LAKE PEIPSI

GEOLOGY


NARVA
Laagna
Vaivara
Toila
Purtse
Savala
Vasavere
Kurtna
Illuka
Konsu L.
Puhatu Mire
Narva R.
Plyussa R.
Ohakvere
Tarumaa
Mäetaguse
Raudi
Kuremäe
Purtse R.
Ojamaa R.
Virunurme
Metsküla
Muraka Mire
Oonurme R.
Iisaku
Varesmetsa
Kamarna
Agusalu
Kuningaküla
Gorodenka Br.
Omuti
Zagriv'e
Kondushi
SLANTSY
Radovel'
Gostitsy
Permisküla
Kirsakov Mire
Jaama
Pootsiku
Alajõgi R.
Roostoja R.
Tudulinna
Sälliku
Alajõe
Remniku
Vasknarva
Vtroya R.
Rannapungerja R.
Lemmaku
Avijõgi R.
Avinurme
Kauksi
Uusküla
Rannapungerja
Smolnitsa
Skam'ya
Hitovo
Kozlov Bereg
Kozlovshchina
Dobruchi
Lohusuu
Kalmaküla
MUSTVEE
Raja
LAKE PEIPSI
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Vaiatu
Torma
Omedu
Rääbise
GDOV
Cherma R.
Gdovka R.
Sääritsa
Ranna
Omedu R.
Vassevere
Toljase
Assikvere
Sassukvere
Kodavere
Gorodishche
Beshkino
Varbevere
Alnavere
Nautrasi
Vetvenik
Torila
Kozlovo
Palamuse
Pala
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Kuna R.
Kokora
Savastvere
Nina
Torokhovo
Välgi
Alatskivi
Lahepera

Lahepera L.
Koosa
Varnja
Klenno
Spitsyno
Zalakhtov'e
Raskopel'
Podlip'e
Mda
Yamm
Praaga
Koosa L.
Kõrveküla
Ostrovtsy
Meerapalu
Piirissaar I.
Podolesh'e
Kavastu
Kalli L.
TARTU
Luunja
Emajõgi R.
Leegu L.
Podborov'e
Zhelcha R.
Remda
Parapalu
Samolva
Võnnu
Agali
Jõepera
Chudskaya Rudnitsa
Chernaya R.
Lochkina R.
Lääniste
Ahja R.
Mehikoorma
Aravu
Pnevo
Put'kovo
Meeksi
Homutovo
Lipno
Kõnnu
Baglitsy
Mooste
Osotino
Mtezh
RÄPINA
Kiidjärve
Meshokol
Tolba R.
Berez'e
LAKE
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Lüübnitsa
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Drislivik
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Treski
Lutepää
Võhandu R.
Molozhva
Nedsaja
Sesniki
Saatse
Vidovichi
Vagula
Pechki
Kalki
Piusa R.
Obdekh R.
VÕRU
PECHORY
PETSERI
Vashina-Gora
PSKOV
PIHKVA
Senno
Vastseliina
Mustjõgi R.
Rõuge
Izborsk
Irboska
Viitina

LAKE PEIPSI

I GEOLOGY

Institute of Geology at Tallinn Technical University

LAKE PEIPSI

GEOLOGY

Compiled and edited by Avo Miidel and Anto Raukas

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Drawings by Paul Pärkma
Layout by Jüri Nemliher
Cover photos by Jaak Nõlvak and Avo Miidel

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INTRODUCTION

Lake Peipsi with its surface area of 3555 km^2 is the fourth largest inland body of water in Europe after Ladoga (18,135 km^2) and Onega (9720 km^2) in Russia and Vänern (5385 km^2) in Sweden.

The coasts of this rich-in-fish lake were inhabited as early as the Mesolithic (the middle of the eighth up to the fourth millennium BC). Peipsi was part of the waterway the merchants of the Hanseatic League used to reach Pskov in Russia. Over many centuries it was an arena of bloody battles and sharp ideological conflicts between different nations.

Peipsi is one of the best-stocked lakes in Europe. Some 95 per cent of the total fish catch in Estonia's inland waters comes from L. Peipsi. Catches over 40 kilograms per hectare are not rare, and the stock of valuable fish, including pike-perch, whitefish, pike, bream, eel, *etc.*, is increasing. The management and regulation of the lake's ecosystem demands profound knowledge of the mechanisms operating in this system.

The lake basin holds over 25 km^3 of relatively clean water and is a promising fresh-water reservoir for the towns of northeastern Estonia and Tallinn. Peipsi supplies water for the boilers of the Estonian and Baltic Thermal Power Stations and drives the turbines of the Narva Hydroelectric Power Station.

The collapse of the Soviet Union in 1991 brought about great changes in the land use and environmental policy. In Estonia and Latvia, the huge state farms were liquidated and small private farms were established. The process is still ongoing. The recent restructuring of economy in the Baltic States and the diminished use of agrochemicals in the Baltic States and Russia have triggered positive trends in environmental management. In recent years, the production of waste water has decreased, the pollution load has reduced and several new efficiently operating sewage treatment facilities have been constructed. As a result, the water quality in the rivers flowing into L. Peipsi has remarkably improved and anthropogenic stress on the lake is lowering.

The maintenance of water purity in L. Peipsi is of utmost importance. In this connection, research into the land structure, landscapes, bottom and coastal deposits of the lake and the prognosis of the lake's geological development have become topical issues.

L. Peipsi has a meridionally elongated basin. Its northern part, which was situated closer to Fennoscandia, is rising, the central part is stable and the southern part is sinking. Thirty rivers, the largest being the Velikaya, the Emajõgi and the Gdovka, run their waters into L. Peipsi. The outflow is only through the Narva River in the rising northeastern part of the lake basin. The mouth of the river has to be regularly cleaned from sandy sediments deposited by longshore drift. The runoff through the Narva River is 5950 m^3 yr^{-1}.

At the present time, the northern part of the lake is rising at a rate of some 0.2-0.4 mm yr^{-1}, while the southern part is sinking at a rate of 1.2 mm yr^{-1}. In the past, the differences between these values were much greater. The above-named processes have caused and are still causing the water to spread from north to south. As a result, vast stretches of lowland on the west coast of L. Peipsi and on the coasts of L. Pihkva are overflooded or embogged. The largest floods occur at the mouth of the Emajõgi River, on the east coast of L. Lämmijärv and along the whole length of L. Pihkva's low coast. The water level in Värska Bay has risen by 10 metres in the course of the last 8000 years.

Seasonal lake level fluctuations are great and exceed 3 metres. The lowest water stand was registered on November 7, 1964 (28.72 m a.s.l.) and the highest on May 12, 1924 (31.76 m a.s.l.). Lake-level fluctuation causes problems in fishery and water transport.

During Soviet time, a great quantity of pollutants from industrial and agricultural sources found their way into the lake. They deteriorated the quality of water and stimulated exuberant growth of reed and bulrush in the nearshore areas. In making long-term prognosis, one should also consider the uneven neotectonic uplift of the Earth's crust and the low spreading of lake water to the south.

Peipsi with a great variability of shore types and favourable bathing places offers good recreational possibilities. Attractive sandy beaches exposed to the southern sun, beautiful pine woods on dunes protecting the lake from northerly winds, good transport-geographical conditions conducive to convenient access, numerous shops and canteens, *etc.*, have made the northern coast popular area among holiday makers. Of special importance is the occurrence of mineral water and curative mud in the southern part of the lake basin.

The importance of L. Peipsi in different fields of economy and its inestimable aesthetic value for holiday makers binds us to careful observation and protection of its natural condition. The lake and its surroundings have been studied since long. The first scientific studies were started after the catastrophic inundations in 1840 and 1844 which caused great damage to adjacent areas. As the fish-breeding in the lake needed some regulation, the first studies were mainly in the field of hydrology and biology of the lake. The first data on the lake's bottom deposits date from the end of the last century. Since then, the problems relating to the structure and geological development of the lake basin have been discussed in many publications.

A network of hydrological stations was developed in the 1920s. Since 1964, complex annual hydrobiological monitoring, and since 1982, monitoring of the lake shores have been carried out. In the 1980s, foundation was laid to large-scale complex research into the lake which involved tens of different institutions.

The Soviet system stimulated close cooperation between Estonian and Russian researchers within extensive, long-term and rather expensive complex projects. Owing to these projects, L. Peipsi is among the most thoroughly and consistently studied lakes in the world. Unfortunately, for political reasons, the results arising from those projects could be published only in Russian and Estonian, and remained unavailable to the international scientific community.

In 1992, Estonia's Russian frontier on the lake was re-established. The cooperation of Estonian and Russian agencies and researchers in environmental monitoring of the lake was discontinued. The monitoring (hydrology, hydrochemistry, phyto-, zoo- and bacterioplankton, macrozoobenthos, radioactive substances and the state of coasts) is ongoing only in the Estonian part of the lake.

A number of bilateral agreements have been signed to regulate transboundary problems between the Republic of Estonia and the Russian Federation. This holds out the hope that cooperation between the researchers of the two countries will continue in the future.

An agreement on border-crossing points on the customs border was signed by the two governments in Moscow on July 9, 1993. On May 1995, an intergovernmental agreement on the protection and regulation of the use of fish stock of the lake was concluded in Pskov. On November 20, 1996, an intergovernmental agreement on the activities of border representatives, and on August 20, 1997, an intergovernmental agreement on the protection and sustainable use of transboundary water bodies were signed.

The most important step in the process of developing cooperation on the protection and sustainable use of the transboundary waters between Estonia and Russia was the establishment of the Joint Commission on Transboundary Waters with the permanent working groups on Water Economy; on Water Protection; on Monitoring and Research; on Cooperation with International and Non-Governmental Organizations and Local Authorities. The first official meeting of the Commission headed by the Chancellor of the Estonian Ministry of the Environment and the Deputy Minister of the Ministry of Natural Resources of the Russian Federation took place in Tallinn on May 19, 1998.

As there are currently no published general surveys on L. Peipsi, the present three-volume monograph (I - Geology, II - Hydrochemistry, III - Flora and Fauna) attempts to fill this gap. Its main purpose is not only to impart scientific information about the lake and its surroundings, but to encourage also the optimal use of the natural resources of the lake and to protect its beauty.

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Avo Miidel and Anto Raukas

1. LOCATION AND TOPOGRAPHY

1. 1. LAKE PEIPSI AND ITS CATCHMENT AREA

Lake Peipsi (3555 km^2), which is situated south of the Gulf of Finland on the border between Estonia and Russia (Fig. 1), ranks fourth in size among the European lakes. The name *Peipsi* comes evidently from the language of ancient tribes who inhabited the area before the Finno-Ugrians arrived. The Russian *Chudskoe ozero* (the lake of *chud*s, the lake on the land of *chuds*) is probably derived from the name *chud*' or *chudy*. It's how the Slavish tribes, who reached the coast of L. Peipsi later, about the middle of the first millennium AD, used to denote the local Finno-Ugric tribes.

The submeridionally elongated catchment basin of L. Peipsi (Fig. 1) measures approximately 160 km in width and 370 km in length, and is shared between Estonia, Russia and Latvia. It is part of the Gulf of Finland's basin and is connected with the latter via the rather short Narva River (77 km) (Fig. 1). The catchment basin of the lake (including the lake surface) has an area of 47,800 km^2, of which 16,323 km^2 is in Estonia, 27,917 km^2 in Russia and 3,560 km^2 in Latvia. It is bordered by the drainage areas of the rivers Plyussa, Shelon' and Lovat' in the east, the Daugava (Zapadnaya Dvina) in the south and the Gauja (Koiva) and the Pärnu in the west. The catchment basin of L. Peipsi, which extends from 59°13' to 56°08'N and from 25°36' to 30°16'E (Fig. 2), is a generally gently undulating glaciolacustrine or till covered plain. The altitude of 100–120 m in its marginal areas lowers down to 30–40 m towards the lake. The border of the catchment area is marked with several uplands: Luga (max altitude 204 m) in the upper course of the Zhelcha and Keb' rivers in the east, Sudoma (294 m) on the southern margin of the drainage network of the Velikaya River, Bezhanitsy (338 m) and Vyaz (277 m) in the south-east. The Pandivere Upland is located in the northern and the Haanja and Sakala uplands in the western part of the catchment.

