appear at the surface as large springs at Sutlema, Tuhala, Kuivajõe, Jõelähtme, Erra, *etc*. In the areas, where the Quaternary cover is thick, large ascending springs are often formed (Saula Siniallikad, Purila springs). On the North-Estonian Plateau, a great many springs occur at the foot of eskers and kames at Iisaku and Kuremäe.

On the islands, the springs are usually attributed to the central infiltration area. On Saaremaa, the largest springs occur on the slopes of the West-Saaremaa Upland (Odalätsi, Pähkla, Viidumäe a.o.).

Numerous springs appear at valleys cutting into the bedrock both in northern and southern Estonia. Usually, these are permanent medium springs: Jõepere and Simuna Katkuallikas in Virumaa, Roosna-Alliku springs in Järvamaa, Tori and Allikukivi in Pärnumaa, Loodi Siniallikad in Viljandimaa, the springs in the Tatra Ancient Valley and at Rõuge in Tartumaa, Koorküla in Valgamaa, Rõuge, Urvaste and Illioru springs in Võrumaa. Small springs occur at the foot of drumlins and on the slopes of morainic hills. The hilly topography of the Otepää and Haanja heights is dotted with small permanent springs. The Väike-Emajõgi River originates in one of those springs in the vicinity of Lake Pühajärv.

Photo 80. Limestones of the Jaagarahu Stage on the Vaika islets are rich in karst phenomena of different types. *Photo by R. Karukäpp.*

ESCARPMENTS AND WATERFALLS

North-Estonian Klint

The North-Estonian Klint is among the most magnificent and spectacular geological monuments in Estonia and north-western Europe (Photos 18,19, 22, 23). Geologically, it is part of a much more extensive landform – the Baltic (Baltic - Ladoga, Ordovician) Klint. The limestone cliff begins from the western coast of the Island of Öland in Sweden, extends under the sea to the Estonia's western coast (Martinsson 1958), runs through Estonia to Russia and disappears under Devonian rocks on the southern shore of Lake Ladoga. The Baltic Klint has a total length of 1100-1200 km (Fig. 253); of this 250 km are in mainland Estonia. The klint is situated on the border of crystalline rocks of the Fennoscandian Shield and sedimentary rocks of the East-European Platform, and serves as an important exposed natural boundary.

The geological section of the North-Estonian Klint is well-known and rather simple. The Cambrian and Ordovician clays, silts, argillites and sandstones cropping out in its lower part are overlain by Ordovician limestones and dolomites (Figs. 254, 255). The structure of the klint changes considerably from the west to the east (Orviku 1940, 1961, Rõõmusoks 1983, Einasto & Saadre 1991, Heinsalu 1990, Мююрисепп 1960а, Орвику 1960б). To the west, the thickness of the *Dictyonema* argillite of the Türisalu Formation and glauconitic sand of the Leetse Formation (the Latorp Stage) decreases (Fig. 254, 255), while that of the carbonate rocks of the Volkhov, Kunda and Aseri stages increases. The upper part of the klint consists of the limestone of the Lasnamägi or Uhaku stages in western, and of the limestone and dolomitized limestone of the Aseri or Kunda Stage in eastern Estonia. The share of Cambrian terrigenous rocks is significant in the section of the central part of the escarpment. Between Selja and Päite, the outcropping Lontova blue clays have a rather great thickness (Fig. 255).

As a whole, the thickness of carbonate rocks in the topmost part of the klint section is relatively small. Terrigenous rocks prevail. Thus, the Estonian name *paekallas* (limestone bank), is a bit misleading.

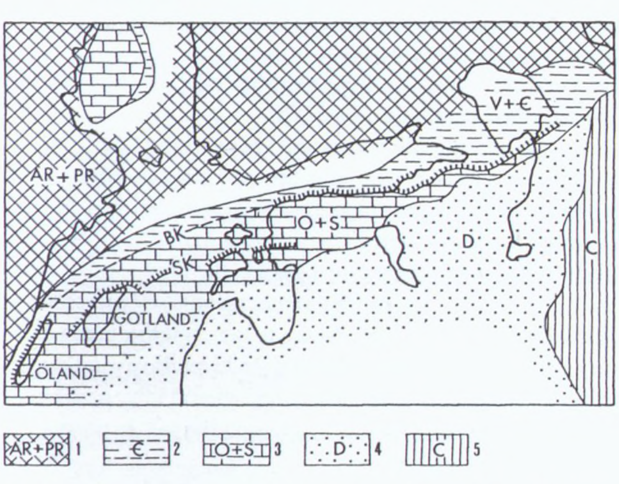


Fig. 253. Location of the Baltic (BK) and Silurian (SK) klints. Outcrops: 1 - crystalline Archaean and Proterozoic; 2 - Vendian and Cambrian; 3 - Ordovician and Silurian; 4 - Devonian; 5 - Carboniferous.

The geological structure of the klint provides convincing evidence of the gentle southward inclination and almost undisturbed occurrence of the bedrock's strata. The farther to the north, the higher in the section the strata of the same age are exposed. Along the coast it creates a delusive impression of the tectonic rebound of the area (Fig. 255). However, in some places tectonic disturbances have essentially changed the position of strata. Due to the Aseri Fault, the southeastern wing of which is more than 20 metres higher than the north-western one, the Kalvi-Aseri and Viru-Nigula-Kõrgküla klint sections repeat each other. In the klint, east of the Purtse Valley researchers (Стумбур 1959) have discovered a series of bedding disturbances, however, it is not yet clear whether these are tectonic or glaciotectonic. The Toila Fault has been acknowledged as tectonic in origin (Pirrus & Vaher 1975).

The height of the klint increases from the west to the east (Figs. 255, 256). On the Island of Osmussaar, where the klint emerges from the waters of the Baltic Sea, it is 6 metres high. Its first point on mainland - on the Pakri Cape, has an absolute height of 24 metres. In Estonia, the klint is at its highest (66 - 67.5 a.s.l.) between the Sagadi and Kandle villages, west of the Selja River. Its relative height is greatest at Ontika (55.6 m) and Päite (43 m). To the east, the height decreases rapidly; in Narva it is only 28-30 m above sea level.

Based on the bedrock exposure and extent of talus, four morphological types of the North-Estonian Klint are differentiated (Einasto & Saadre 1991):

1. Vertical cliff without talus, the bedrock is exposed in its full extent. This type occurs only in the sections open to immediate wave erosion (Osmussaar Island, Pakri islands, the Cape of Pakri, at Türisalu, Rannamõisa, Päite and Utria).

2. Limestone scarp, the lower part is covered by talus. This most widespread type of the escarpment occurs, for instance, in the vicinity of Tallinn, between Saka and Toila, within the Lahemaa National Park.

In the event of the above-mentioned two types, the margin of the limestone escarpment is dissected by tectonic joints which cut into rocks and divide the escarpment into small blocks. Most of the joints are of NW-SE or NE-SW orientation, however, farther in the east there are also joints running in a north to south or east to west direction (Teichert 1927a, Хейнсалу и Андра 1975).

3. Scarp or gentle slope with talus, completely covered with vegetation, the bedrock is not exposed. This kind of klint occurs between Viru-Nigula and Kõrgküla in the Lahemaa National Park.

4. The completely buried escarpment, not traceable in the topography. It is of rather limited distribution and occurs, for instance, between the Valgejõgi and Loobu rivers.

Tammekann (1940) divided the klint into klint peninsulas and interstitial klint bays. Between Osmussaar and Ladoga, he differentiated altogether 49 klint peninsulas and bays, of which 37 are within Estonia.

The klint peninsulas are NW-SE or N-S stretches of land with their steep extreme points often washed by the sea. In accordance with the dip of the bedrock's strata, the height of the limestone escarpment decreases from the north-west to the south-east, occasionally up to its interruption or complete

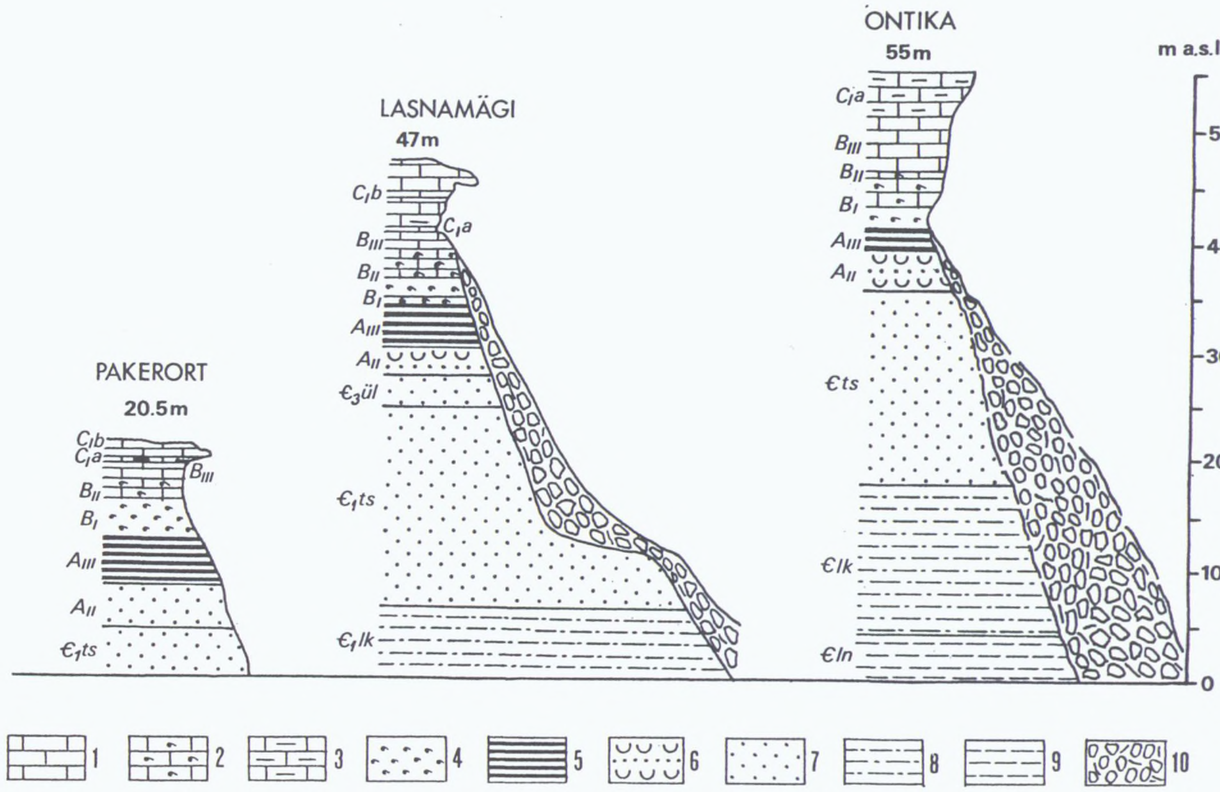


Fig. 254. Sections of the North-Estonian Klint (after Einasto & Saadre 1991). Legend: 1 - limestone; 2 - glauconitic limestone; 3 - oolitic limestone; 4 - glauconitic sand and sandstone; 5 - graptolitic argillite (*Dictyonema* shale); 6 - *Obolus* sandstone; 7 - fine-grained quartzose sandstone; 8 - siltstone; 9 - blue clay; 10 - talus.