The catchment area of L. Peipsi belongs to the forest

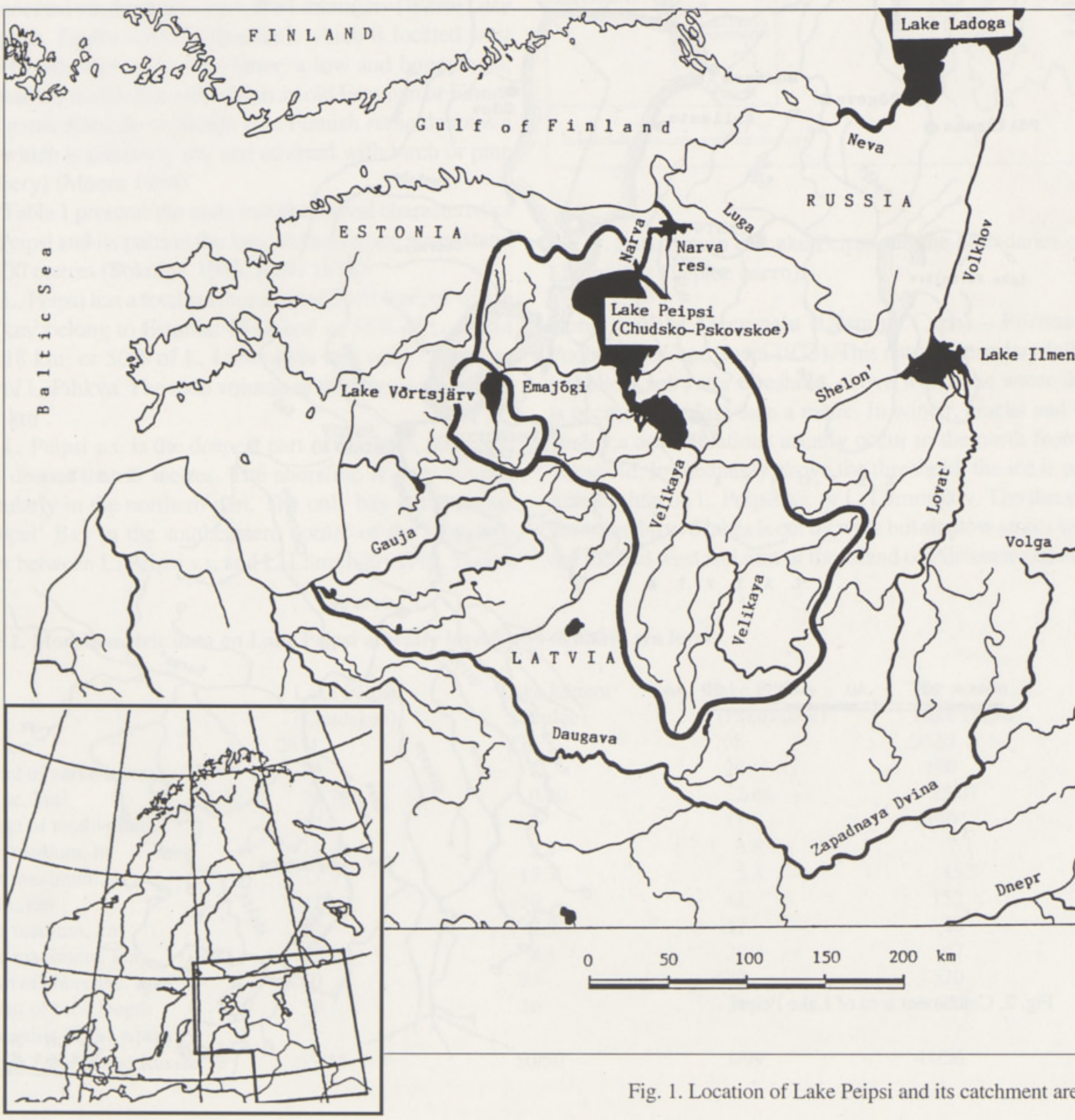

Fig. 1. Location of Lake Peipsi and its catchment area

zone of the East Baltic Geobotanical Subprovince (Laasimer 1964). The proportion of forested land is highest (60–70%) in NW at Alutaguse, where the pine forests prevail. Towards the south, the forested area decreases and the proportion of broad-leaved trees increases. In the southern part of the catchment basin the forests cover only 30–40% of the area; the proportion of deciduous trees is high. Mires and bogs make up 10% of the catchment area. Together with bogged forests and meadows they account for 15–20% of the excessively damp land. The proportion of mires is highest in the drainage areas of the rivers flowing in the northern part of the region: Põltsamaa — 25%, Rannapungerja — 20% and Gdovka — 15% (Sokolov 1983). In the drainage area of the Velikaya River, mires cover only 2% of the land area. The proportion of arable land is highest (60–70%) in the western and southern, and lowest (20–30%) in the northern part of the catchment basin.

The catchment basin of L. Peipsi holds more than 4500 lakes, the largest of which is Võrtsjärv (270 km^2), while all the other lakes are less than 1 km^2 in area. There are about 240 inlets into L. Peipsi. The largest rivers are the Velikaya (catchment area 25,200 km^2), the Emajõgi (9745 km^2), the Võhandu (1423 km^2) and the Zhelcha (1220 km^2) (Fig. 2), which all together form about 80% of the whole L. Peipsi's

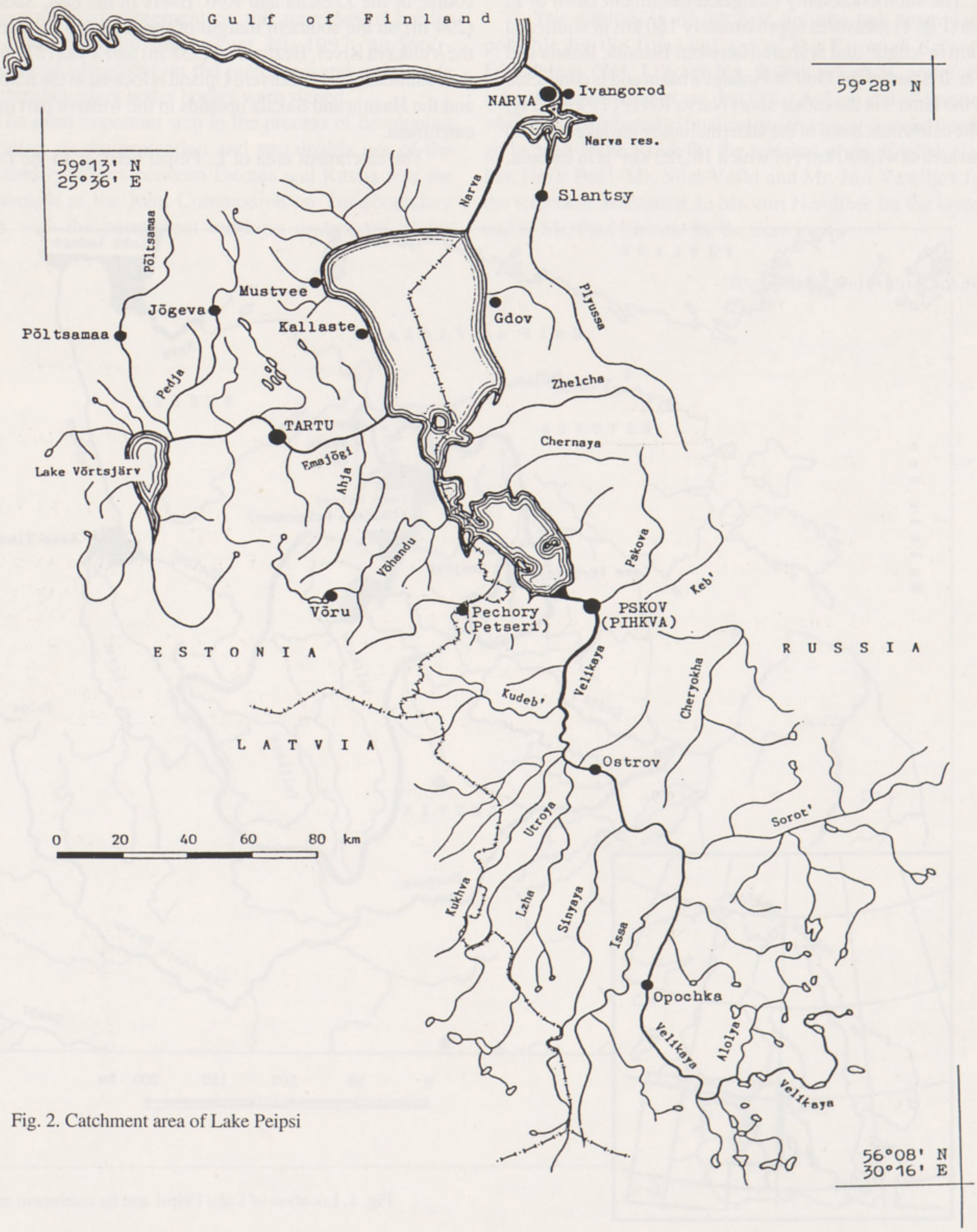

Fig. 2. Catchment area of Lake Peipsi

catchment basin and account for 80% of the total inflow controlling, thus, the water regime of the lake. The other bigger rivers include the Piusa (796 km^2), the Rannapungerja (601 km^2), the Chernaya (530 km^2), *etc.* The Velikaya River has a length of 430 km, three other rivers are over 100 km and 25 rivers over 10 km in length. Some 88% of the watercourses flowing into L. Peipsi are rivulets and ditches, less than 10 km in length. The only outflow is through the Narva River, which starts from the northeasternmost corner of the lake and takes its waters into the Gulf of Finland.

The annual mean runoff through the Narva River is more than 12 km^3. The state of the lake is affected by forests, agricultural lands, townships and the industrial region in its north, each of which has its specific influence on the lake.

Despite the large surface area, L. Peipsi is a shallow body of water with a rather simple configuration of the coastline. L. Peipsi, or rather the lake system, consists of three parts, each bearing a name of its own (Fig. 3). The northernmost, the largest and deepest part is Lake Peipsi *sensus stricto* (*Chudskoe ozero* in Russian), which is sometimes called Suurjärv (*Big Lake*). The southernmost part is Lake Pihkva (*Pskovskoe ozero* in Russian). These two lakes are connected via the strait-lake Lake Lämmijärv (*Warm Lake* in English, *Teploe ozero* in Russian), which is located west of the Remda Peninsula. The latter, a low and boggy area, has evidently derived its name from an old Estonian or Finno-Ugric name *Rämeda* or *Rämja* (the Finnish *räme* denotes a mire, which is relatively dry and covered with birch or pine shrubbery) (Moora 1964).

Table 1 presents the main morphological characteristics of L. Peipsi and its parts at the long-term average water stand of 30.00 metres (Sokolov 1983, Jaani 1973).

L. Peipsi has a total surface area of 3555 km^2, of which 1570 km^2 belong to Estonia: 1442 km^2 or 55% of L. Peipsi *s.s.*, 118 km^2 or 50% of L. Lämmijärv and only 10 km^2 or 1.3% of L. Pihkva. The total volume of the lake as a whole is 25.07 km^3.

L. Peipsi *s.s.* is the deepest part of the lake, 80% of it being deeper than 6 metres. The shoreline is very simple, particularly in the northern part. The only bay is the scenic Raskopel' Bay in the southeastern corner of the lake. The border between L. Peipsi *s.s.* and L. Lämmijärv (Fig. 3) runs

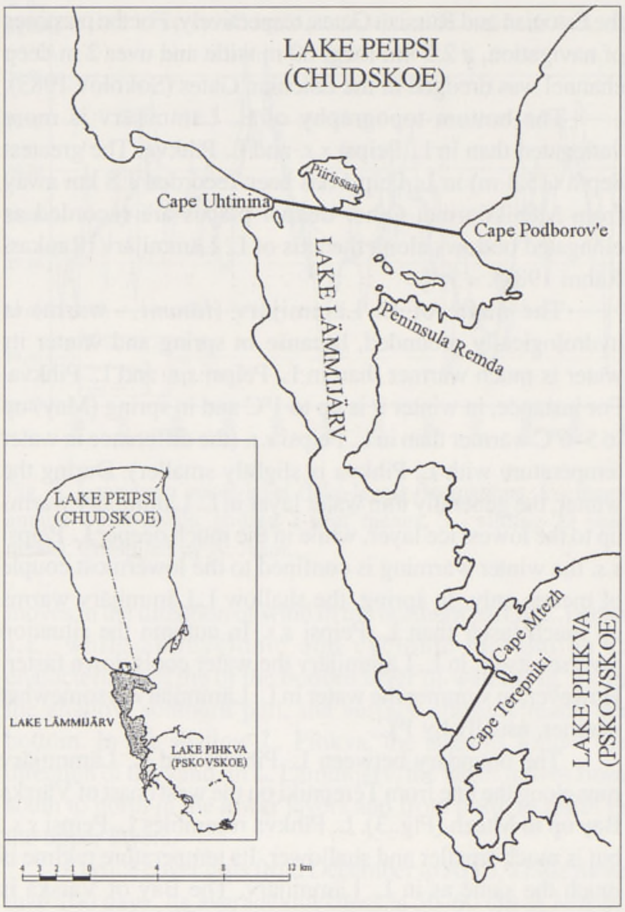

Fig. 3. Subdivision of Lake Peipsi and the boundaries of L. Lämmijärv (Teploe ozero).

along the line Meerapalu (Uhtinina Cape) – Piirissaar – Podborov'e Cape (Jaani 1973). This natural boundary follows a shallow subwater threshold, above which the water-depth is occasionally less than a metre. In winter, cracks and long push ice accumulations usually occur to the north from the threshold. Immediately above the threshold, the ice is much thinner than on L. Peipsi *s.s.* or L. Lämmijärv. The threshold between the two lakes is cut by wide but shallow straits which are located west and east of the Island of Piirissaar and called

Table 1. Morphometric data on Lake Peipsi at water level by 30 m above sea level.

	Lake Peipsi (Chudskoe)	Lake Lämmi (Teploe)	Lake Pihkva (Pskovskoe)	The whole Lake Peipsi
Area, km^2	2611	236	708	3555
Per cent of surface area	73	7	20	100
Volume, km^3	21.79	0.60	2.68	25.07
Per cent of total volume	87	2	11	100
Depth medium, m	8.3	2.5	3.8	7.1
Depth maximum, m	12.9	15.3	5.3	15.3
Length, km	81	30	41	152
Width medium, km	32	7.9	17	23
Width maximum, km	47	8.1	20	47
Length of shoreline, km	260	83	177	520
Per cent of total length	50	16	34	100
Distribution of the aquatory between Estonia and Russia, %	55/45	50/50	1/99	44/56

the Estonian and Russian Gates, respectively. For the purposes of navigation, a 2.5 km long, 60 m wide and over 2 m deep channel was dredged in the Estonian Gates (Sokolov 1983).

The bottom topography of L. Lämmijärv is more variegated than in L. Peipsi *s.s.* and L. Pihkva. The greatest depth (15.3 m) in L. Peipsi has been recorded 1.8 km away from Mehikoorma. Other deeper places are recorded as elongated hollows along the axis of L. Lämmijärv (Raukas, Rähni 1981).

The name of L. Lämmijärv (*lämmi* - warm) is hydrologically grounded, because in spring and winter its water is much warmer than in L. Peipsi *s.s.* and L. Pihkva. For instance, in winter it is up to 1°C and in spring (May) up to 5–6°C warmer than in L. Peipsi *s.s.* (the difference in water temperature with L. Pihkva is slightly smaller). During the winter, the generally thin water layer in L. Lämmijärv warms up to the lowest ice layer, while in the much deeper L. Peipsi *s.s.* the winter warming is confined to the lowermost couple of metres only. In spring, the shallow L.Lämmijärv warms up much faster than L. Peipsi *s.s.* In autumn the situation reverses itself: in L. Lämmijärv the water cools down faster. However, in summer the water in L. Lämmijärv is somewhat warmer, usually by 1°C.