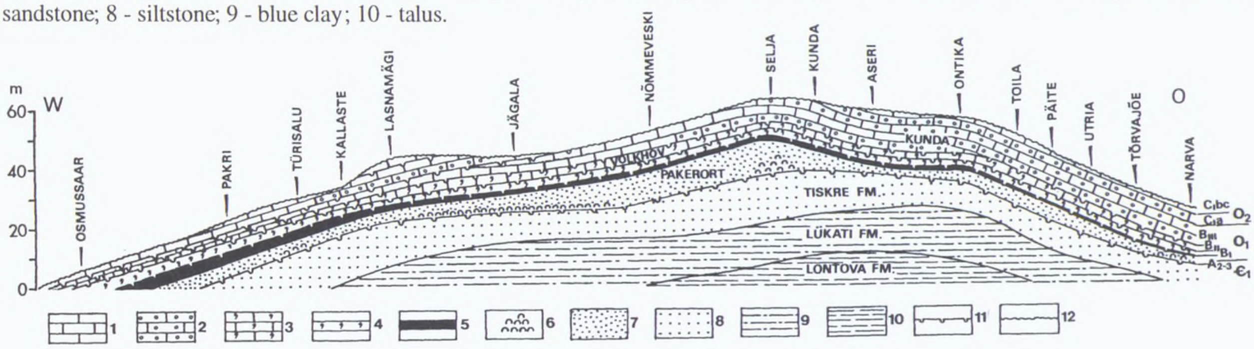


Fig. 255. Geological section along the edge of the Baltic Klint (compiled by R. Einasto). Legend: 1 - limestone; 2 - oolitic limestone; 3 - glauconitic limestone; 4 - glauconitic sand and sandstone; 5 - graptolitic argillite (*Dictyonema* shale); 6 - *Obolus* conglomerate; 7 - *Obolus* sandstone; 8 - fine-grained quartzose sandstone; 9 - siltstone; 10 - blue clay; 11 - discontinuity surface; 12 - bedrock surface.

disappearance. On the Pakri Klint Peninsula, Ordovician strata are exposed at a length of several kilometres, providing excellent possibilities for their study.

In klint bays, the bedrock is often several tens of metres below sea level. Klint bays are filled with up-to-100-metre-thick Quaternary sediments of the last ice age which are overlain by the Baltic Sea deposits. Beach ridges of the Ancylus Lake, Litorina Sea and Limnea Sea often run across the bays. The largest klint bays are Vääna, Tallinn and Valgejõe-Loobu. In Tallinn klint bay, three buried valleys have been identified (Кюннапуу и др. 1981, Тавaст и др. 1983). In the deepest of these valleys the bedrock is at least 145 metres below sea level. Klint bays are pre-Quaternary river valleys, dredged and widened by the ice sheet (Miidel & Tavast 1978, Тавaст и Раукас 1982 a.o.).

After the retreat of the last ice sheet, several North-Estonian rivers formed 15–35-metre-deep valleys with abundant rapids in klint bays. Spilling over the edge of the klint, the rivers formed waterfalls (Fig. 256).

On the basis of the height and shape of the limestone escarpment, width of the coastal plain and other morphological characteristics, Tammekann (1940) divided the North-Estoinian Klint into four parts.

1. Between Osmussaar and Tiskre, the klint forms a steep continuous escarpment with a rather narrow strip of shore in front of it. The limestone plateau is low, however, its height increases steadily up to 42 m above sea level. In several places, the sea is eroding the cliff. There are several klint peninsulas of northwest orientation jutting out into the sea. In places, the klint forms two terraces.

2. Between Tiskre and Kalvi, the gently sloping escarpment is half-buried by talus or covered with vegetation. The above-described morphological klint types 2, 3 and 4 prevail. The escarpment is dissected by numerous klint bays. The up-

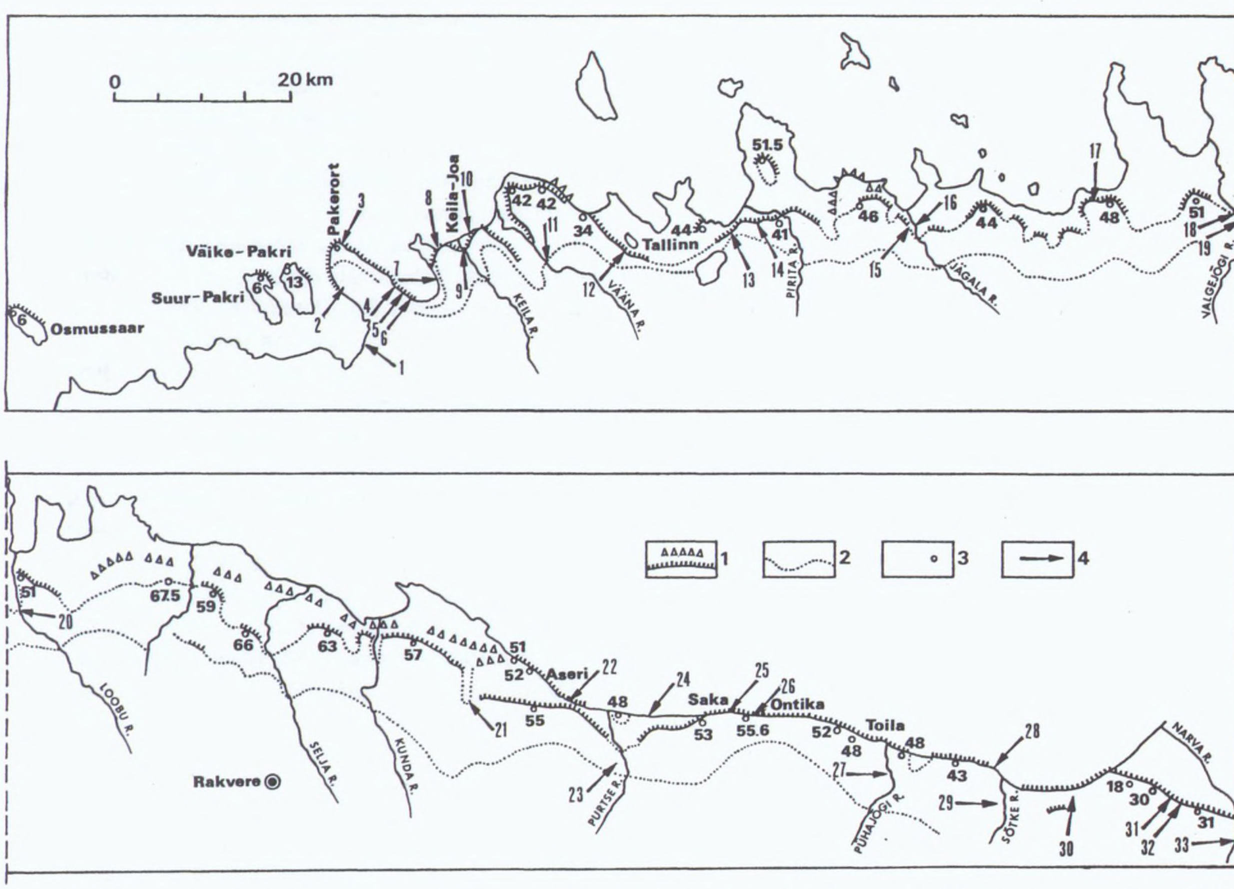


Figure 256. Sites of waterfalls: 1 - Madise (1.5); 2 - Paldiski; 3 - Pakri (5.5-5.8); 4 - Valli (2.2); 5 - Põllküla (2.0-2.5); 6 - Kersalu (2.5-3.0); 7 - Treppoja (5.6); 8 - Tornimäe (2.0-3.0); 9 - Keila (5.7); 10 - Türisalu (2.5); 11 - Vahiküla (Vääna) (4.5); 12 - Harku (1.3); 13 - Hundikuristik (3.8); 14 - Purde (4.5); 15 - Jõelähtme (2.1); 16 - Jägala (8.1); 17 - Turjekelder (4.0); 18 - Vasaristi (3.6); 19 - Nõmmeveski (1.2); 20 - Joaveski (5.1); 21 - Kohina-Linnamäe (2.5); 22 - Aseri (1.5); 23 - Uhaku (1.2), 24 - Ahermu (1.2); 25 - Karjaoru (8.0); 26 - Valaste (26.1); 27 - Aluoja (5.0-5.8); 28 - Ukuoru (4-5); 29 - Langevoja (3.9); 30 - Utria (3.2), 31 - Orasoja (1.1); 32 - Tõrvajõe (2.2); 33 - Narva (6.5).

Legend: 1 - North-Estonian Klint (white triangles - lower escarpment); 2 - outcrop area boundary of the so-called klint stages (BII - C_1b); 3 - height of the klint (a.s.l.); 4 - site of waterfall, in brackets - total height of the waterfall in metres.

to-20-km-wide coastal lowland in front of the klint has several minor scarps and an abundance of coastal formations of the Baltic Sea. The peculiar features of this klint section are well observable within the Lahemaa National Park (Linkrus 1976) where the escarpment is at its highest.

3. Between Kalvi and Meriküla, the klint approaches the sea and continues as an abrupt cliff with a thick layer of talus (up to 20m) at its foot. At Päite and Utria, the sea reaches the foot of the cliffs. Klint types 1 and 2 are most widespread. Between Saka and Toila, the cliff forms an unbroken and straight wall reaching 50-55 m above sea level. Blue clays crop out in the lower part of the escarpment, which due to the springs, promote landslides and soil creep. Eastwards from Toila, the escarpment gradually diminishes in height, and near the Meriküla Village it is only 30 m above sea level.

4. Between Meriküla and the Säsi River, at first the height of the escarpment continues to decrease and at Narva it is only 25-30 m above sea level. This section continues to Russia and reaches its highest point near Koporye (130-140 m a.s.l.). The escarpment is covered with vegetation and near the Town of Narva turns into a gently dipping slope, difficult to observe.

This chapter will not deal with the origin of the klint, because the genesis of the Baltic Klint has already been discussed in the subchapter Bedrock Topography. The only thing we would like to point out is that most researchers support the opinion that the klint is a denudational landform which is part of the cuesta-like bedrock topography of the East-European Plain (Tammekann 1949, Martinsson 1958, Шмидт 1883, Марков 1931б, Орвику 1960б, Таваст и Раукас 1982, Малаховский и Грейсер 1987). Several researchers (Шмидт 1883, Марков 1931б, Tammekann 1949) maintain that the klint started to develop during the period before the Middle-Devonian transgression. However, most likely, it happened in the Palaeogene when rifting in the northern part of the Atlantic caused the uplift of Fennoscandia and northwestern part of the East-European Platform which, in its turn, triggered again denudation processes (Пуура 1980). The klint obtained its present shape during the past 10,000-11,000 years as a result of the erosion by the waters of the Baltic Sea and its forerunners. On the Pakri Cape, the escarpment has been eroded in the course of some 7000 years. It is obvious that during that period the klint has recessed in a good deal leaving a flat eroded platform in front of it

The North-Estonian Klint is undoubtedly the most impressive natural monument in Estonia. In 1996, it was chosen as a symbol of the Estonian landscapes by the readers of the journal "Eesti Loodus" (Estonian Nature). For geologists it provides excellent conditions for investigating the outcropping Cambrian, Lower- and Middle-Ordovician rocks. On the basis of the klint or close-lying stratotypes, three Cambrian formations (Tiskre, Lükati, Lontova) and several members, five Lower and Middle Devonian stages (Pakerort, Varangu, Kunda, Aseri, Lasnamägi) and more than 20 members have been differentiated (Hints *et al.* 1993, Менс и Пиррус 1977). Besides, changes in facies and fauna can be studied on the basis of rocks untouched by tectonic processes and containing an abundance of well-preserved fossils which are exposed at a length of 250 km along the limestone escarpment. The klint has highly controlled Pleistocene and Holocene processes (Таваст и Раукас 1982).

Several sections of the klint have been designated as conservation areas. These are the stretches with a diversity of plant species in the deciduous forests skirting the foot of the klint, with the natural power and splendour attracting thousands of toursists, with a great aesthetic value teaching people to honour and protect nature, As early as 1957, the Saka-Ontika-Toila Nature Landscape Reserve was founded and the Rannamõisa Park was taken under protection. In 1996, a landscape reserve was founded at Pakri to protect the limestone escarpment. However, there are still several nature monuments in need of protection, like the Türisalu and Utria cliffs, *etc.* The Saku-Ontika-Toila Landscape Reserve should be expanded to the east to ensure preservation of this unique landform and to provide access to its most interesting and scenic sites.

West-Estonian Klint

The West-Estonian (Silurian) Klint runs from the mainland through the Island of Muhu to the northern coast of the Island of Saaremaa. At Vilsandi Island and Vaika islands it submerges into the Baltic Sea, continues westward on the sea floor and appears from the sea on the northern and western coasts of Gotland Island (Fig. 253).

The West-Estonian Klint is divided into four parts (Aaloe & Miidel 1967). The first section between Kergu and Pärnu-Jaagupi is represented by small knobs and hillocks consisting of hard reef dolomites of the Jaagarahu Stage. These small hillocks are usually 0.5-1.0 m, occasionally even more than 6 m high (Aaloe 1958).

Between Pärnu-Jaagupi and Mihkli (at Kirbla, Photo 23), Mihkli, Salumägi, Lihula, *etc.*) up-to-15-m-high hills are encountered. Their northern and northwestern slopes are higher and hold ancient cliffs cut by the waves of the Ancylus Lake and Litorina Sea. On the lee sides of some hills, spits were formed (Раукас и др. 1965). In the upper part of the hills hard, massive cavernous dolomites and at their base relatively soft medium-bedded cavernous dolomites of the Jaagarahu Stage are outcropping (Аалоэ 1956).

The third section includes an active cliff at the northwestern coast of Kesselaid Islet, and active (Kautliku, Püssina) and passive cliffs (Rannaniidi, Tupenurme, Üügu) at the northern coast of Muhu Island. The cliffs, eroded into the northern and northwestern slopes of hills have been described in particular detail by Aaloe (1958, Аалоэ 1956, Aaloe & Miidel 1967), Luha (1934, 1937, 1940b) and Märss (Мярсс 1988). According to these authors, the height of the cliffs ranges from 3 (Rannaniidi, Tupenurme) to 15 m (Üügu). They are built up either solely of hard massive cavernous reef dolomites (Rannaniidi, Tupenurme) or of these rocks on top and bedded platy dolomites below. In the lower part of the Kesselaid Cliff, dolomite marls of the Jaani Stage are outcropping. At the Kautliku Cliff, the sea is eroding marlstones of the Jaani Stage only. The Tupenurme Cliff dates from the Ancylus Lake, the Üügu Cliff with its roomy wave-cut notches from the Litorina and Limnea seas.

On the Island of Saaremaa, the West-Estonian Klint forms a system of cliffs, intensively eroded by the sea. The height of the cliffs ranges from 2.5 m (Paramaja, Liiva, Undva) to 21 m (Panga Cliff). The cliffs have cut mainly into marlstones, domerites, argillaceous dolomites and crinoidal limestones of the Jaani Stage. Only the Pulli and Panga (Photo 22) cliffs expose massive or bedded cavernous dolomites of the Jaagarahu Stage.

The West-Estonian Klint, unlike the North-Estonian Klint, does not form a continuous escarpment but consists of isolated small hills and cliffs. The hills, built up from hard reef dolomites resistant to exogenic processes, are classified as rocky drumlins, reworked by the waters of the Baltic Sea (Luha 1934, 1937, 1940b; Aaloe, 1958, Таваст и Раукас 1982).

The most important cliffs and hills like Salevere Salumägi, Kirbla, Üügu, Rannaniidi, Tupenurme and Panga are under protection.

Waterfalls

In Estonia there are at least 28 natural waterfalls with a height of one or more metres. Together with the waterfalls, damaged or strongly changed by man the number will reach 33 (Fig. 4). Basing on the circumstance that almost all the waterfalls, with a few exceptions only, are located on the North-Estonian Klint, Orviku (1935a) introduced the term "North Estonian Fall Line" into scientific literature. As the Baltic Klint with its numerous falls crosses Estonia's eastern frontier and continues in western Russia, the Estonian fall line is actually part of the more extensive Baltic Fall Line.

According to the classification by Schwarzbach (1967), Estonian waterfalls are destructive with more or less headward erosion. They are divided into consequent and subsequent falls. Many waterfalls in Estonia, like those on the Pakri Peninsula, at Vahiküla, Valaste, Karjaoru are consequent, *i.e.* they plunge down from the pre-existing escarpments, not formed by rivers. Still, the secondary subsequent falls are in the majority. This group includes the falls which initially came into being as consequent waterfalls but later, as a result of erosion, lost their original significance and turned into subsequent ones (Keila, Jägala, Langevoja, Narva a.o.). The height of the waterfalls varies greatly, ranging most frequently from 3 to 8 metres. The highest waterfall (26 m) in Estonia is situated at Valaste in the Saka - Ontika - Toila Landscape Reserve. In this man-made waterfall the water drops from the mouth of a ditch. In all likelihood, the falls of Pakri (with a drop of 5.5-5.8 m), Karjaoru (two steps, total height 8 m), Madise (height 2.5 m) and Tornimäe (height 3 m) were also generated by man, since they are also related to ditches.

Not all waterfalls are spilling over the edge of the klint, but some are situated rather far inland, on the tributaries of larger rivers (Langevoja and Kohina-Linnamäe falls), but not on the main river itself.

Most of waterfalls have only one step (Narva, Jägala (Photo 19), Keila, Langevoja, Türisalu a.o.). However, there are some smaller ones with up to six steps which form cascades (Treppoja, Joaveski, Aluoja a.o.). For instance, the fall at Joaveski has 6 steps with a total drop of 5.2 m (Teemusk 1967), the distance between the uppermost and lowermost steps being 160 m. The cascades were formed on a stepped bedrock surface where the loose deposits, once covering the bedrock, had been washed away by the running water.