The boundary between L. Pihkva and L. Lämmijärv runs along the line from Terepniki on the west coast of Värska Bay up to Mtezh (Fig. 3). L. Pihkva resembles L. Peipsi *s.s.*, but is much smaller and shallower. Its temperature regime is much the same as in L. Lämmijärv. The Bay of Värska is situated in the northwestern corner of L. Pihkva.

L. Peipsi has 35 islands and islets. In additon to those there are about 40 islets at the mouth of the bigger rivers (the Velikaya – 40, the Võhandu – 3, the Emajõgi – 2, the Logina (Koosa) – 2, and the Samolva – 2). The total area of islands and the islets is 29 km^2, which forms 0.8% of the surface area of the lake. The islands are concentrated in the southern part of the lake - in L. Pihkva and L. Lämmijärv (in L. Peipsi *s.s.* small islets occur only in Raskopel' Bay). The biggest islands are Kolpino (11.1. km^2), Piirissaar (7.5 km^2) and Kamenka (4.0 km^2) near the western coast of L. Pihkva (Fig. 2). The Talabskij Archipelago, which consists of three islands and is situated off the east coast of L. Pihkva, is a peculiar phenomenon; it was formed along the margin of the ice tongue which penetrated into L. Pihkva basin during the Otepää Stade of the last glaciation (Photo 1). The archipelago consists of Talabek or Zalit, Verhnij or Belov islands with the small Talabenets Islet between them. Talabek Island is isolated from the mainland by the Beregovoij Ksush Strait. Talabek and Talabanets are separated by Talabanetsk Ksushi Strait, and Talabanets and Verhnij by the Vorota or Zheroglya Strait (Sokolov 1983).

The water-level fluctuations in L. Peipsi are considerable. During the course of the last 80 years, an amplitude of 3.04 metres has been registered. The average water level is 30 metres above sea level. The lowest water level of this century was registered on November 7, 1964 (28.72 m a.s.l.) and the highest (Photo 2) on May 12, 1924 (31.76 m a.s.l.). A distinct rhythmicity (Fig. 4) is observed in lake-level fluctuations (Jaani 1973).

Considerable water level fluctuations in L. Peipsi cause changes in both the surface area and volume of the lake. During the highest (31.76 m) water-level, its surface area was estimated at 4330 km^2, and the volume of water at 32.128 km^3. At the lowest water-stand, these values were 3480 km^2 and 20.98 km^3, respectively. Thus, the surface area of the lake may vary by the value of 850 km^2 and the water volume by 11.15 km^3.

The southwesterly and southerly (45–50%) winds predominate in the Peipsi Basin causing water level changes in the northern part of the lake, where the width of the beach ranges from 5 to 15 metres. The coastal slope is relatively gentle and even a half-a-metre rise in the water level may trigger intensive erosion of the coastal formations. In the western part of the lake, the wind frequently blows from the west and in the eastern part from the south, west and north-

Photo 1. An island in the Talabskij Archipelago – a relict of an end moraine which is subject to the effect of the lake's advance. Photo by E. Rähni.

Photo 2. Flood at Vasknarva on May 21, 1924. Photographer unknown.

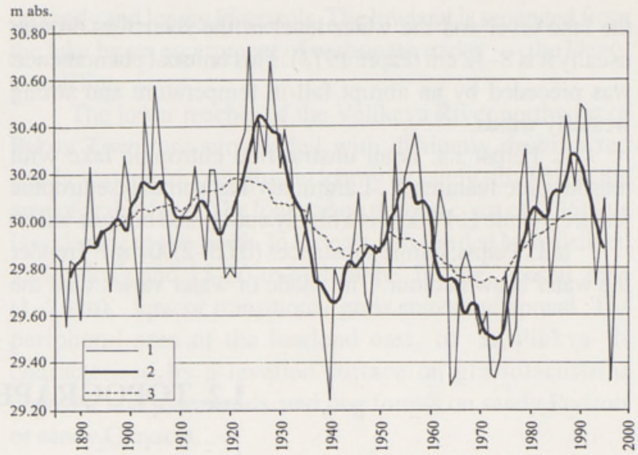

Fig. 4. Long-term water-level (lake-level) fluctuations. 1 - mean annual water-levels, sliding 7-year means, 3 - sliding 23-year means. Compiled by A. Jaani.

west. The average wind-speed is 4–5 m s^{-1}. The maximum speed — 20 m s^{-1} — was recorded at Tiirikoja when the wind was blowing from the west-southwest and 24 m s^{-1} at Gdov when the wind was blowing from the west-northwest. The strongests winds blow in October, while the calmest month is July.

Waves in L. Peipsi are steep and short. When the wind blows with the speed 8 m s^{-1}, the waves are 60–70 cm high (Sokolov 1983). Waves of that height are most common in L. Peipsi (57%). The highest waves (240 cm) were recorded in 1961 and 1962, when the wind-speed reached 20 m s^{-1}. Wind drift, wind gradient (compensation), flow, seiche and internal pressure streams (Kallejärv 1973) occur in L. Peipsi. The system of streams is different in the different parts of the lake (Peipsi *s.s.*, Lämmijärv and Pihkva). The surface water moves in the direction of wind in the northern part of L.Peipsi *s.s.*, while anticyclonic and cyclonic circulation is characteristic of the of the bottom layer of water (Fig. 5). In the shallow southern part, the surface currents reach the bottom. In the shallow L. Pihkva, the streams follow the direction of the wind. In L. Lämmijärv, the water moves from south to north in the lower layers and from north to south in the upper layers.

The ice cover lasts from December to April, a little more than 100 days. Its average thickness is 50–60 cm in March. Autumn-winter ice drift and abundant formations of frazil ice are common phenomena on the Narva River. In November 1971, an extraordinary ice jam formed at the outflow of the Narva River near the village of Vasknarva and blocked the water flow from the lake. As a result, the difference between

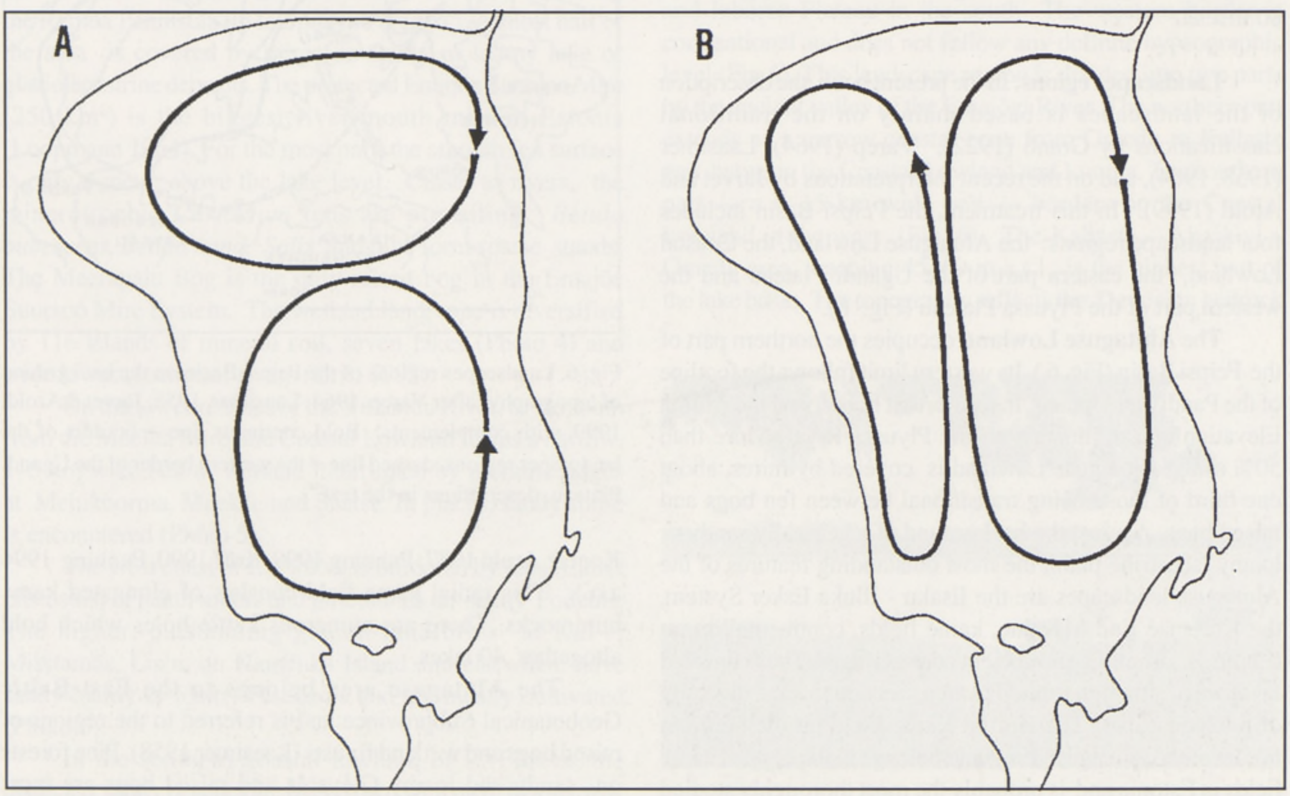

Fig. 5. System of currents in the bottom layer of water in L. Peipsi with the prevailing W-E (A) and N-S (B) winds. Compiled by T, Filatova & A. Jaani.

the lake level and the water level in the river was 64 cm; usually it is 8–12 cm (Eipre 1973). This unusual phenomenon was preceded by an abrupt fall in temperature and strong westerly winds.

L. Peipsi *s.s.* is an unstratified eutrophic lake with mesotrophic features, L. Lämmijärv has some dyseutrophic features, while L. Pihkva is a highly eutrophied body of water.

In L. Peipsi, humic substances (0.05–2.00 mg l^{-1}) render the water brown colour. The shade of water varies with the parts of the lake, changing from brownish and yellowish-brown in L.Pihkva and L.Lämmijärv to greenish or even greenish-blue in L. Peipsi *s.s.*.

The content of dissolved mineral salts in water is relatively low, usually 150–350 mg l^{-1}. According to the hydrochemical classification, the water of L. Peipsi belongs to the hydrocarbonate type; the chief constituents HCO_3^- and Ca^{2+} are followed by Mg^{2+} and SO_4^{2-}, while Na^+ ja Cl^- are of secondary importance. The lake has a good oxygen regime.

1.2. TOPOGRAPHY AND LANDSCAPES

Lake Peipsi is located in a flat lowland area which has no distinct boundaries and, therefore, the definition of the lake basin is complicated. In topography, the 50 m isoline can be taken formally as a boundary, but this would cause difficulties in identifying the limits of the lake basin in the north and west (Fig. 6). There is no geological boundary for the lake basin either, because the main lithological boundary between the Ordovician and Silurian limestones and the Devonian sandstones crosses the basin. The distribution of ice marginal formations and the character of Quaternary deposits within the basin do not notably differ from those observable at some distance from the lake.

In view of the above, the boundaries suggested for the Peipsi Basin are conventional rather than fixed by topography. A reasonable way of definition is on the basis of the catchment area of the lake (Järvet & Arold 1999) limiting oneself to its lower central and northern parts (Fig. 6). The altitudes of the contemporary topography within the below-described conventional borders of the basin are between 30 and 80 m, and only the highest tops of single landforms rise higher than 80 m a.s.l.

Landscape regions. In the present book, the description of the landscapes is based mainly on the traditional classifications by Granö (1922), Varep (1964), Laasimer (1958, 1964), and on the recent interpretations by Järvet and Arold (1999). In this treatment, the Peipsi Basin includes four landscape regions: the Alutaguse Lowland, the Coastal Lowland, the eastern part of the Ugandi Plateau and the western part of the Plyussa Plateau (Fig. 6).

The **Alutaguse Lowland** occupies the northern part of the Peipsi Basin (Fig. 6). Its western limit follows the footline of the Pandivere Upland. In the north it borders on the Ahtme Elevation and in the east on the Plyussa River. More than 50% of the Alutaguse Lowland is covered by mires, about one third of those being transitional between fen bogs and raised bogs. Against the background of a generally sandy or loamy lacustrine plain, the most outstanding features of the Alutaguse landscapes are the Iisaku - Illuka Esker System, the Kuremäe End Moraine, kame fields, continental dunes (Photo 3), drumlins and ancient coastal ridges. These lineated landforms, although limited in area, account for the diversity of the landscapes. The Kurtna Kame Field on the northern border of the Alutaguse Lowland belongs to the biggest kame fields in Estonia and is probably the most thoroughly studied kame landscape (Tavast, Raukas 1982, Karukäpp 1987,

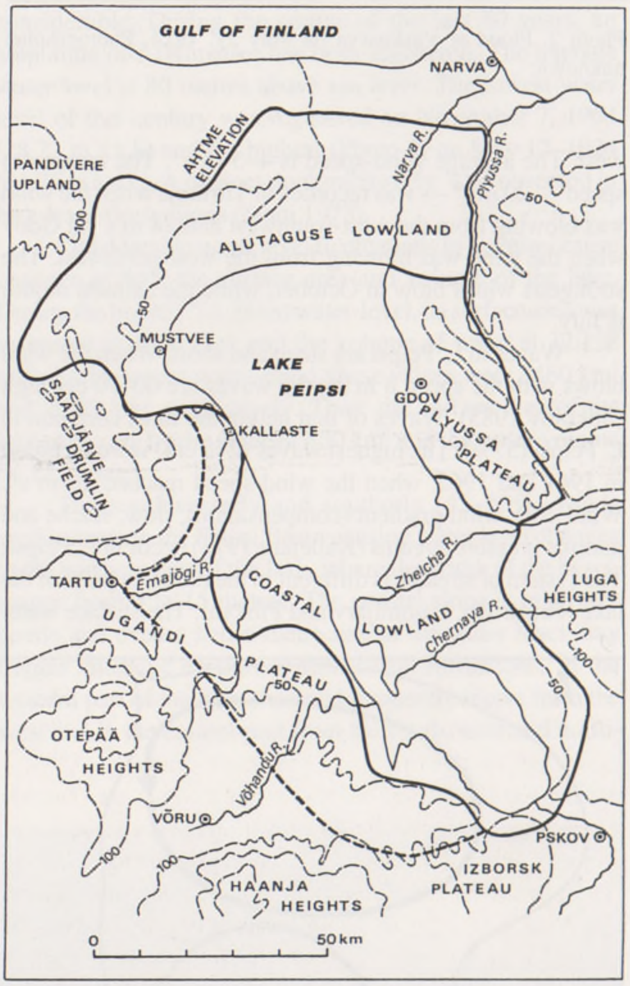

Fig. 6. Landscapes regions of the Peipsi Basin on the background of topography (after Varep, 1964; Laasimer, 1958; Järvet &Arold, 1999, with complements). Bold continous line – borders of the landscapes regions, dashed line – the western border of the Ugandi Plateau, descriptions in the text.