Geologically, the Estonian waterfalls are similar in construction due to their relation to the Baltic Klint. The upper part of waterfalls consists of hard Ordovician limestones and dolomites, usually of Kunda and Volkhov stages (Keila, Jägala, Aluoja a.o.). These horizontal strata overlie the weaker, either thin-bedded clayey carbonate rocks, sandstones or shales of Ordovician and Cambrian age, usually of the Billingen, Hunneberg, Varangu and Pakerort stages (Keila, Jägala, Linnamäe, Turjekelder, Valaste, Langevoja and Tõrvajõe falls).

Sometimes the uppermost part of a waterfall consists of carbonate rocks belonging to the Aseri (Narva Falls) or even to the Lasnamägi and Uhaku stages (Pakri). The cascades at Treppoja (Photo 21) and Vahiküla are formed by carbonates of the Lasnamägi Stage. Many small waterfalls are related to the lower escarpment of the klint, being formed in Cambrian terrigenous rocks (*e.g.* Aseri, Ahermu, Türisalu).

Some waterfalls (Keila, Jägala, Langevoja a.o.) are typical cap-rock falls, *i.e.* the uppermost part is resistant to erosion and forms a protective cap. They are not in their initial position, but are receeding upstream. Evidence is derived from deep and steep-walled canyons formed downstream of waterfalls, and in some cases also from historical data. Thus, comparison of old Swedish and recent maps has shown that during 1688-1931 the waterfall at Jägala retreated 42 m, *i.e.* at an average rate of 17.3 cm per year (Kaljuvee 1931). This result is in good accordance with the data presented by Venjukov (Венюков 1882), according to which the Jägala Fall receded 3.21 m or 16 cm per year during the course of 20 years (Photo 19). By means of old maps it was established that from 1862 to 1977 the Keila Fall regressed 11 m or 9.7 cm per year (Kumari 1977). As a result of undermining of the upper hard layers, a roomy cave was formed in the soft rocks. Its depth reaches 3 m at the waterfalls of Keila and Jägala.

Due to the erosion of falling water, plunge pools with a depth from 2-3 m (Keila and Jägala waterfalls) to 6 m (Narva Falls) were formed in front of waterfalls. The upstream migration of waterfalls is marked by canyons, exceeding 12-20 m in depth (Keila, Jägala and Narva). The total upstream distance of migration ranges from some hundred metres to some kilometres (Narva Falls). The process is considerably favoured by tectonic joints dividing the bedrock into small blocks and determining the direction of retreat (Мийдел 1971, Miidel 1976).

However, not all waterfalls have retreated. The site of cascades (Treppoja, Vahiküla a.o.) has not changed in the course of time. The falls are practically stable because there are no geological preconditions promoting their recession (Miidel 1976).

All Estonian falls are young in geological sense. They were formed in the post-glacial period, according how land emerged from the waters of the Baltic Sea. In eastern Estonia, the formation of waterfalls probably started after the drainage of the Baltic Ice Lake. The majority of waterfalls in western Estonia came into being either after the transgression of the Ancylus Lake or even considerably later, *i.e.* in the time span 9300-3000 years ago.

The Narva Falls were among the greatest waterfalls in Europe, not so much because of there height (6 m) but with the discharge. In the periods of floods, the water discharge extended to 1500-1800 m^3 per second. However, in 1955 the Narva River was dammed up during the construction of the Narva Hydroelectric Power Station and since then the falls have been dry.

In view of the scenic beauty, aesthetic and scientific value, the major waterfalls including Treppoja, Keila, Jägala, Joaveski, Narva, Langevoja, Tõrvajõe, were put under protection in 1957.

PEATLANDS

There are 9836 mires with a total area of 1,009,101 ha in Estonia, which cover 22.3% of the land area (Orru 1992, 1995). In 1994, 1.05 million tonnes of peat were produced in Estonia. This was twice as much as the annual natural increment (*ca.* 500,000 tonnes). During 1980-90, the amount of peat produced was even four times as high as its natural increment in mires. By today, 120,000 ha of mires have been drained for agricultural and 180,000 ha for silvicultural purposes, and 30,000 ha for peat production. Mining of oil shale in northeastern Estonia has destroyed 2000 ha of mires, to which 100 hectares is added annually (Ilomets *et al.* 1995).

Mires as geological monuments are invaluable, since they present an excellent basis for studing the process of peatland formation. During its accumulation, peat stores information about environmental conditions. The botanical composition of peat and preserved fossils impart valuable information about the evolution and age of mires. On this basis, palaeogeographic and palaeoclimatic conclusions can be drawn which are urgently needed for directing human activities in the fields associated with the Earth's crust, ground- and surface water, *e.g.* recultivation of mining regions, amelioration, *etc.* Peat extraction is accompanied by drainage of the area breaking the natural development of the mire. Hence, preservation of water level is of primary importance in keeping the natural processes ongoing in mires.

The main reasons for the preservation of a part of our peatlands in an undrained condition are as follows:

- conservation of the reserves of pure water (the peat layer filters precipitation which is contaminated nowadays);
- conservation of peat resources;
- conservation of plant resources (particularly berries, honey and medicinal plants);
- protection of rare plant and animal species;
- protection of medicinal waters and curative mud;
- protection of landscapes for recreational purposes (tourism, hunting, fishing);
- preservation of areas serving as comparison standards in scientific research.

The first mire reserve (1109 ha) was founded at Ratva in 1938. Most of the mire reserves date from 1980. During 1993-

95, local reserves were reformed into national nature reserves and the Soomaa National Park and the Alam-Pedja Nature Reserve were established. At present, 95 mires with a total area of 1799 sq km were partially or fully protected in Estonia (Fig. 257). The main purpose of mire reserves is to preserve the mire ecosystem (botanical-zoological aspect). Besides, mires may be treated as pure water accumulators. In view of the above, protection of mires as water preservation areas is of great significance, since it would quarantee the quality and quantity of water reserves (hydrological-hydrogeological aspect) in the regions of intensive land use (Kink & Andresmaa 1995).

Table 78 shows the number and area of mire reserves and mires in protected areas by counties. It also points out the number of mires in most critical condition which urgently need to be conserved as water preservation areas. For instance, on Hiiumaa and Saaremaa, where the water regime is extremely sensitive to human impact, all the mires should be conserved as water preservation areas. Protected mires together with water preservation areas should cover ca 30% (at present 18%) of the total area under mires (Kink, unpublished data).

H. Kink has worked out a hydrogeoecological study method which enables to assess the prospective of any single mire as a water conservation area. The method takes into consideration the geological conditions of the mire, particularly its type and genesis, the load of the drainage area, its water reserves, and allows to predict potential changes in the water regime (Kink & Metslang 1981).

In fens, fed by groundwater, the intensity of nutrition must be taken into consideration. As a result of the draining of mire, the groundwater table will drop. In karst areas, this negative aspect may manifest itself within an area of 1-2 km, in sandy regions within 0.5 - 1 km around the mire (Kink 1975). If drains cut into an aquifer which contains water under artesian pressure head, the amount of the water to be removed will increase abruptly.

In the transitional and raised bogs, formed as a result of terrestrialization of lakes, peat usually overlies lake mud or lime. It isolates the bog water stored in the peat layer from the subsurface groundwater protecting, thus, the latter from draining. In terms of water preservation, the density of hydrographic network in peatland is of great importance, because in spring meltwater drains into rivers and ditches and flows rapidly away. The water regime is endangered in the vicinity of bog-pools which are particularly sensitive to human impact (Photo 81). For instance, in the Viru Mire a drain cut open several bog-pools, which then accounted for ca 90% of the water removed from the mire.

The mires, formed as a result of paludification of mineral soils, do not have any water-proof layer of lake sediment and the subsurface groundwater is hydraulically connected with mire water, except the cases when peat overlies clay or loam. If peat is underlain by sand or fissured limestone, the natural groundwater protection is weak or absent.

Particularly endangered are the so-called "hanging" mires (*e.g.* Meenikunno) where peat overlies a layer unsaturated with water. If ditches cut in this layer, the whole mire may run dry. Peculiar man-generated "hanging" mires occur in north-eastern Estonia where the groundwater table has dropped as a result of mining and the connection between the bog and groundwater is lacking (Andresmaa *et al.* 1994).

The effect of bog water on surrounding areas depends on the type of landscape, sediments and dissection of the area and is not always reflected in water level changes only. For instance, in karst areas the acid mire water may significantly promote the development of karst processes.

In 1978, the researchers of the landscape ecology working group at the Institute of Geology undertook monitoring of the chemical composition of mire water in Estonia's nature reserves. The main purpose was to study the formation of water under different geological conditions (Andresmaa *et al.* 1993). Mire water in general is characterised by a heightened concentration of nitrogen compounds due to the decay of organic matter. The raised bog water is low in the mineral matter. The chemical composition of water in the spring-mire (Viidumäe) is much the same as in groundwater and the concentration of nitrogen is lower than in bog water. The surface and groundwater composition in unprotected areas is much the same as in mires (Meenikunno). In the peatlands, receiving air-borne pollutants from towns and industrial enterprises, mire water is affected. As an illustration serves the Kunda