Kont & Arold 1987, Punning 1990, Koff 1990, Punning 1994 a.o.). This radial kame field consists of elongated kame hummocks. There are numerous kettle holes which hold altogether 40 lakes.

The Alutaguse area belongs to the East Baltic Geobotanical Subprovince and is referred to the regions of raised bogs and wetland forests (Laasimer 1958). Pine forests on sandy and loamy Gleysols and raised bogs are most widespread. The biggest mires (and mire systems) are Muraka

Photo 3. Continental dunes at Sälliku. Photo by A. Raukas.

(207 km²), Puhatu (57 km²), Kirsakov Mokh (50 km²), *etc.* Dunes and coastal ridges are covered by lichen-pine or *Calluna*- pine forests on Podzols. East of Rannapungerja, the northern and northwestern lake shore is backed by low beach ridges and dunes. The western part of the Alutaguse area at an altitude of 36–40 m is levelled, covered by mixed forests on overmoistured sandy Gleysols. The area of arable land on the loamy Cambisols and Luvisols and loamy Gleysols is in the northwestern outer part of the region and in the coastal area between the Rannapungerja River and Mustvee Town. Arable land makes up only 5% of the area, but is rather fertile (Kask 1996). As a special local type of tillage, the high-furrow cultivation is practiced in horticulture on the two-member sandy-loamy Gleysols.

Coastal Lowland. The Coastal Lowland at an altitude of 32–45 m makes up the largest part of the basin on both sides of the lake. It extends from Kallaste southwards, and around lakes Lämmijärv and Pihkva to the north including the Remda Peninsula up to Spitsyno (Fig. 6). Almost half of the area is covered by mires on sandy or loamy lake or glaciolacustrine deposits. The protected Emajõe Suursoo Mire (250 km²) is the biggest river-mouth mire in Estonia (Loopmann 1964). For the most part, the smooth fen surface is only a metre above the lake level. Closer to rivers, the minerotrophic *Caricetum* fens are prevailing. *Betula pubescens, Betula nana, Salix* and pine form sparse stands. The Meerapalu Bog is the only raised bog in the Emajõe Suursoo Mire System. The wetland landscape is diversified by 116 islands of mineral soil, seven lakes (Photo 4) and wide lower reaches of rivers (Orru 1995).

On the lower reaches of the Võhandu River, to the south from the Meelva Mire, the Coastal Lowland forms a narrow, 1–5 km wide belt of wetland interrupted by moraine ridges at Mehikoorma, Meeksi and Saatse. In places, sandy shore is encountered (Photo 5).

The west coast of L. Pihkva is bordered by fens (Eutric Histosols) or heath moors and pine stands on sandy Podzols. The higher, outstanding glacial landforms at Kul'e, Mikitamäe, Lis'e, on Kamenka Island and elsewhere have sandy-loamy or loamy Planozols and are mostly cultivated (Photo 6).

In the southern coastal lowland of L. Pihkva, the dissected sandy beaches are overgrown with reed. In a narrow belt, well-drained agricultural land alternates with brushwood on sandy and loamy Planozols. The lowland is separated from the lake by an escarpment of carbonate rocks — the Devonian Klint.

The lower reaches of the Velikaya River northwest of Pskov Town are surrounded with distinctly drumlinized gently waving topography on loamy or sandy till. This is the area of arable land. The lowland on the east coast of L. Pihkva is widening to the north. In the east, it is borded by a distinct, 35-km-long and 15–20-m-high slope. In the coastal zone (1–2 km), fens or transitional grass mires are spread. The peripheral area of the lowland east of L. Pihkva is characterised by a levelled surface on glaciolacustrine deposits, and pine stands and bog forests on sandy Podzols or sandy Gleysols.

On the Remda Peninsula (31–34 m a.s.l.), the coastal lowland on sandy glaciolacustrine and lake deposits is occupied by fens, transitional bogs and raised bogs. The monotonous levelled lowland is dissected by the Chernaya and Zhelcha rivers and their paludified flat-plains. The kame fields hold a great number of lakes. Sand in the topmost part of the kames is almost everywhere reworked by aeolian processes. The hummocky kame fields (Podlip'e, Goristoe and Znamenka) and the dune ridge systems of north-east to south-west orientation (Photo 7) are covered by pine stands on sandy Podzols. In the kame fields, deep kettle holes with mires and lakes are most typical (Photo 8). The gently sloping Pnevo – Remda – Piskopovo belt of ice marginal formations, less than 10 m in height, can be traced by the areal distribution of more fertile soils and arable land.

The Remda Peninsula with its typical kame and dune landscapes and beautiful clean-water lakes is the wildest and most picturesque part of the Peipsi Lowland.

The eastern part of the **Ugandi Plateau** borders on the Alutaguse Lowland in the north and on the Haanja Heights and Izborsk Plateau in the south The western border is conventional and does not follow any definite hypsographic level (Fig 6). This landscape region is divided into two parts by the ancient valley of the Emajõgi River. The northern part extends as a narrow coastal zone from Omedu to Kallaste and between the Coastal Lowland and Luunja. The southern part — a 6–15-km wide belt — borders on the Coastal Lowland in the west (Fig. 6). The Kallaste – Alatskivi – Omedu area, reaching 45–69 m a.s.l., is the highest part of the lake basin. The topography reflects the Devonian bedrock

Photo 4. Lake Leegu in the Emajõe Suursoo Mire. Photo by A. Miidel.

Photo 5. Sandy beach at Mehikoorma. Photo by A. Miidel.

Photo 6. Glacial marginal formations at Lis'e. Photo by A. Miidel.

elevation and the influence of drumlinization. Valleys (Sassukvere, Torila, Savastvere) have been eroded into the big drumlin-like landforms. The Kallaste Escarpment in the Devonian sandstones is the highest cliff on the lake coast (Photo 9.)

The arable sandy and loamy Planozols and Podzoluvisols on loamy till and glaciolacustrine deposits are rather fertile. The agricultural landscapes prevail in this part of the lake basin.

The southern part of the Ugandi Plateau is mostly at an altitude of 45–70 m. Here the Devonian Plateau is covered by a less-than-10-m-thick Quaternary cover (till and glaciolacustrine sand or loam), and cut by the Ahja, Võhandu and Piusa river valleys into several segments. In contrast to the levelled topography, the Mustoja Kame Field has a dissected hummocky landscape. It consists mainly of small kame hillocks, less than 10 m in height, with kettle holes between them. In the northern part, the kame field has been transformed by aeolian processes (Lutepää Dune Field in the vicinity of Värska).

The concentration of the Pontic and Pontosarmatic elements of flora in the Mustoja Kame Field and the Lutepää Dune Field suggests suitable climatic, topographic and soil conditions, and migration of these elements via the Velikaya River (Eilart 1963). Houseleek (*Sempervivum soboliferum*), as one of the Pontosarmatic indicator of flora (Photo 10), has its western habitat boundary in southeastern Estonia and on the east coast of L. Peipsi (Gorodishche Kame Field, Photo 11).

The area belongs to the South-Estonian region of light pseudopodzolic soils or Planozols (Reintam 1997). In a southerly direction, the thickness of the glaciolacustrine cover on till increases, and the Planozols are replaced first by Podzoluvizols and further — at Palumaa — by Podzols.

The area between Tartu and Põlva offers much more suitable soil conditions for agriculture. The proportion of arable land is gradually diminishing southwards. The forested areas dominate to the south from the Võhandu River. Depending on the moisture conditions, the pine forests contain lichen, moss, *Calluna* or *Vaccinium*.

Of the **Plyussa Plateau,** only the western part extending from the coast of L. Peipsi to the Plyussa River belongs to the lake basin described here (Fig. 6). At an height of 40–55 m a.s.l., the slightly drumlinized and undulating till plains are dominating. The coastal belt of sandy-loamy Planosols and Luvisols, *ca* 15 km in width, is actively used for agricultural purposes. The peripheral zone is occupied by secondary forests of birch and pine with *Calluna* and *Vaccinium* or association of birch, aspen and alder. The lichen — pine association can be found only to the south from the Kuna River in the Torokhovo Esker and kame complex.

Against the background of the undulating plain, the limited areas of kame hummocks (Zigoska – Beshkino, Torokhovo), hummocky moraine (nearby the mouth of the Kuna Valley), deep glaciofluvial valleys (Kuna, Cherma) and also a marginal esker at Torokhovo dissect the landscape.

The landscape regions of the lake basin offer different preconditions for land use. The resistance to natural changes

Photo 7. Steep downwind slope of the dunes at Podlip'e, Remda Peninsula. Photo by R. Karukäpp.

Photo 8. Lake Dolgoe in the Plotkino Kame Field, Remda Peninsula. Photo by R. Karukäpp.

Photo 9. Cliff in the Devonian sandstones at Kallaste. Photo by A. Miidel.

Photo 10. Houseleek (*Sempervivum soboliferum*) on the southern slope of a kame in the Gorodishche Kame Field. Photo by R.Karukäpp

Photo 11. Gorodishche Kame Field. Photo by R. Karukäpp.

and response to human impact are also different. The contemporary landscapes here suffer less from human activities than in surrounding areas.

In the lake basin, the processing of agricultural production and fish is the main branch of industry. Pskov, with the population of over 340,000 in 1994, is the biggest town in the region. It was mentioned as early as 862 and 903 AD in the Russian Chronicle *The Tale of Bygone Years*. Pskov is famous for its church and monastery buildings dating from the 12th –17th centuries. It is the principal cultural, economic and educational centre of the Pskov Region.

There are acting oil-shale mines in the northern part of the lake basin. In underground mining, mainly two kinds of technology have been used: room-and-pillar mining technology, where the basic roof of the mined-out area is supported by oil-shale pillars, and narrow-web mining by cutter – loader with a total collapse of the roof. During Soviet time, the calculated dimensions of the pillars for the room-and-pillar technology had to ensure the safety during the mining period only. As a result, some 2130 ha or 2.9 % of the mined-out area had collapsed by 1993 (Kaljuvee 1993). Deformation of the surface results in the accumulation of water. Paludification of the area has been rather fast and the natural balance of the forests and mires has been destroyed (Photo 12). No environmental safe technology has been elaborated so far.

Photo 12. Collapsed surface of the Kalina Mire above the exhausted Viru oil-shale mines. Photo by R.Karukäpp.

2. GEOLOGY OF THE LAKE BASIN

2. 1. BEDROCK GEOLOGY

Precambrian basement. The thickness of the crust in eastern Estonia increases slowly from about 45 to 50 km (Ankudinov *et al.* 1994) towards the south. The crust consists mostly of basement rocks as the thickness of the sedimentary cover is small ranging from 0.3 km in the north to 0.6 km in the south.

Palaeoproterozoic metamorphic and igneous rocks of the Peipsi area belong to the Svecofennian Domain. The supracrustal rocks were deposited less than 2 Ga ago, mostly around 1.9 Ga (Koistinen 1996), with the latest postorogenic input 1.62–1.65 Ga ago.

Metasediments (schists and gneisses) are common in the north. Gneisses containing biotite, cordierite, garnet and sillimanite are characteristic of the Alutaguse Zone (Fig. 7), where they intercalate with biotite gneisses and form a complex of the same name (Puura *et al.* 1976). The gneisses are migmatized by plagioclase-microcline granite or pegmatite.

Basic to intermediate **metavolcanic rocks** are common in the south. They are divided into two main groups by their mineral composition: hornblende-pyroxene and biotite-hypersthene gneisses (Koistinen 1996). The gneisses have been subject to the charnockitic and plagioclase–orthoclase-granitic migmatization. Biotite-hypersthene gneisses mostly occur in the vicinity of Tartu and Otepää, 35 and 60 km west of L. Peipsi, respectively. Acidic metavolcanic rocks are less abundant.

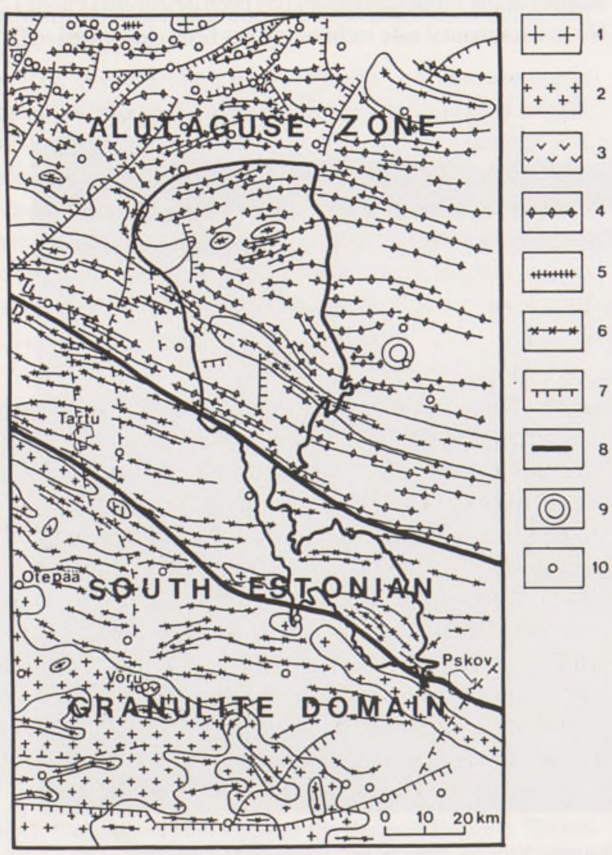

Fig. 7. Palaeoproterozoic basement of the Lake Peipsi area. Adapted from Koistinen (1996): 1 - rapakivi granite; 2 - granite; 3 - mafic plutonic rocks; 4–6 - trace of dominant strike trends of mica schist and mica gneiss (4); graphitic schist (5); mafic and intermediate metavolcanic rocks (6); 7 - Phanerozoic fault; 8 - Proterozoic shear zone: U – upthrown side, D – downthrown side; 9 - crater; 10 - borehole finished in the basement.