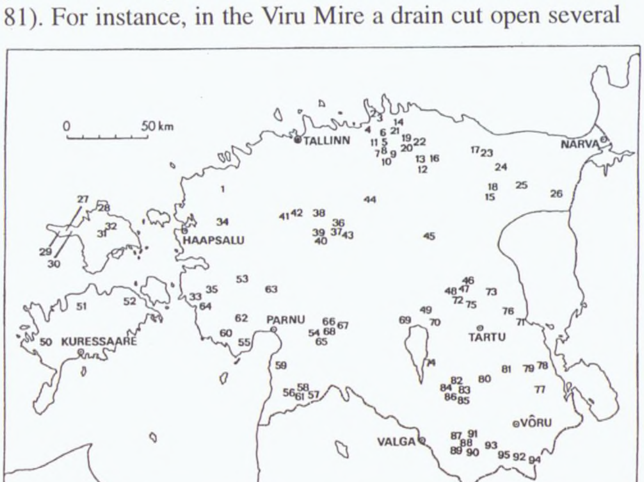


Fig. 257. Mire reserves (underlined) and mires in protected areas as of January 1, 1997. **Harju County**: 1 - Suursoo; 2 - Aabla; 3 - Hara; 4 - Kahala; 5 - Suru, 6 - Viru; 7 - Koitjärve; 8 - Salu; 9 - Muki; 10 - Lille; 11 - Salgu. **Lääne-Viru County**: 12 - Kiku; 13 - Kõrbse; 14 - Laukasoo; 15 - Luusaare (Venevere); 16 - Neeruti; 17 - Sämi-Kuristiku; 18 - Tudu-Järvesoo; 19 - Tuksmani; 20 - Udriku; 21 - Uuesmõisa; 22 - Vohnja. **Ida-Viru County**: 23 - Uljaste; 24 - Sirtsi; 25 - Muraka; 26 - Agusalu. **Hiiu County**: 27 - Eiste; 28 - Kodeste; 29 - Kõivasoo; 30 - Kõrigu; 31 - Pihla; 32 - Määvli. **Lääne County**: 33 - Nehatu; 34 - Marimetsa; 35 - Tuhu. **Rapla County**: 36 - Aela-Viirika; 37 - Loosalu; 38 - Mahtra; 39 - Keava; 40 - Palasi; 41 - Kodila-Linnuraba; 42 - Hagudi. **Järva County**: 43 - Piiumetsa (Rumbi); 44 - Epu-Kakerdi. **Jõgeva County**: 45 - Endla; 46 - Kaiavere; 47 - Lava; 48 - Ulpe; 49 - Umbusi. **Saare County**: 50 - Viidumäe (Lümanda); 51 - Pelisoo; 52 - Koigi. **Pärnu County**: 53 - Avaste; 54 - Kikepera; 55 - Lindi; 56 - Nigula; 57 - Rongu; 58 - Ruuna; 59 - Tolkuse; 60 - Tõhela; 61 - Tõrga; 62 - Võlla; 63 - Kodesma; 64 - Soometsa; 65 - Ojasaare. **Viljandi County**: 66 - Kuresoo; 67 - Valgeraba; 68 - Ördi; 69 - Parika. **Tartu County**: 70 - Emajõe-Pedja (Laeva); 71 - Emajõe-Suursoo; 72 - Punnasjärve; 73 - Sirnuvere; 74 - Valguta; 75 - Vasula; 76 - Laukasoo. **Põlva County**: 77 - Meenikunno; 78 - Meelva; 79 - Valgeraba; 80 - Ahijärve; 81 - Jõe. **Valga County**: 82 - Nuustaku; 83 - Pori; 84 - Rebaste; 85 - Vidrike; 86 - Pühajärve; 87 - Köstrejärve; 88 - Palu; 89 - Koobassaare; 90 - Väike-Apja; 91 - Lauksilla. **Võru County**: 92 - Kellamäe (Vanamõisa); 93 - Kaugjärve; 94 - Murati; 95 - Ubajärve.

Table 78. Protected mires and prospective water protection areas

County	Mire reserves	Mires in protected areas	Prospective water protection areas	Per cent of protected mires in the county
	Number/Area (ha)			
Harju	1 / 5793	10 / 8971.1	10 / 27993	47
Lääne-Viru	3 / 2893.6	8 / 4673.1	4 / 3234	23
Ida-Viru	4 / 19293.1	-	9 / 8691	24
Hiiu	-	6 / 3359.1	2 / 1342	67
Lääne	3 / 6453.2	-	6 / 15921	46
Rapla	7 / 10078	-	3 / 3702	25
Järva	1 / 582	1 / 4245	15 / 9458	19
Jõgeva	1 / 8162	4 / 9781	4 / 3156	33
Saare	1 / 364	2 / 2922	9 / 3725	33
Pärnu	13 / 34436.5	-	6 / 6327	35
Viljandi	4 / 22195.5	-	8 / 16727	53
Tartu	4 / 25886	3 / 2867	3 / 1720	32
Põlva	5 / 3305	-	13 / 4055	19
Valga	-	10 / 1047	11 / 9719	32
Võru	1 / 1546	3 / 1072	8 / 2039	12
Total:	48 / 140987.9	47 / 38937.3	111 / 117809	33

area where the pH value in mire water is occasionally higher than 7 (Kink, unpublished data).

In recent years, attention has focused on the contribution of mires to the global carbon circulation. The circulation of substances through the mire's ecosystem binds atmospheric carbon dioxide into organic compounds. As a result of draining, the process reverses itself; the CO_2 added to the atmosphere from mires promotes the development of the greenhouse effect. In Estonia, about 300,000 ha of mires have preserved in their natural condition which means that some 70% of our mires have been drained or are within the area of the influence of drainage, and peat no longer accumulates in these mires. Under the present economic conditions it is not possible to maintain all draining systems, to say nothing of creating new ones. Therefore, in the peatland with groundwater nutrition the water has to be dammed up to create conditions for peat increment.

Photo 81. Bog-pools are especially sensitive to human impact. Most of the Laukasoo Mire in the Lääne-Viru County is under nature protection. *Photo by M. Orru.*

According to the World Conservation Union (IUCN), some 50% of peat-forming environments have vanished from the face of the Earth (The Trondheim... 1994), in some European countries the mires no longer exist. Protection of mires as water and plant preservation areas is topical all over Europe (Akkerman 1982, Succow & Jeschke 1986). We are lucky to have many mires preserved in an almost virgin state, which gives Estonia an important responsibility for international mire conservation. The threats to our mires will grow in near future because peat is gaining in importance in Estonian economy, both for export and inland energy production. However, we have to learn from the dramatic experience of other countries and take measures to protect our mires and preserve them for our descendants.

PANDIVERE WATER PROTECTION AREA

The Pandivere Upland is part of Intermediate Estonia (Varep 1964) and one of the largest karst areas in Eastern Europe. Granö (1922) classified the Rakvere - Pandivere region as an area of large eskers and knolls with abundant fields, groves, big villages and manors.

The state of waters within the Pandivere Upland has been studied in the course of several decades (Aruja *et al.* 1976, Eipre 1987). Deeply worried about the critical situation in the late 1980s, the authorities of the Rakvere and Paide districts together with the Commission on Nature Conservation of the Estonian Academy of Sciences appealed to the Government of the Republic with the proposal to establish a water protection area in this region. The Pandivere Water Protection Area (PWPA) was founded on December 8, 1988 with the purpose "... to protect the most important water resources of great economic importance, the ground and surface water formation area of North and Central Estonia, to avoid irretrievable environmental damages caused by industry or other economic activities." (Decree No.744, 9.1. of the Supreme Council of the Estonian SSR). This programme area was the first of its kind in Europe. The main objective of PWPA is to support the protection and sustainable use of the surface and groundwater resources through the main water and land users (Raukas 1993).

Currently, the PWPA is managed by the Lääne-Viru and Järva county governments. Within the PWPA there are the lands of 18 communes, the towns of Rakvere, Tamsalu and Tapa and the settlements of Ambla, Järva-Jaani and Lehtse.

The water preservation area is situated on the Pandivere Upland. The upland consists of Ordovician and Silurian limestones, dolomites and marls, making up a complex with a total thickness of 200 metres, which provides water for the northern Estonia's villages and towns (Savitskaja & Kink 1987). The water supplied by carbonate rocks is fresh, its chemical composition being HCO_3-Mg-Ca. The largest water reserves are related to karst zones and springs (100-200 litres per second) where all North-Estonian rivers start from (Fig. 258). Under natural conditions, groundwater is unprotected or poorly protected against pollution. The primary source of groundwater recharge is precipitation which percolates into ground.

The pollution proofness of groundwater is controlled by the geological situation in the region. From the hydrogeological point of view, the intensity of infiltration is of great significance. According to the level of groundwater protection, the following zones are distinguished in the PWPA:

- **unprotected areas**. The thickness of the sedimentary cover is less than 2 m. Surface water passes freely through the rocks in the alvar areas and karst fields;
- **poorly protected areas.** The layer of till is 2-10 m thick;
- **moderately protected areas.** The cover of clayey sand exceeds 10 m in thickness or there is a more than one metre thick loamy screen which has not been damaged by man. This zone includes also areas where the piezometric level of artesian water lies near the ground level;
- **well protected areas.** There must exist an undamaged cover of clay or loam with a thickness greater than 3 m.

The above-presented criteria allow the following subdivision of the PWPA:

- **water preservation areas** include the regions worth of protection from the aspect of water and landscape (*e.g.* karst fields and springs). These areas are classified as nature reserves with restricted human activity that could affect the quality of water;
- **areas of regulated use** include the regions with unprotected groundwater and the protected zones of surface water. Agricultural activities are allowed, but restrictions on land use are more severe (*e.g.* 80 kg of nitrogen fertilizers per hectare, 1 animal unit per hectare);
- **areas of restricted use.** Groundwater is poorly protected. The amount of nitrogen fertilizers allowed per hectare is 100 kg, the number of animal units is 1.5. Percolation of waste water into the soil must be avoided. Foundation of landfills or other potential pollution sources is forbidden in the area;
- **areas with general regime.** Groundwater is moderately or well protected. The regulations are less strict, undertakings with a potential pollution effect will not pose a threat to the environment.

As a rule, groundwater is poorly protected in the most fertile lands of the Pandivere region, and well protected in mire landscapes.

The distribution of water consumption is unfavourable in the Pandivere WPA. For instance, in the Järva County, 17% of private wells are situated in unprotected, 82% in poorly protected and only 1% in moderately protected areas.

Under the conditions of the present economic depression, the consumption of water is decreasing and the quality of water in the upper aquifers shows the signs of improvement. As a rule, the concentrations of NO_3^- and Cl^- have been decreasing more rapidly, while the decline in SO_4^{2-} has been slower (Table 79). The improvement of water quality owing to the restricted use of fertilizers is evident. Contamination of groundwater with nitrogen compounds in Estonia is frequently decisive in the formation of the surface water quality, because in large areas as, for instance, the Pandivere region, the rivers are recharged mostly from the uppermost aquifer.

However, it should be born in mind that Pandivere has always been and will be the area with the most intensive

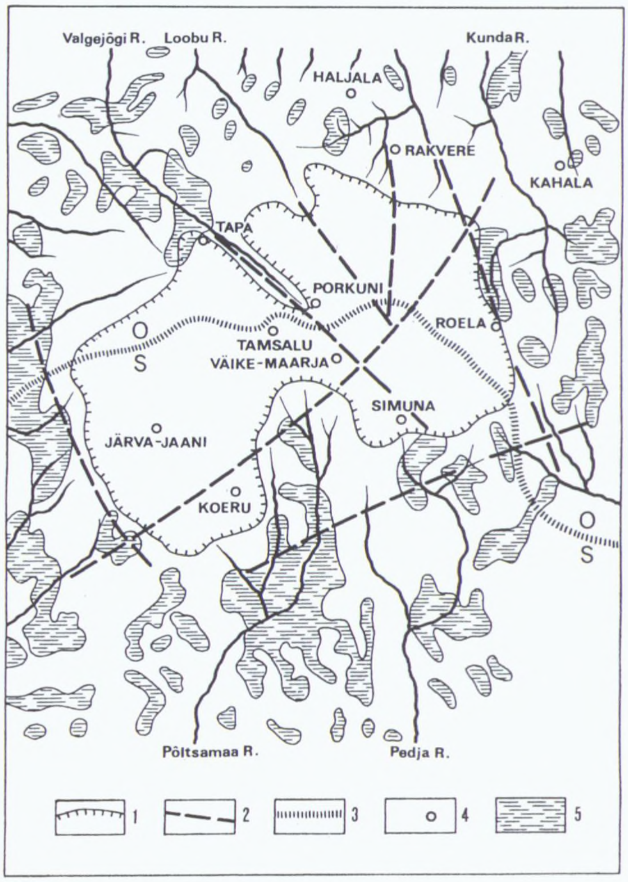


Fig. 258. Pandivere Upland: 1 - roof; 2 - zones of disturbances; 3 - outcrop boundary of Ordovician and Silurian rocks; 4 - towns and settlements; 5 - mires.

Table 79. Changes in average content of NO_3^-, Cl^- and SO_4^{2-}

Parameter	Year	Content		
		NO_3	Cl^-	SO_4^{2-}
Spring water (average content)	1991	25.1	17.3	40.5
Change from previous year	1992	-1.7	+1.2	-1.2
Change from previous year	1993	-3.5	-3.3	-0.2
Change from previous year	1994	-6.3	-2.0	-4.6
Private wells of the Pandivere Upland (average content)	1991	26.0	20.0	39.6
Change from previous year	1992	-3.8	-1.1	-1.8
Change from previous year	1993	-3.2	-2.3	+1.1
Change from previous year	1994	-5.4	-1.8	-4.2
Water supply wells (average content)	1991	23.5	26.4	39.8
Change from previous year	1992	-5.5	-3.1	-2.5
Change from previous year	1993	-3.6	-4.6	-1.7
Change from previous year	1994	-4.4	-0.5	-2.5

agriculture in Estonia.

As already mentioned, the Pandivere WPA abounds in karst formations and springs. Besides, it holds a lot of big erratic boulders and limestone cliffs (Fig. 259). The development of both surface and buried karst forms has been promoted by the location of the area on the watershed, relative heights and zones of tectonic disturbances. The roof of the upland is dotted with temporal lakelets. Of the 357 karst areas registered in the Pandivere WPA (Karst... 1994), 15 have been entered into the Book of Primeval Nature and 12 are classified as water preservation areas (Fig. 259).

Important water reservoirs are the Assamalla Karst Field,

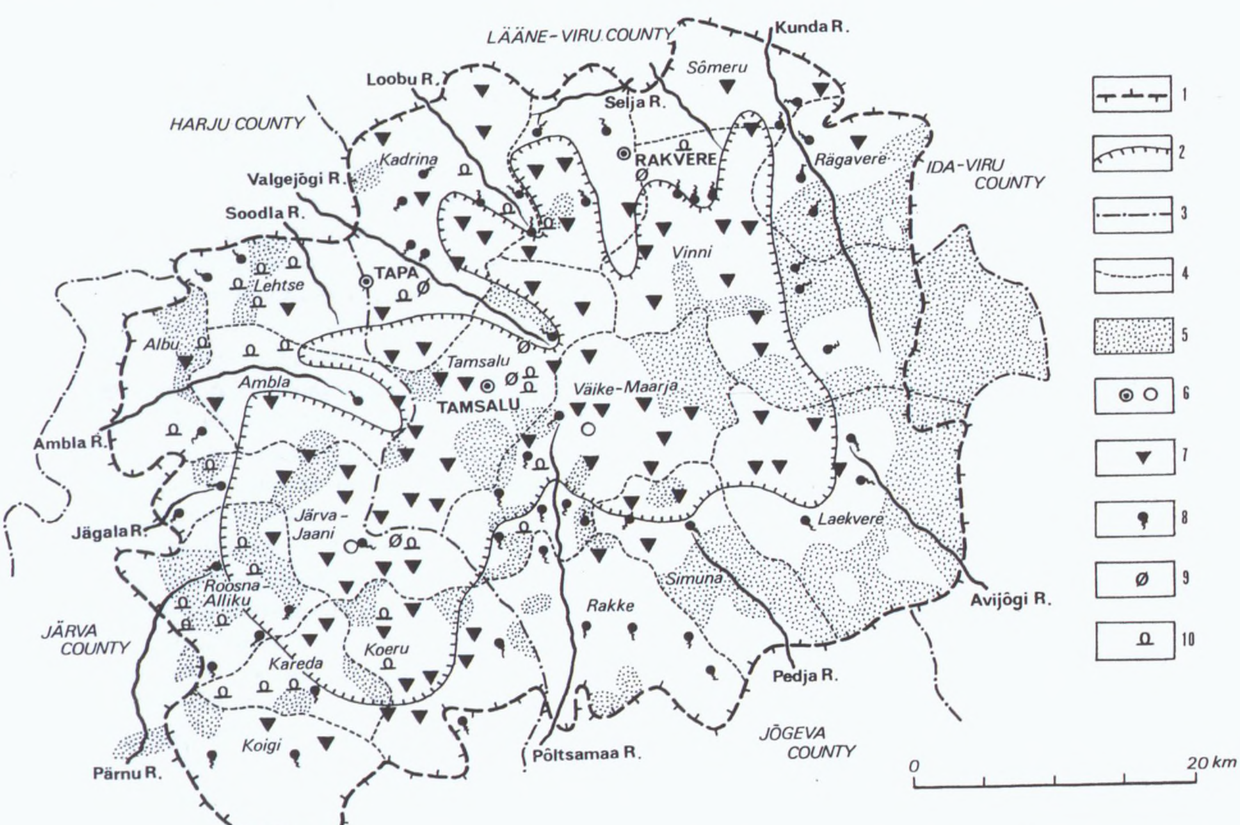


Fig. 259. Geological monuments and water reserves under protection. Boundaries of the: 1 -water protection area; 2 - roof of the upland; 3 - counties; 4 - communes; 5 - forests; 6 - towns and settlements; 7 - karst forms; 8 - springs; 9 - outcrops; 10 - erratic boulders.

and the karst lakes of Äntu (depth 8 m) and Porkuni-Võhmetu-Lemküla. From the scientific point of view, most valuable are alkalitrophic, particularly temporary karst lakes. Lake Porkuni and the clear-water Lake Äntu Sinijärv provide favourable habitats for interesting species of flora and fauna (Mäemets 1977). Of the 90 springs and spring areas within the PWPA, 28 have been included in the Book of Primeval Nature, 20 are located in the water preservation areas.

Agricultural pollution load is generated by both point and areal sources. Major point-pollution sources are cattle farms, agricultural production processing plants and former military sites. Areal pollution is caused by misuse of fertilizers. In the WPA, 100 kg of nitrogen is allowed to be used per hectare. The potential agricultural load is 1,500,000 population equivalent. Some 30% of polluting substances enter natural waters. In the river flow, the annual amount of nitrogen varies from 10 to 82 kg per hectare, in the water of shallow wells it is 25 to 60 mg per litre.

In accordance with the decision of the Council of the Pandivere WPA of November 13, 1990, a ground- and surface water monitoring network was laid out with an aim of studying the state of water within the protection area. More than 5000 water samples of the Pandivere WPA have been studied. The results obtained were used in compiling the maps (scale 1 : 10 000) on natural water protection, technogenic load and environmental situation. Nitrogen balance was assessed, recommendations for improving the situation were given.

The conservation of the Pandivere WPA is of great significance, first of all, in terms of the groundwater resources and the rivers flowing out from this region. The goals of the water protection are:
- preservation of drinking water;
- ensuring the recharge of deeper aquifers with high-quality water;
- preservation of the quality of water on the upper courses of rivers;
- preservation of rare landscape features.

NATURE CONSERVATION AREAS AND COMPLEX MONUMENTS

Nature protection has long traditions in Estonia. In 1910, a bird sanctuary was established on the Vaika islets in the West-Estonian Archipelago. In 1957, the third Nature Conservation Law of Estonia was approved. The Law of the Protection of Nature in Estonia, approved in 1990, summed up the achievements in this field. In 1994, there were four national parks, four nature reserves, one nature park, 13 landscape reserves, 25 mire reserves and two programme areas in the Republic (Peterson 1994).

All the areas under protection are in the landscapes typical of Estonia. Research into the hydrosphere and geosphere (Kink *et al.* 1996) was commenced with mapping (1:10 000); constant checkup of the chemical composition, level and runoff regime of surface and groundwater has been carried out since 1978 (Fig. 260). In the Pandivere Water Protection Area the monitoring network was laid out in 1990. In 1991, the number of monitoring sites reached 173.