Orogenic plutonic rocks are common everywhere. Synorogenic granitoids are mainly migmatite granites (Koistinen 1996). In the north, in the Alutaguse Zone, they are represented by plagioclase–microcline granites, while in the south charnockites and plagioclase–orthoclase granites occur. A small gabbronorite pluton, probably synorogenic, has been drilled at Võru, 45 km southwest of L. Pihkva.

Rapakivi granite, known as the Ereda Pluton, occurs 35 km north of L. Peipsi. The pluton is composed of homogenous pink-grey coarse-grained porphyritic granites (Soesoo & Niin 1992). A specific feature of the Ereda Pluton is weak late-postmagmatic deformation (Kuuspalu 1975, Kirs 1986).

Palaeoproterozoic metamorphism was either low P/high T type, or andalusite-sillimanite type (Koistinen 1996). In the south, in the granulite facies gneisses, widespread garnet and cordierite, formed by the breakdown of biotite and sillimanite, indicate prograde metamorphism. The PT-conditions have been calculated using several geothermometers and geobarometers (Hölttä & Klein 1991). These give temperature estimates for the prograde stage of metamorphism of 700–800°C and pressure estimates of 5–6 kbar or greater. In the more heterogenous gneisses of the north the dominant metamorphic grade is high temperature amphibolite facies. Geothermobarometry, mostly of the biotite + garnet "sillimanite" assemblage and of cordierite, estimates prograde metamorphism at 600–700°C and 3–5 kbar.

Sedimentary bedrock. The Late Proterozoic, Early and Middle Palaeozoic seas exhibiting multitude of depositional and erosional phases covered the area now occupied by L. Peipsi. They formed the sedimentary bedrock including the Vendian, Cambrian, Ordovician, Silurian and Devonian (Table 2). Three lithological macrounits of the Upper Vendian–Middle Palaeozoic sedimentary bedrock can be separated: Upper Vendian–Lower Ordovician siliciclastic supergroup, Middle Ordovician–Silurian carbonate supergoup and Devonian mainly siliciclastic supergroup.

The Late Vendian sea inundated the area from the east, where the large Moscow Basin had been formed. The Vendian involves sandstones, siltstones and claystones of the Kotlin Stage. The thickness of the Vendian decreases in a southwesterly direction from 110 to some 50 m. After the uplift of the area at the end of the Vendian, accompanied with the denudation, the Cambrian sea again inundated it from the east. It is possible that at the eastern coast of L. Peipsi there are some rocks of the Redkino Stage but the whole area is covered over with the claystone of the Lontova Stage, the thickness of which decreases towards the southwest from 70 to some 40 m. The end of the Lontova Age is marked with a distinct break in sedimentation that terminated the westward

Table 2. The sedimentary bedrock of Lake Peipsi and surrounding area (Raukas & Teedumäe 1997, Webby 1997)

System	Series	Regional Stage
DEVONIAN	UPPER	Pļaviņas
	MIDDLE	Amata
		Gauja
		Burtnieki
		Aruküla
		Narva
		Pärnu
	LOWER	Rēzekne
		Tilžė
SILURIAN	LLANDOVERY	Raikküla
		Juuru
ORDOVICIAN	UPPER	Porkuni
		Pirgu
		Vormsi
		Nabala
		Rakvere
		Oandu
		Keila
		Haljala
		Kukruse
	MIDDLE	Uhaku
		Lasnamägi
		Aseri
		Kunda
		Volkhov
	LOWER	Billingen
		Hunneberg
		Pakerort
CAMBRIAN	UPPER	
	MIDDLE	
	LOWER	Vērgale
		Dominopol'
		Lontova
VENDIAN	UPPER	Kotlin

transgression (Rozanov & Lydka 1987). The Moscow Basin had risen above sea level and remained in such a state up to the Middle Cambrian.

The Dominopol' sea inundated the Peipsi area from the west, where a new subsiding area – the Baltic Basin – formed. It was opened to the west and had a direct connection with the ocean. The alternating claystones and sandstones of the Dominopol' Stage (thickness 0–6 m) and sandsones of the Ljuboml' and Vērgale stages (0–21 m) are distributed over the northern part of the area. Because of incomplete sections, the highly complicated evolution of the area in the Middle (Raukas & Teedumäe 1997, p.188) and Late Cambrian is difficult to follow. The thickness of the Middle (0–35 m) and Upper (0–11 m) Cambrian increases towards the south. The total thickness of the Cambrian ranges from 80 m in the southwest to 200 m in the northeast. The Middle Cambrian, and the lower and upper parts of the Upper Cambrian are represented by sandstones, whereas the middle part of the Upper Cambrian is predominantly composed of argillaceous rocks. The Lower Ordovician is siliciclastic as well. Up-to-4-m-thick sandstone of the Pakerort Stage occurs at the northern coast of L. Peipsi. The whole area is covered over with a thin (less than 1 m) blanket of siltstone and sandstone of the Hunneberg and Billingen stages.

During the Middle and Late Ordovician the area was part of the northern flank of a shallow cratonic sea in which carbonite sediments accumulated. In general, the Ordovician carbonites of the area are more argillaceous than those in western Estonian (Põlma & Kleesment 1985). The Middle Ordovician (30–50 m thick) and the lower part (40–60 m) of the Upper Ordovician including most of the Oandu Stage consist of argillaceous limestones and marls. The thickness of the Upper Ordovician ranges from 50 to 150 m. The end of the Middle Ordovician, and beginning of the Late Ordovician (Uhaku and Kukruse ages) was a period when kerogenous, kukersite-bearing deposits accumulated at the northern coast of L. Peipsi. A 1–2-m-thick commercial oil-shale seam belongs to the base of the Kukruse Stage. The nearest opencast oil-shale mine is situated 25 km north of L. Peipsi. During the Haljala and Keila ages, numerous volcanic ash (metabentonite) intercalations formed. The Rakvere and Nabala stages mainly consist of more or less dolomitized aphanitic limestones which southwards become more argillaceous. Rocks of the Nabala Stage form an outcrop area at the northern coast of L. Peipsi (Fig.8). Outcrops of the stage may be found in the channel of the Rannapungerja River and on the Narva riverside at Permisküla. The Vormsi and Pirgu stages are represented by argillaceous limestones

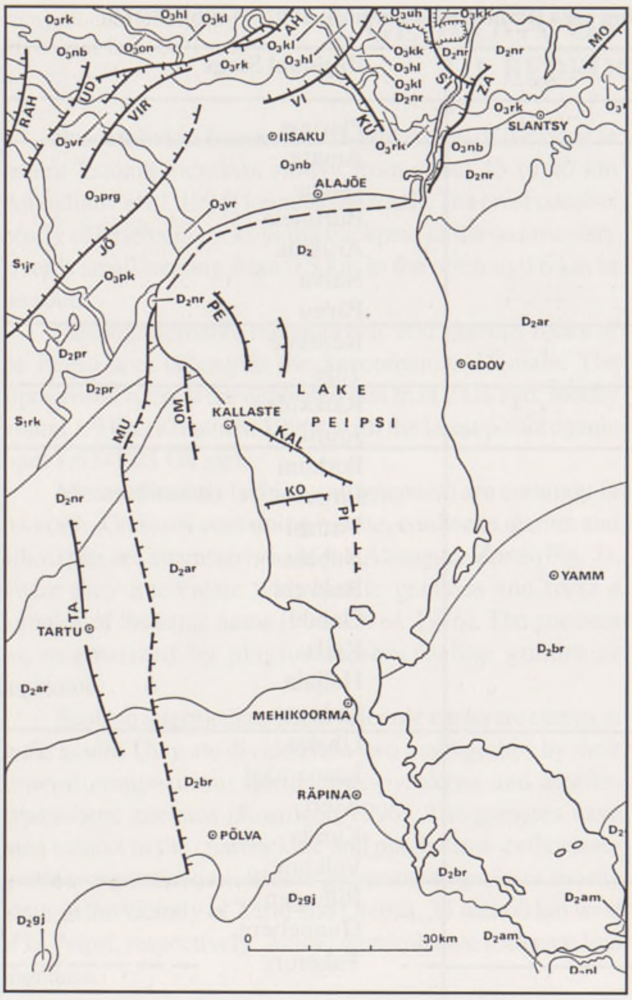

Fig. 8. Bedrock geology of L. Peipsi and surrounding area (adapted by R. Vaher from Suuroja 1997). Outcrop areas of the stages, Upper Ordovician: O_3uh - Uhaku, O_3kk - Kukruse, O_3hl - Haljala, O_3kl - Keila, O_3on - Oandu, O_3rk - Rakvere, O_3nb - Nabala, O_3vr - Vormsi, O_3prg - Pirgu, O_3pk - Porkuni; Lower Silurian: S_1jr - Juuru, S_1rk - Raikküla; Middle Devonian: D_2pr - Pärnu, D_2nr - Narva, D_2ar - Aruküla, D_2br - Burtnieki, D_2gj - Gauja, D_2am - Amata; Upper Devonian: D_3pl - Pļaviņas. Zones of disturbances: AH - Ahtme, JÕ - Jõgeva, KAL - Kallaste, KO - Kolkja, KU - Kuremäe, MO - Monastyrek, MU - Mustvee, OM - Omedu, PE - Peipsi, PI - Piirissaare, RAH - Rahkla, SI - Sirgala, TA - Tartu, UD - Udriku, VI - Viivikonna, VIR - Virunurme, ZA - Zagriv'e

distributing in the western part of L. Peipsi. A general regressive trend of basin development is recognizable in the Pirgu Age. Sandy limestones of the Porkuni Stage occur more than 5 km west of the lake. The total thickness of the Ordovician ranges from 110 to 180 m. Silurian (Juuru Stage) limestones and marlstones are also spread more than 5 km west of the lake.

Most of L. Peipsi belongs to the Devonian outcrop area (Fig. 8). At the end of the Silurian and beginning of the Devonian the Peipsi area was above sea level, being subject to denudation. The Devonian transgression started from the southwest. Silt- and sandstones (2.1 m) of the Lower Devonian Tilžė Stage have been palaeontologically determined only in the Värska drill core (Sorokin 1981). Sandstones (lower part of the section) and dolomitic marls (upper part) of the Rēzekne Stage (0–30 m) spread further to the north and sandstones of the Middle Devonian Pärnu Stage (0–35 m) even more. The transgression reached its maximum in the Narva Age. The lower part of the Narva Stage consists mostly of dolomitic marl with interlayers of dolomite and dolomitic claystone. Siltstone and sandstone intercalating with dolomitic marl and claystone form the upper part of the stage. In the south, the Narva Stage is up to 90 m thick. The Aruküla (0–80 m), Burtnieki (0–70 m) and Gauja (0–35 m) stages consist mostly of sandstones with interbeds of siltstone, claystone and infrequent dolomitic marlstone. The Amata Stage (0–20 m) is represented by alternating sand- and siltstones with interbeds of claystones. The lower part of the Upper Devonian Pļaviņas Stage consists of rhythmically alternating dolomite and dolomitic marl. The upper part of the stage is represented by limestones and dolomites. The thickness of the Pļaviņas Stage reaches 30 m. The outcrop areas of the Middle and Upper Devonian stages form northeast-trending, regularly arranged belts, the age of which decreases in a southerly direction (Fig. 8). The outcrop area of the Pļaviņas Stage occurs at the southern coast of L. Pihkva. In the south, the thickness of the Devonian exceeds 300 m. The total thickness of the sedimentary bedrock ranges from 300 m in the north to 600 m in the south.

2.2. TECTONICS AND FORMATION

The sedimentary bedrock of Estonia is divided into three **tectonic stages**: Baikalian (Vendian and Lower Cambrian, Lontova Stage incl.), Caledonian (Postlontovan Cambrian, Ordovician, Silurian and Lower Devonian, Tilžė Stage incl.) and Hercynian (Posttilžėan Devonian), named mostly by geologists of the Baltic countries (Suveizdis 1979). The tectonic stages are separated with regional unconformities. All these stages are represented in the Peipsi area.

Within the limits of the lake, the basement plunges in a SSE direction from -280 m at Rannapungerja to -600 m near Pskov (Fig. 9) having a very gentle (8') **regional dip**. The regional dip of the sedimentary bedrock strata ranges from 8 to 10'. Local variations in the dip are clearly observable in several sections (Figs. 9, 10) of the area. In the southwest, the mainly monoclinal regional dip terminates at the Haanja-Lokno Uplift, which is the eastern part of the 200-km-long Valmiera – Lokno Uplift. The basement rises to the level of -425 m on the crest of the Lokno Anticline, and descends rapidly in a southerly direction to the level of -800 m (Fig. 9).

The largest, mostly monoclinal structure of the area is complicated by many minor structural features of various kind described in detail in northeastern Estonia (Vaher *et al* 1962, Puura 1986, 1987), on which a wealth of data has become available owing to extensive exploration for oil shale and phosphorite (over 10,000 boreholes per 2,900 km²).