Lahemaa National Park, 650 km in area, was founded on June 1, 1971, to protect ecosystems typical of the northern part of Estonia and cultural landscapes with historical, cultural and geological monuments (Fig. 260). Natural landscapes and the reserve make up 70% of the park.

The park covers the Coastal Lowland, the North-Estonian and North-East Estonian limestone plateaus and the Kõrvemaa area where the various relief forms of glacial origin, including kames, radial eskers, glaciofluvial deltas and glaciolacustrine plains, occur in their classical form. The pre-Quaternary bedrock consists of Upper Proterozoic, Cambrian and Lower Ordovician carbonaceous and terrigenous rocks. The overlying marine, glacial, glaciofluvial and gaciolacustrine deposits are usually 1-20 metres thick.

The North-Estonian Klint runs through the park and is sectioned by numerous klint headlands with klint bays between them. The klint headland at Muuksi has an absolute height of 47 m and a relative height of 20 m. On klint headlands, where the limestone bedrock crops out on the surface, numerous *alvars* are encountered. At the edge of the klint, the Valgejõgi and Loobu rivers, flowing into the Gulf of Finland, have eroded deep canyon-like valleys with a striking series of waterfalls on the bottom, including Nõmmeveski (Photo 52), Vasaristi (Photo 78) and Joaveski. Some 150 springs occur under the cliff. The park holds 14 lakes. Five karst areas with underground streams and springs occur on the plateau.

The limestone cliff at Muuksi-Tsitre, four waterfalls, the valley of the Valgejõgi River, four outcrops, six karst areas and ten springs have been included in the Book of Primeval Nature. Mires cover 8.5% of the Lahemaa area. Lahemaa is rich in big erratic boulders and stone fields. 122 boulders have been listed in the Book of Primeval Nature. *Majakivi*, the largest boulder in Lahemaa, has an height of 7 m and is situated on the Juminda Peninsula. Big boulder accumulations occur on the Käsmu Peninsula and in the park of Palmse Manor. There is the Boulder Museum near the seaside village of Altja (Viiding 1981).

Groundwater is stored in the Quaternary deposits and in the lime- and sandstones of the sedimentary cover. The water resources at Lahemaa have no natural protection or it is insignificant (Fig. 261). To study the water stored in the Quaternary, Ordovician and Cambrian-Ordovician aquifer systems, a network of groundwater monitoring sites from 46 wells (depth 2.5-20 m) at 18 sites was founded. At a depth of 10 m, the studies did not reveal any traces of pollution in the natural landscape of the Muuksi alvar, while in the regions with intensive agriculture the pollution load in drinking-water proved heaviest in karst areas and on alvars. The aim of the monitoring was to collect data and on this basis to recommend measures for protecting water in areas with similar natural conditions and anthropogenic influence (Kink *et al.* 1994, 1986).

Matsalu Ornitological Reserve (Wetland) 48,460 hectares in area, was founded in 1957 in the catchment area of Matsalu Bay (Fig. 260). In 1957, it was designated as a wetland area of international importance. The main lines of protection included the checking of eutrophication of the bay and the management of areas with a regulated use of nature.

The reserve in the typical western Estonian landscape is situated on a plateau where Lower Silurian carbonate rocks

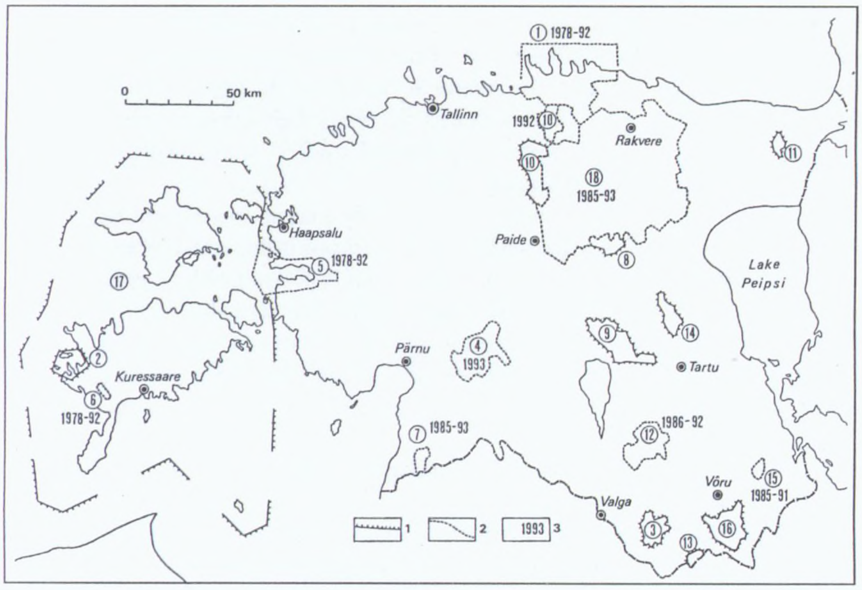


Fig. 260. **National parks:** 1 - Lahemaa; 2 - Vilsandi; 3 - Karula; 4 - Soomaa. **Nature reserves:** 5 - Matsalu; 6 - Viidumäe; 7 - Nigula; 8 - Endla; 9 - Alam-Pedja. **Landscape reserves:** 10 - Kõrvemaa (the northern part was a former military area); 11 - Kurtna; 12 - Otepää; 13 - Paganamaa; 14 - Vooremaa. **Mire reserve:** 15 - Meenikunno. **Nature park: 16 - Haanja. Programme areas:** 17 - West-Estonian Archipelago Biosphere Reserve; 18 - Pandivere Water Protection Area.

crop out under a thin (2-3 m) clays and marine sands. The buried Kasari Valley is 33 metres deep.

Two scarps (Salevere and Kirbla), three springs (Salevere Silmaallikad) and a great number of boulders have been included in the Book of Primeval Nature.

Nine rivers run their waters into Matsalu Bay. 94% of groundwater reserves are stored in the Silurian-Ordovician limestones. The water in the topmost aquifer system has no natural protection in elevations, beach ridges and eskers (Kink & Kaljumäe 1987).

The water monitoring system is based on profiles running through the bay and along the left bank of the Kasari River. There are altogether 21 monitoring sites, with the depth of the wells ranging from 5 to 30 metres (Kink 1991). The water of the wells in natural landscapes was practically devoid of nitrogen, while heightened concentrations (> 100 mg/l) of Cl and SO_4 were registered in the sea-side areas. In the cultural landscape, the concentration of N in the topmost aquifer was at its highest in 1980-85. The pollution of water is decreasing.

Viidumäe Nature Reserve, 1194 ha in area, was founded in 1957 in the western part of Saaremaa Island (Fig. 260). A conspicuous topographic feature is the former coastal escarpment. Its relative height ranges from 10 to 18 m and absolute height from 16 to 51 m. The Quaternary cover consists of glacial, glaciofluvial, lake and sea deposits and is up to 30 m thick.

The foot of the escarpment is rich in springs. The groundwater gushes out at the surface as large point springs (38) or by way of seepage in a large area. The springs in the reserve are most water-abundant after the years with heavy precipitation, and have been included in the Book of Primeval Nature. Spring bogs cover some 10% of the reserve's area and support luxuriant growth of rare plant species. 17 species have been placed under protection.

Otepää Landscape Reserve, 23,031 ha in area, (Fig. 260) was founded in 1970. Large relief forms attaining a relative height of 10-20 m are surrounded by hills or eskers. In the center of the heights there is a denudationally dissected plain of Middle and Upper Devonian sandstones. The largest rivers of South Estonia start from the heights. The landscape reserve holds altogether 65 lakes. Groundwater is stored in the sands and gravels of the Quaternary cover and the bedrock. The interstratal groundwater is protected, the subsurface groundwater and surface water are poorly protected. In 1986, 20 observation sites were founded for monitoring purposes. The studies showed that in this area the water quality was better than in northern Estonia. The quality of surface water was affected by Otepää Town (Kink *et al.* 1986, Kink 1990).

Vilsandi National Park, 167 km² in area (Fig. 260), was founded in 1910 on the Island of Vaika, restored in 1957 as the Vilsandi Nature Reserve, reformed into a national park in 1993. The task of the park is to safeguard the whole ecosystem of islands with special attention to seabird and coastal plant communities.

The bedrock is formed of secondary dolomites of the Jaagarahu Stage (Aaloe *et al.* 1983). The Quaternary cover is thin and prevailingly of marine origin. Karren – small holes and cracks formed by precipitation and sea water – is a common phenomenon on the exposed surface of dolomites on many islands (Photo 80). However, it is also found as deep as 4 - 5 metres below sea level.

Karula National Park, 103 km² in area, was formed as a landscape reserve in 1979, and reformed into a national park in 1993 (Fig. 260). The area exhibits a picturesque landscape of morainic hills and kames with 33 interspersed lakes.

Soomaa National Park, 367 km² in area, was founded in 1993. It is situated on the boundary of Upper and Lower Estonia in the catchment area of the Navesti, Halliste and Raudna rivers (Fig. 260). The catchment area of the rivers within the Soomaa National Park totals 4000 km². The national park consists of the Kikepera, Ördi, Kuresoo and Valgeraba mires.

In 1993, geological conditions and the chemical composition of water were studied in particular detail in the area (Kalm *et al.* 1994). The upper part of the bedrock is composed of sandstones and clays of the Narva Stage, which are overlain by till, glaciofluvial and glaciolacustrine deposits with a total thickness of 5-30 m.

The mires, within the national park (except Valgesoo) formed as a result of the filling-up of lakes. There is a layer of gyttja under the peat stratum which isolates the mire water from the deeper-lying groundwater. The hydrographic network of mires responds easily to human activities. In the Valgesoo Mire, which formed as a result of paludification of sandy soils, the water is slightly or moderately protected. Water is unprotected in the catchment areas.