Narrow linear **zones of disturbances**, 1 to 4 m in width,

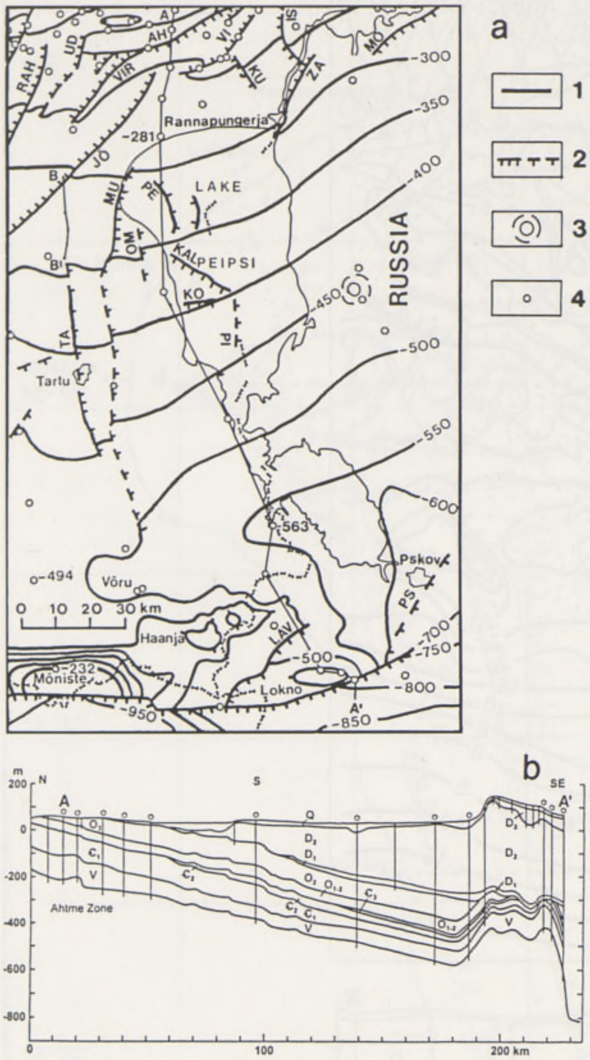

Fig. 9. Structure-contour map (a) of L. Peipsi and surrounding area (adapted by R. Vaher from Sildvee & Vaher 1995) and section (b, by R. Vaher): 1 – contour on top of the Palaeoproterozoic basement; 2 – flexure above a basement fault: defined (solid), assumed (dashed); 3 – crater; 4 – borehole, finished in the basement. A-A', B-B' – location of the geologic sections (section B-B' see Fig. 15). Zones of disturbances: AH – Ahtme, JÕ – Jõgeva, KAL – Kallaste, KO – Kolkja, KU – Kuremäe, LAV – Lavry, MO – Monstyrek, MU – Mustvee, OM – Omedu, PE – Peipsi, PI – Piirissaare, PS – Pskov, RAH – Rahkla, SI – Sirgala, TA – Tartu, UD – Udriku, VI – Viivikonna, VIR – Virunurme, ZA – Zagriv'e.

intersect both the sedimentary cover and the crystalline basement. These zones divide the area into numerous blocks of various sizes. As a rule, the zones of disturbances occur as a flexure above a basement fault combined with anticline in the upthrown side, and with syncline in the downthrown side. The total vertical amplitude A_t (the distance between trough and crest surfaces) in the sedimentary bedrock can be considered as consisting of components, known as "anticlinal" (A_a), "monoclinal" (A_m), and "synclinal" (A_s).There are joint (fracture) zone(s) on the more steeply dipping limb and, in places, fault(s) in the joint zone.The folds in the zones of disturbances are in general very gentle and usually 5 to 10, but never more than 50 m high.

The Viivikonna Zone is described in the Viivikonna Openpit about 30 km north of L. Peipsi. Here the dominant feature of the structure is the anticline (A_a = 5.4 m, A_m = 1.8 m, A_s = 0.8 m; A_t = 8 m). The beds in its southeastern limb have a very gentle average dip of 20' SE (max. 55'), in the northwestern limb the dip is 7° NW (Puura & Vaher 1997). This anticlinal belt is ascribed to fault movements in the basement, because its shape is very asymmetrical. In the 125-m-wide fracture zone on the more steeply dipping limb, the intensity of jointing increases towards the centre, and some oil shale beds are substituted by karst clay. In the 40-to-50-m-wide part of the structure, called the shatter zone, 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 Ahtme Zone is described in the Ahtme Mine (Kattai & Vingissaar 1980), 15 km west of the Viivikonna Openpit section. In the Ahtme section, the dominant feature is a flexure above a basement fault (A_m = 13 m, A_s = 5 m, A_a = 0 m; A_t = 18 m). This limb has a dip of 1 to 18° SE. The fracture zone is 200 m wide, and its karsted central part (shatter zone) is 80 m wide. In the latter, there is a fault with the throw of 3 m. Along the Ahtme Zone of disturbances one can recognize contrasting movements related to the succeeding periods. Thus, the flexure in the Ordovician strata above a basement fault is due to the rise of the northwestern block, while a minor fault in the Ahtme Zone is indicative of somewhat later rise of the southeastern block.

In the eastern part of the Baltic Oil-Shale Basin, where the Devonian rocks are distributed (Fig. 10), the Zagriv'e Zone extends as far as the northeastern part of L. Peipsi. According to the drilling data, the dominant feature of the zone is a flexure above a basement fault (A_m= 8–12 m, A_a= 1–4 m, A_s= 2–3 m; A_t up to 16 m). Both the Ordovician and Devonian are disturbed and, thus, the Zagriv'e Zone is post-Devonian (Tuuling 1988).

Fig. 10 shows also an escarpment, known as the "Wesenberg Escarpment" or "Wesenberg Klint". The steep, cliff-like escarpment is 30–45 m high. It is clearly observable on the base of the Devonian, but missing on the base of the Upper Ordovician (Fig. 10b). Consequently, it is a pre-Devonian erosional scarp overlain by Devonian deposits. Thus, in this area linear structures of different origin and age have been found.

A sedimentary bedrock structure within the limits of L. Peipsi was investigated by means of high resolution shallow marine seismic reflection profiling. The net of seismic profiles (Fig. 11) was shot during summer 1994 (Noormets *et al.* 1998). The contours on top of the Ordovician (Fig. 12) show a regional dip of about 8' and numerous local structures (Miidel *et al.* 1999). Morphologically, the zones of disturbances here are similar to the Ahtme and Zagriv'e zones, as the dominant feature is the flexure above a basement fault. For example, according to the interpretation by R. Noormets, the Peipsi Zone in section 4–4' (Fig. 13a) has a total amplitude A_t = 24 m (A_m= 17m, A_s = 4 m, A_a = 3 m). Eight seismic profiles intersecting the zone, suggest 15 km for its length. Fig. 14 shows that in section 1–1' (Lohusuu) the Kallaste Zone has a total amplitude A_t =31 (A_m = 27m, A_s = 3m, A_a = 1m).

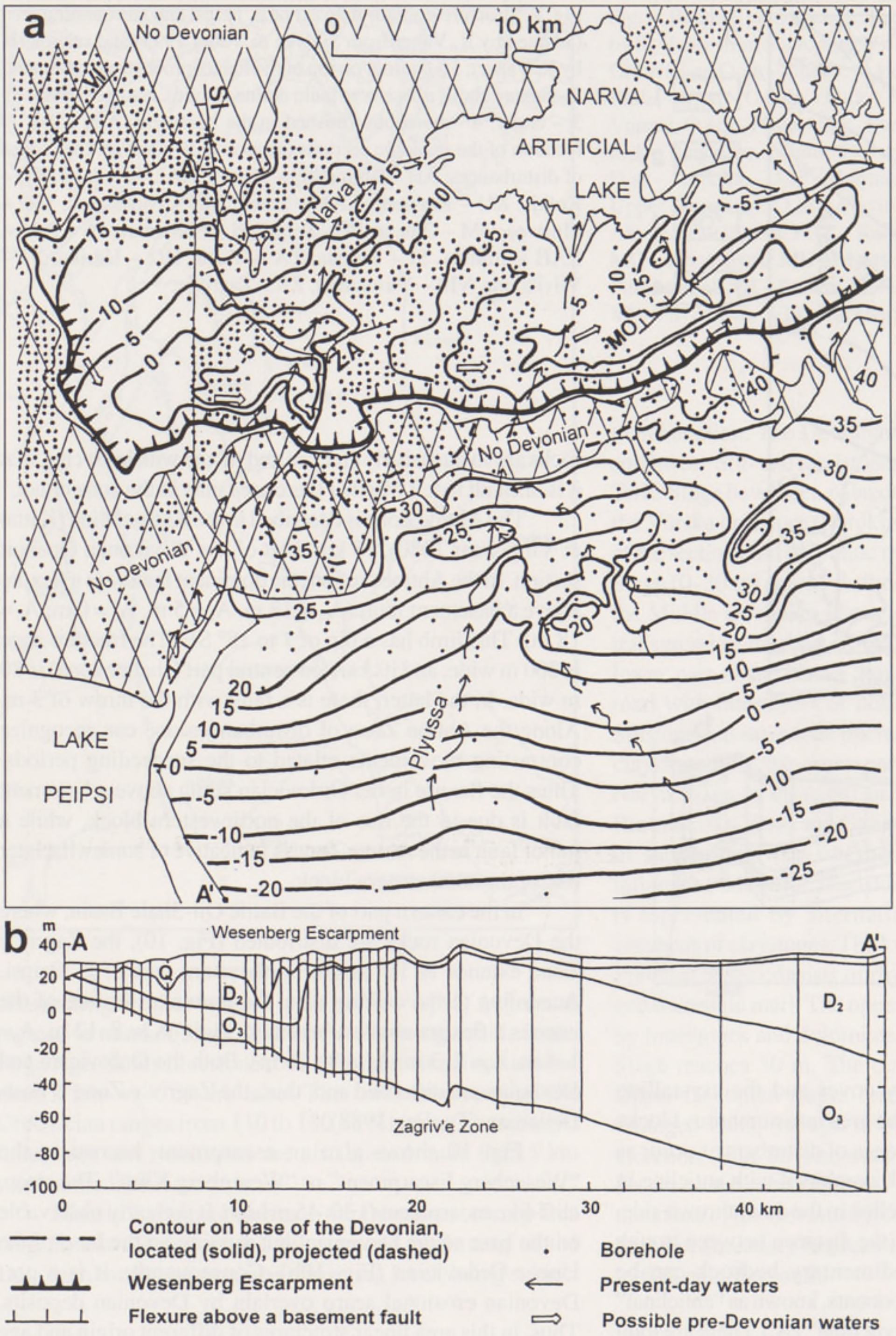

Fig. 10. Bedrock structure in the eastern part of the Baltic Oil-Shale Basin, map (a – adopted from Tuuling 1988) and section (b – by R. Vaher). Zones of disturbances: MO – Monastyrek, SI – Sirgala, VI – Viivikonna, ZA - Zagriv'e.

In section 2-2' (Rannapungerja), the total amplitude of the zone A_t = 48 m (A_m = 39 m, A_s = 7 m, A_a = 2 m). For the Kolkja Zone, in section 1–1' A_t = 25m (A_m = 20 m, A_s = 3m, A_a = 2 m) and in section 2–2' A_t = 18 m (A_m = 15 m, A_s = 2 m, A_a = 1m).

Small **basins and domes**, 0.1 to 1 km in diameter, with an amplitude of 0.5–1.5 m have been found in oil-shale mines (Baukov & Kotlukov 1973). The largest (0.5–0.6 km), Narva Basin is known 29 km north of L. Peipsi in the Narva Openpit because of exploration for oil shale in 1961. The basin is oval in outline. In the centre of the structure, the Ordovician strata lie 50 m below the level of surroundings. Figure 12 shows four basins (closures), 2 to 6 km in diameter, within the limits of L. Peipsi. In the centre of basins, the Ordovician strata lie some 10 to 20 m below the level of surroundings. Some 5 to 15 m high local anticlines (area 2 km × 4–6 km × 12 km) occur as noses in Fig. 12.

An interesting structure is the Mishina Gora **Crater** 18 km east of L. Peipsi (Fig. 8). In the crater, the Quaternary sediments are in places underlain by Ordovician, Cambrian and basement rocks surrounded by an outcrop area of the Middle Devonian rocks. The rim-to-rim diameter is 4–5 km. It is suggested that the structure is a cryptovolcanic (Shmaenok & Tikhomirov 1974) or a meteoritic impact crater (Puura *et al.* 1994).

It would be reasonable to deal with the formation and age of the Peipsi Basin in a broader sense, taking under consideration also the surrounding area. The Narva–Luga

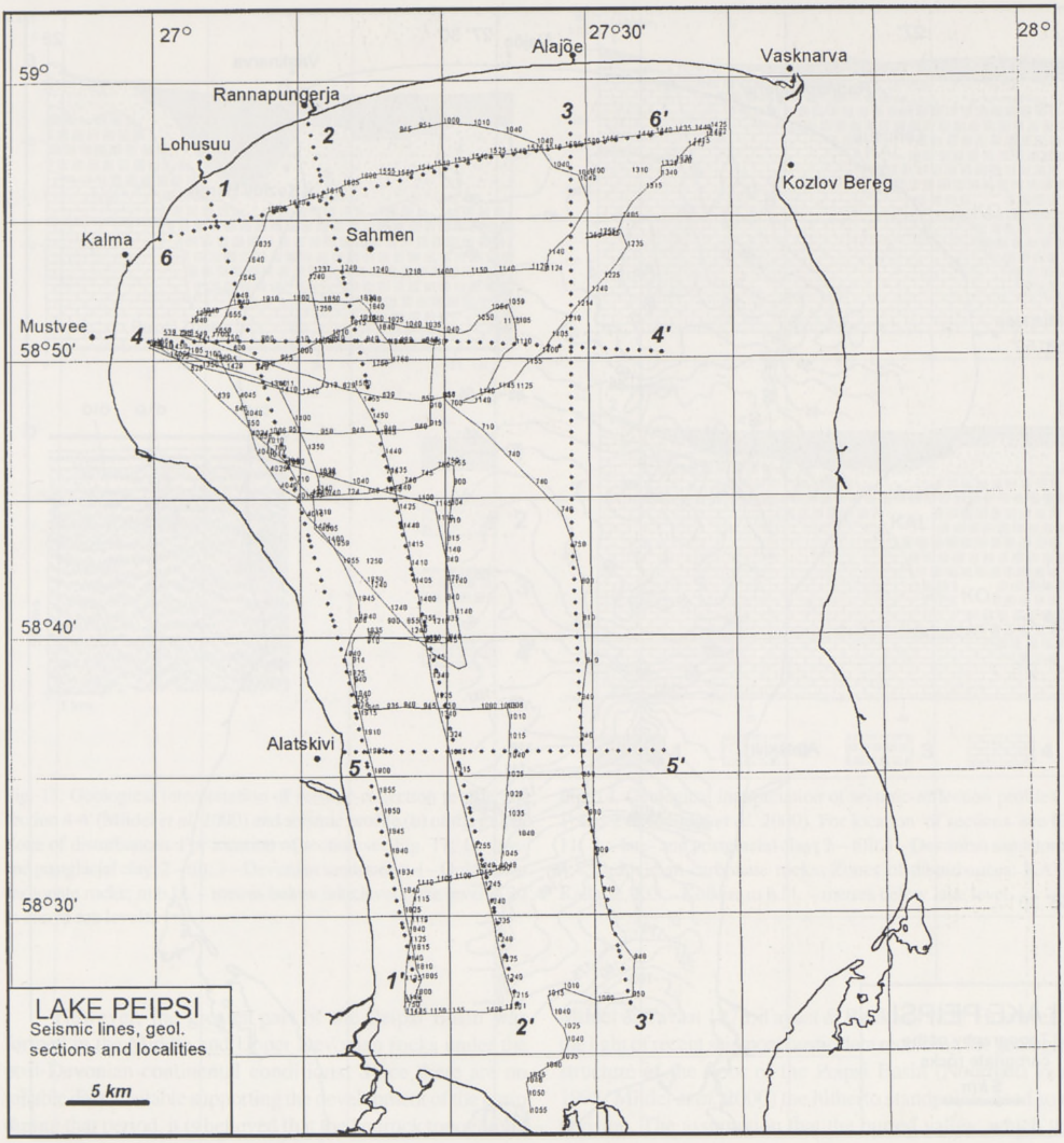

Fig. 11. L. Peipsi with the location of seismic profiles (solid) and geologic sections (dotted) (Noormets *et al.* 1998).

Lowland, bordering the Peipsi Basin on the north, is believed to be very old. In the northern part of the Leningrad District of Russia, and in North-East Estonia, the Middle Devonian rocks rest unconformably on the Ordovician rocks (Schmidt 1883, Markov 1931, Tammekann 1949, a.o.), which suggests that the Narva–Luga Lowland formed before the Middle Devonian. Kajak (1964) maintains that at least the northern part of the lake basin came into being before the Narva Age.

The relief buried under the Middle Devonian sediments has been studied in a particular detail by Sammet (1961) and Tuuling (1988a). According to their data, the pre-Devonian topography has a general SSE inclination . Its major element – a 10–12-km-wide negative topographical feature running from west to east – is presumably a river valley, which was cut in the Ordovician rocks. The channel is infilled by Devonian sediments, 30–45 m in thickness. The southern slope of the valley is known as the Wesenberg Escarpment, because it consists of hard limestone of the Rakvere Stage. As is known, the terrace from the limestones of the Rakvere Stage runs through the whole of northern Estonia (Schmidt 1883, Tammekann 1949).

According to Heinsalu (1971) and Gazizov (1971), the development of the karst topography in the oil-shale basin in northeastern Estonia started before the Middle Devonian. At the beginning of the Devonian or even before the Devonian, the Jõhvi (Ahtme) bedrock elevation (Puura *et al.* 1987), the Pandivere Upland and the Narva–Luga Lowland (Puura *et al.* 1999) formed in their general lines. Recent data on the distribution of Middle Devonian sediments in the northern part of the Peipsi Basin and on the southeastern slope of the Pandivere Upland support pre-Devonian age of the above-mentioned topographic features of the bedrock (Kiipli 1989).

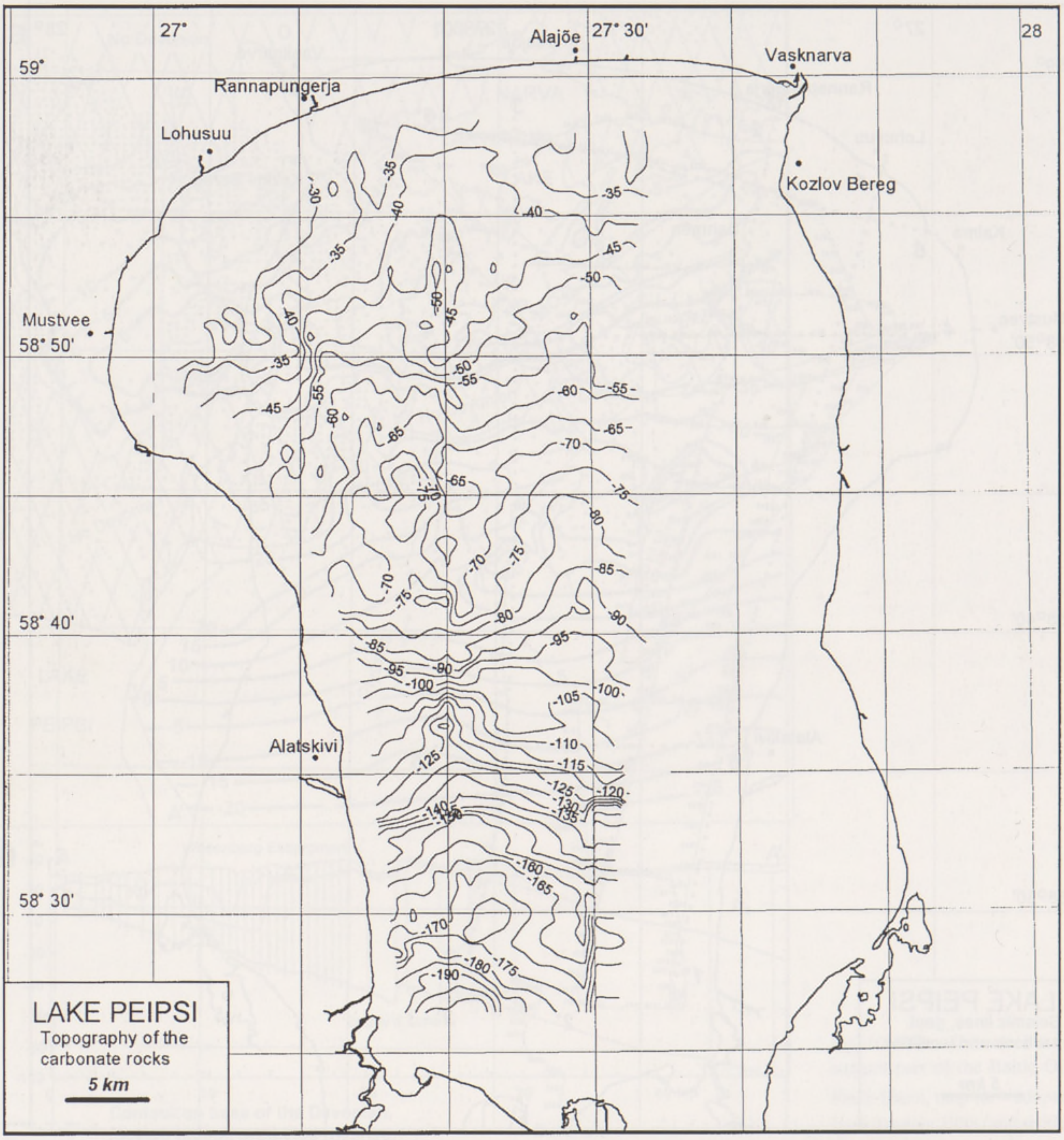

Fig. 12. L. Peipsi, contour on top of the Ordovician (Miidel *et al.* 2000). Contours show the depth below lake level (30 m a.s.l.).

A meridionally oriented profile in the vicinity of the Narva River shows clearly how the upper surface of the Upper Ordovician rocks rises slowly from under the Middle Devonian Narva Stage, exposes under the Quaternary sediments, and disappears then again smoothly under the Devonian rocks. It is noteworthy that the upper surface of the Ordovician rocks has not essentially changed under the Quaternary deposits. This suggests that the bedrock surface was partly formed in pre-Devonian times as a result of continental denudation, and was somewhat reworked by glacial erosion.

There are some interesting data available concerning the spread of Devonian rocks on the southwestern slope of the Pandivere Upland, or rather on the boundary of the Viru–Harju Plateau and the Peipsi Basin which is transitional in places. Borehole records show that between the villages of Vaiatu and Süvalepa the rocks of the Pärnu Stage, seldom of the Narva Stage, overlie the carbonate rocks of the Juuru and Raikküla stages (Fig. 15). Middle Devonian rocks form "patches" in a large valley, over 16 km in width and 40–50 m in depth, which was eroded into the upper surface of the Silurian rocks (Kiipli 1989). In places, the scarp is even comparable with the North-Estonian limestone escarpment, *e.g.* in the vicinity of Rääbise Village where the height of the bedrock falls 49 m within a 1.2-km-long stretch (Kiipli 1989).

The above-presented data prove convincingly that in the northern slope of the lake basin the pre-Devonian topographic features of the bedrock were re-exposed and are erosional-denudational in origin. In the southern part of the lake basin it is buried under Middle and Upper Devonian sediments.

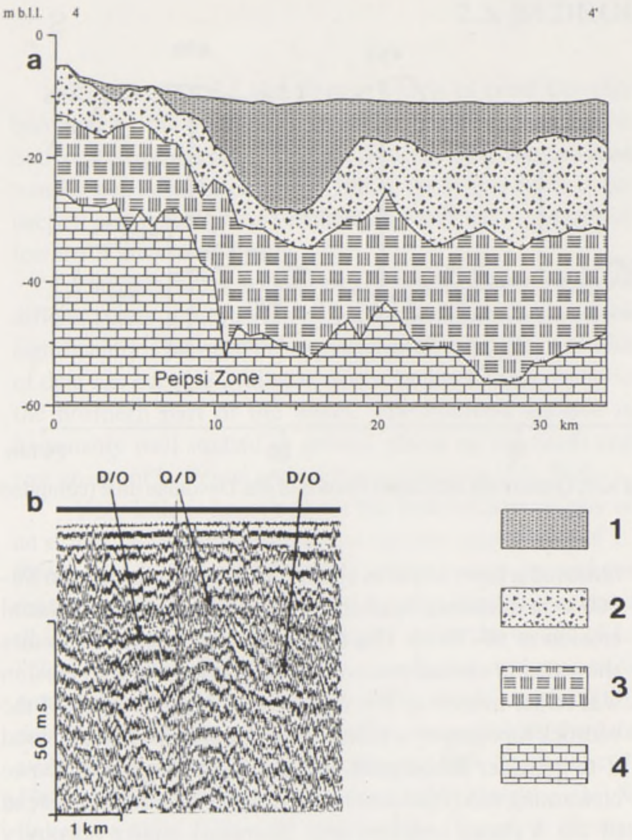

Fig. 13. Geological interpretation of seismic-reflection profile (a), section 4-4' (Miidel *et al.* 2000) and seismic profile (b) of the Peipsi Zone of disturbances. For location of section see Fig. 11: 1 – late- and postglacial clay; 2 – till; 3 – Devonian sandstones; 4 – Ordovician carbonate rocks; m b.l.l. – metres below lake level (lake level is 30 m above sea level).

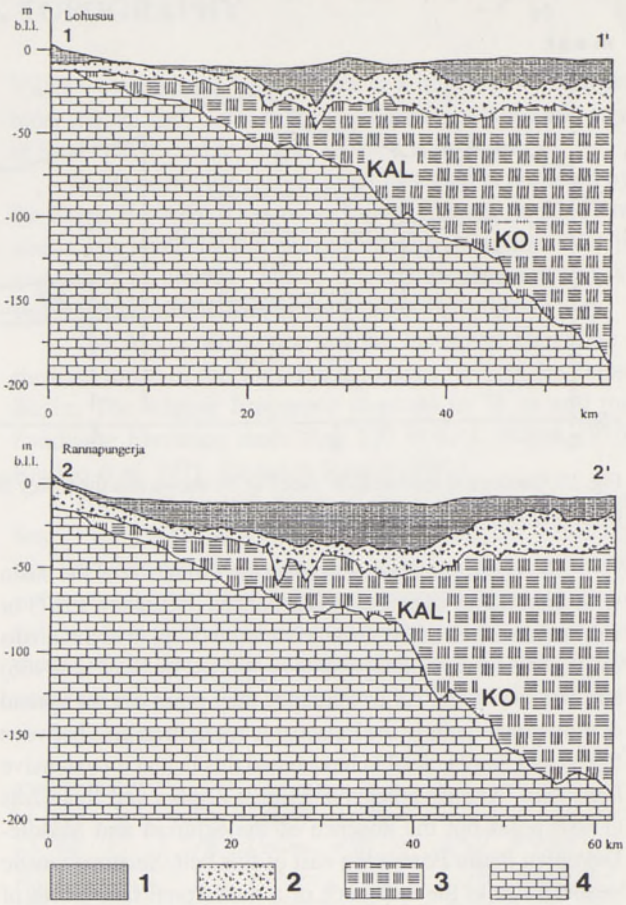

Fig. 14. Geological interpretation of seismic-reflection profiles 1–1' and 2–2' (Miidel *et al.* 2000). For location of sections see Fig. 11: 1 – late- and postglacial clay; 2 – till; 3 – Devonian sandstones; 4 – Ordovician carbonate rocks. Zones of disturbances: KAL – Kallaste, KO – Kolkja; m b.l.l. – metres below lake level.

However, the greater part of the Peipsi Basin was formed in the Middle and Upper Devonian rocks under the post-Devonian continental conditions. Since there are no reliable data available supporting the development of the basin during that period, it is believed that the bedrock topography of the lake basin was formed in pre-Quaternary times and was afterwards reworked by the Pleistocene glaciers (Tammekann 1949, Tavast & Raukas 1982). The valleys featuring the bedrock topography and infilled by Quaternary sediments are traditionally recognized as pre-Quaternary in age (Sammet 1961, Kajak 1963, 1970, Kvasov 1975, 1979, Miidel & Tavast 1978, Tavast & Raukas 1982), although sediments older than the Quaternary have never been found in these valleys. It is supposed that the pre-Quaternary watershed, established in Russia, extended to the northern part of the lake basin, where the buried valleys are lacking (Sammet 1961, Miidel 1966). The rivers north of this watershed flowed towards the north, and the rivers south of it ran their waters to the south-west, towards the present-day Gulf of Riga. From the east, the rivers flowed either into the lake basin (Isachenkov 1969) or straight across the lake basin to Estonia (Kvasov 1975). It has been assumed that the area, which is now the Peipsi Basin, held several valleys along which rivers flowed towards the south (Tammekann 1949, Miidel & Tavast 1978, Tavast & Raukas 1982). However, in the light of recent seismoacoustic data recording the geological structure of the floor of the Peipsi Basin (Noormets *et al.* 1998, Miidel *et al.* 2000) the hitherto standpoints need some revision. The assumption that the buried valley, which was discovered in the vicinity of Mustvee, extends in a northeasterly direction up to the Sahmen Shallow (Tavast & Raukas 1982) has found no confirmation. These data do not support the existence of a meridionally-oriented valley either, unless it is situated to the east from the studied area, which is unlikely.