Nigula Nature Researve, 27.71 km² in area, was founded in 1957. This peatland (Fig. 260) in the transition area between the Sakala Upland and the Pärnu Lowland is a mire typical of the southwestern part of Estonia. The upper part of the bedrock is composed of sandstones of the Devonian Burtnieki and Aruküla stages. The sandstones are overlain by till (1-2 m), glaciolacustrine and marine sediments. The mire formed as a result of overgrowing of a Late-glacial lake. As the result, gyttja (20 cm) underlies the peat layer (7 m). Buried drumlins project above the mire surface as mire islets. Since the mire water is isolated from the groundwater stored in mineral soils, its natural protection is good (Kink *et al.* 1996).

The bog water is acid (pH 4.7), the lake water is slightly acid. The water in bog-hollows is richer in Cl, SO_4 and Ca than in bog-pools. The Nigula Nature Reserve is the most important water reservoir in the southern part of the Pärnu County (Kink 1993).

Endla Nature Reserve, 81.62 km² in area, was founded as a mire reserve in 1981, and reformed into the Endla Nature Reserve in 1985 (Fig. 260). In 1910, a peatland experimental station was opened at Tooma. It was the first of the kind in the Russian Empire into which Estonia at that time belonged. At the station, attention focused on peatland amelioration. In 1950, a hydrometeorological station was put into operation in the mire.

The Endla Nature Reserve is a typical mire system with abundant pools. The peat layer is 3-4, occasionally 7.3 metres thick. The aim of the reserve is to preserve the ecosystem and safeguard the water resources of the Pandivere Water Protection Area. The largest (Norra) and deepest (Sopa) springs of Estonia are located within the Endla Nature Reserve.

Alam-Pedja Reserve, 270 km² in area, was founded as a wetland and forest reserve on the flooded area on the lower courses of the Pedja and Põltsamaa rivers (Fig. 260). All landscape types of Ancient Võrtsjärv are represented in the reserve: mires, coastal formations and insular heights. The nature reserve serves as a zone of protection to Võrtsjärv which is one of the two major lakes in Estonia.

Kurtna Landscape Reserve, 25.41 km² in area, was founded in the northeastern part of Estonia in 1987. The reserve holds complexes of natural ecosystems which have been most heavily affected by human activities in Estonia.

The basic geomorphological unit of the landscape reserve is the Kurtna Kame Field, featured by hillocks, small ridges and mires. The kame field is dotted with 40 lakes in glaciokarst depressions. The Kurtna Landscape Reserve (Fig. 260) lies between the densely populated and heavily industrialized oil shale mining region and sparsely inhabited territory with big forests and mires of Alutaguse. The reserve borders on mines, peat milling fields and numerous plants. Due to the extremely complicated ecological situation in the region, complex studies were carried out in the area during 1985-94 (Punning 1994).

Kõrvemaa Landscape Reserve, 217.20 km² in area, was founded in 1957, expanded in 1971 and 1991. Until 1990, the northern part of the reserve was controlled by military forces of the former Soviet Union. In 1990, when the access to the area became easier, researchers of the Institute of Geology undertook geological studies in the northern part of Kõrvemaa (Fig. 260).

Kõrvemaa is located in the outcrop area of carbonate rocks. The upper surface of the bedrock is dissected by ancient valleys, the contemporary relief is diversified by morainic hills, eskers and glaciofluvial deltas. The area abounds in peat bogs and picturesque lakes - relics of large post-glacial lakes. The present-day Kõrvemaa is a forest landscape, which provides habitats for 650-700 plant species. 15 species are under protection.

Paganamaa Landscape Reserve, 11.07 km² in area, was founded in the southeastern part of Estonia in 1979 to preserve glacial relief forms and lakes.

Vooremaa Landscape Reserve, 99.0 km² in area, was founded in the central part of Estonia in 1946 to preserve typical drumlin landscape with numerous marshes and lakes. The complex monitoring area at Saadjärv was founded in 1994 (Keskkonnaseire 1995).

Meenikunno Mire Reserv, 17.57 km² in area, was founded in southeastern Estonia in 1981, when a network of 30 mire reserves was created. It includes the Meenikunno Mire with sand ridges and bog islands. Peculiar water absorption funnels occur on the boundary of the mire and mineral soil area.

West-Estonian Archipelago Biosphere Reserve. This programme area (15,600.74 km²) was founded in 1990 on the islands of western Estonia where a complex preservation strategy of the territory was established. The reserve covers also a large part of the Baltic Sea and includes numerous islands (Saaremaa, Hiiumaa, Vormsi, Muhu, Ruhnu) and islets. The area is rich in geological monuments, including karst areas, springs, big boulders and geological outcrops, which have been or will be included in the Book of Primeval Nature.

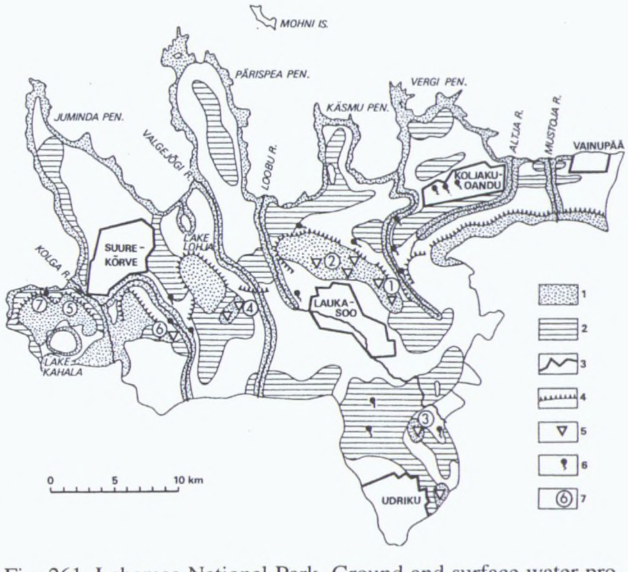


Fig. 261. Lahemaa National Park. Ground and surface water protection: 1 - unprotected area; 2 - weakly protected area; 3 - boundary of nature reserves; 4 - klint; 5 - karst forms; 6 - spring; 7 - water protection areas.

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Trükitud Tallinna Raamatutrükikojas

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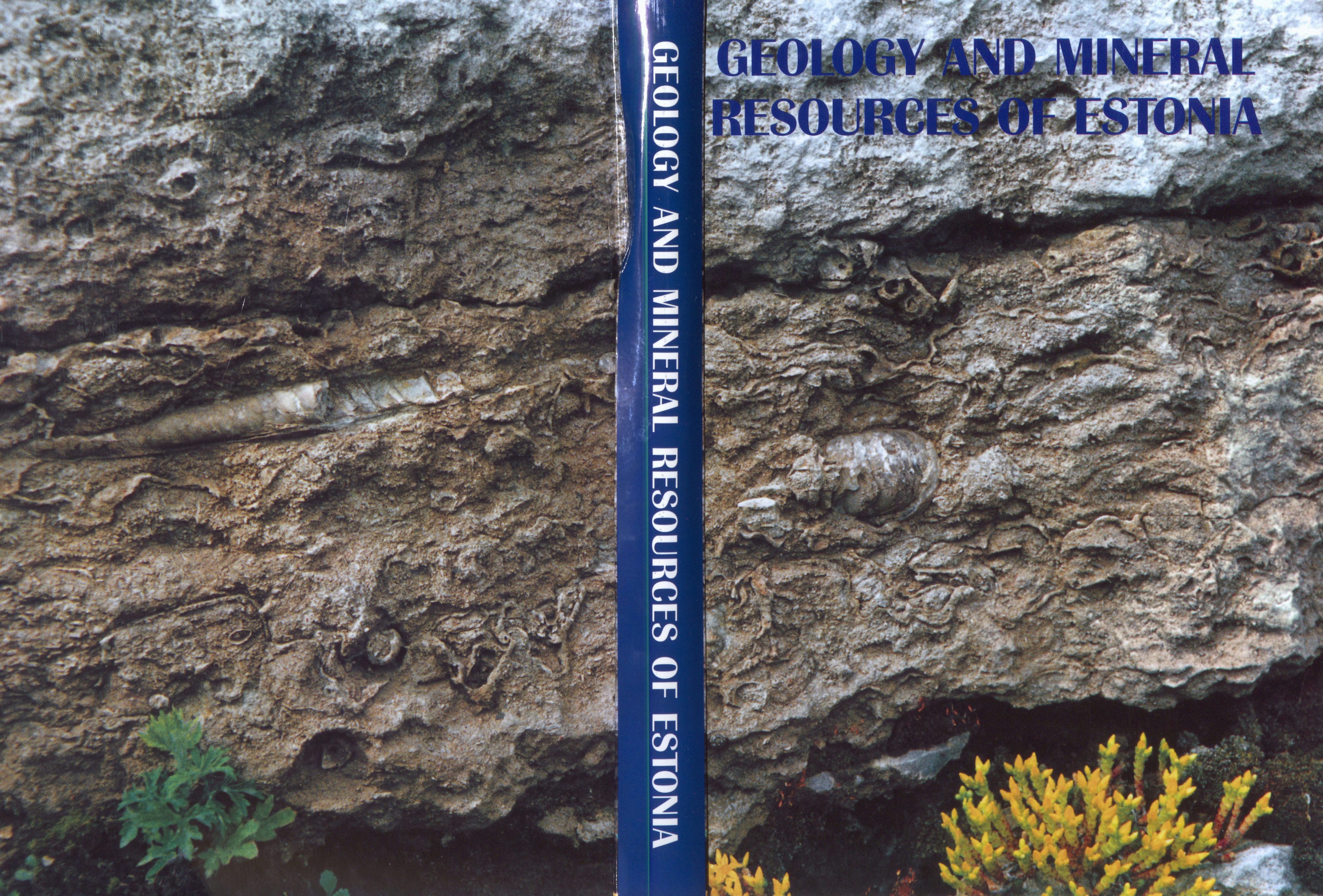

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