Opinions differ as to the formation of the Peipsi Basin. Some researchers maintain that the basin and the adjacent lowlands are denudational land forms (Markov 1931, Orviku 1960). Others (Sammet 1961, Nikolaev 1967, Shul'ts 1969, Kaplan & Suveizdis 1970, Zander & Salomon 1971, Baeva 1978, a.o.) state that it was tectonics that played the leading role. Sammet (1961) pointed out several facts which, in his opinion, support the tectonic origin of the basin. These include, for instance, the falling of old valleys into the basin, the occurrence of scarps on the west coast of the lake and folded structures in L. Pihkva. According to Kajak (1964), the faults trending NE and NW were an important prerequisite to the tectonic genesis of the Narva-Luga and Ojamaa lowlands.

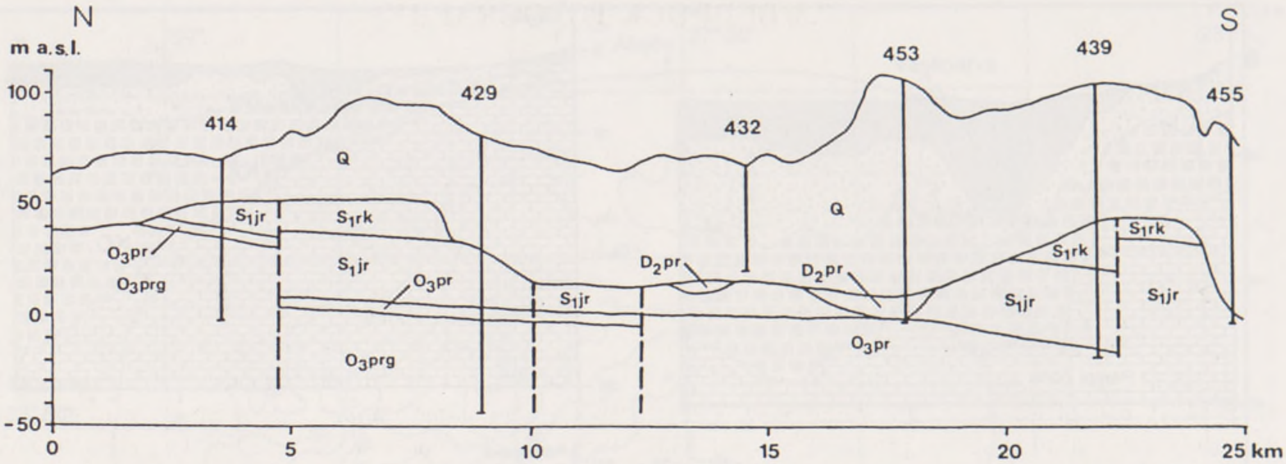

Fig. 15. Geological section B-B' (see Fig. 9) shows that the valley filled with Quaternary sediments formed in pre-Devonian time (compiled by A. Põldvere, Geological Survey of Estonia).

Several researchers associate the formation of the basin with meridional basement disturbances (Nikolaev 1967) or with a mobile belt of the same orientation (Kaplan & Suveizdis 1970). The former standpoint is supported by Shul'ts (1966) and Baeva (1978). As an evidence they point out the spread of orthogonal faults in the Palaeozoic rocks in an area between Pskov and Narva. Shul'ts has treated this belt as an extensive lineament. Among other indications, Kaplan and Suveizdis (1970) point out the absence of the Silurian and Middle-Devonian Pärnu Formation east of this belt. Seismoacoustic studies refer to the existence of a meridional fault north of Piirissaar Island and the Paldiski–Pskov Deep Fault running through the lake basin from northwest to southeast (Miidel *et al.* 2000) which is reflected in the thickness of the Devonian complex and a rapid deepening of the Ordovician/Devonian contact.

However, there are also data contradicting to the existence of the regional meridional fault zone or lineament in the Peipsi Basin. The studies by Zander and Salomon (1971) suggest the existence of both orthogonal and diagonal joints in the surroundings of Pskov. The valleys cutting the basement in the Izborsk area change the direction at a right angle which suggests the occurrence of the joints of NW and NE strike (Bekker 1924). Joints of northwestern orientation are widespread in the Narva canyon and waterfall. Judging by the facies models by Männil (1966), this belt is reflected neither in the thicknesses of the Ordovician sediments nor in the distribution of facies. The eastern boundary of Silurian sediments is regarded erosional (Kaljo 1970). Also the anomalies of the magnetic and gravity fields and gradient zones are prevailingly of NW, NE or W-E orientation. The results of the most recent seismoacoustic studies do not confirm the tectonic origin of the lake basin either. The distribution of the Devonian thicknesses does not contour the lake basin. In general, the isopachous lines run either perpendicular to or across the lake, which shows that the lake basin is not a Devonian structure (Miidel *et al.* 1999).

Several researchers have stressed the great importance of glacial erosion on the development of the basement in the northwestern part of the East-European Plain. According to Makkaveev (1975), in northwestern Russia the glaciers removed a layer of rocks about 62 m, on lowlands even 80–100 m in thickness. Isachenkov (1982) has estimated glacial erosion at 60–70 m. The results of seismoacoustic studies show that in several parts of the Peipsi Basin glacial erosion was rather intensive. The origin of the basic features of the bedrock topography within L. Peipsi can be well understood if to consider lithological and geomorphological evidence concerning the dynamics of the glacier in the basin, the spread of the Vyborg rapakivi and Suursaari quartz porphyry (Raukas 1963), and the orientation of ice striae and pebbles in the till (Raukas *et al.* 1971) which show that the glacier intruded into the Peipsi Basin via the lowlands surrounding the Ahtme Elevation (Raukas *et al.* 1971, Tavast & Raukas 1982). A wedge-shaped eminence in the bedrock topography in the northern part of L. Peipsi formed evidently between two ice tongues in the shade of the Ahtme Elevation (Miidel *et al.* 2000). West of this eminence, a wide bedrock trough-shaped valley developed. Taking into consideration Tuuling's (1988a) data on the bedrock topography in the western part of the Narva-Luga Lowland, it is quite possible that a similar valley-like relief form developed also in the northeastern part of the Peipsi Basin.

The glacial erosion was at its highest in the central part of L. Peipsi where two ice flows got together (Miidel *et al.* 2000). This conclusion is supported by the distribution of the thickness of the Quaternary sediments, particularly till, in the basin (Noormets *et al.* 1998). The thickness of the Quaternary sediments increases from 5 m in the north to 50 m in the central part of the lake. The thickness of the till increases in the same direction, reaching 25 (max 33) m in the isometrical hollows eroded by ice (Fig. 21). It is noteworthy that the features of the bedrock topography are also well observable in the relief of the upper surface of the till (Fig. 18).

Basing on the above, it is supposed that the greater and deeper part of the lake basin formed as a result of glacial erosion, evidently during the course of several ice ages (Miidel *et al.* 2000). The amount of glacial erosion in the middle part of the lake basin versus the bedrock topography in the surrounding land may be estimated at least at 50–60 m, which agrees well with the results by Makkaveev (1975) and Isachenkov (1982).

2.3. BEDROCK TOPOGRAPHY

The Basin of Lake Peipsi which is conditionally bordered by 40–50 contour line is a negative form in the bedrock topography. Its boundaries are sometimes transitional. After Isachenkov (1982), the Peipsi Basin is the deepest part of the Pskov–Velokoretsk Lowland — a negative feature of wide areal extent.

The present knowledge on the bedrock topography in different parts of the meridionally elongated lake basin varies significantly. Owing to the exploration for oil shale, a wealth of data has become available on the bedrock topography in the northern part of the basin. The bedrock surface is reasonably well studied in several places on the north and east coasts of L. Peipsi and on the south coast of L. Pihkva.

Very little is known about the bedrock topography in an extensive swampy plain in the eastern coastal area of L. Pihkva and L. Lämmijärv and in a mire system around the mouth of the Emajõgi River. Until recently, there were almost no data as to the bedrock topography under the waters of L. Peipsi. Despite the investigations conducted in the western part of the lake (Noormets *et al.* 1998, Miidel *et al.* 2000), there is still a large area to be studied. Nevertheless, the data available to date (Sammet 1961, Kajak 1963, Sammet *et al.* 1967, Kotlyukova 1969, Isachenkov 1969, Malakhovskij & Bakanova 1971, Rähni & Tavast 1981, Tavast & Raukas 1982, Noormets *et al.* 1998 a.o.) allow description of main features of the bedrock topography in the lake basin (Fig. 16).

In the west and north, the lake basin borders on the Viru–Harju Plateau and the Narva–Luga Lowland. The latter has been considered to belong to the lake basin either entirely (Tavast & Raukas 1982) or partly (Kajak 1970, Raukas *et al.* 1971). The development of the Narva–Luga Lowland and the Peipsi Basin has been closely related. The lowland connects the Basin of Peipsi with the Basin of the Gulf of Finland. Therefore, it would be reasonable to treat the Narva–Luga Lowland or, at least its western area, as part of the Peipsi Basin. And the more, that there is not any pronounced morphological boundary between them.

According to Tuuling (1988a), the bedrock in the Narva–Luga Lowland is prevailingly 20–30 m a.s.l. In general, the bedrock surface descends towards the central part of the basin. The Narva Valley is not distinctly traceable in the bedrock topography. Against the background of the generally flat relief, an elongated branching and closed hollow is distinguishable under the present Puhatu Mire. In this negative feature, the bedrock surface sinks nearly to sea level. In the middle reaches of the Narva River, a flat valley falls into the Plyussa Valley, cut in the bedrock. Vasavere, on the eastern slope of the Ahtme Eminence, is the largest buried valley. The bottom of this buried valley is at a depth of 17 m b.s.l. Besides the above-named negative features, there are some smaller bedrock hillocks, irregular in shape and arrangement, which rise by 5–10, seldom more metres, above their immediate vicinity. The largest of those eminences are located in the surroundings of the Zagriv'e and Agusalu villages, where the bedrock is more than 45 and 35 m a.s.l., respectively. In a remarkable prolonged eminence on the western bank of the Plyussa River, opposite the Gostitsy Village, the bedrock surface rises to a height of 40 or even more metres a.s.l. Some hills are situated in the outcrop area of hard Ordovician carbonate rocks, *e.g.* at Zagriv'e.

In the north, the Narva–Luga Lowland is bordered by the North-Estonian Klint rising at this point up to 25-28 m above sea level. In the bedrock topography, the klint is expressed as a slope, not as a steep escarpment and is, therefore, inconspicuous in the present topography.

The Ahtme Eminence and the Pandivere Elevation - medium forms of the Viru–Harju Plateau – adjoin the Peipsi Basin. The Ahtme Eminence rises up to 76 m and the Pandivere Elevation more than 130 m a.s.l. (Kajak 1970, Raukas *et al.* 1971, Tavast & Raukas 1982).

The northwestern slope of the Pandivere Elevation lowers from 80 m to 40 m and passes, thereafter, smoothly into the lake basin. However, at some points the transition is rather steep. After Kiipli (1989), the slope descending from a height of 60 m towards the lake basin represents a surface dissected and denudated into carbonate rocks of the Raikküla and Juuru stages. Ten kilometres to the southwest from

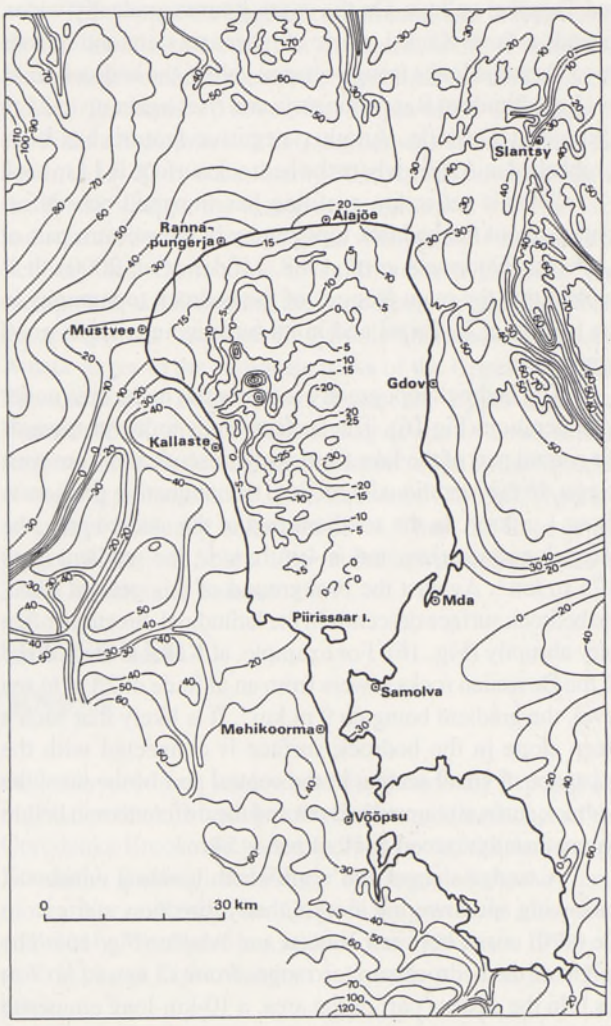

Fig. 16. Bedrock topography of the Peipsi Basin. Compiled using data by Kajak 1963, 1970, Sammet *et al.* 1967, Isachenkov 1969, 1969a, Malakhovskij & Bakanova 1971, Tavast & Raukas 1982, Miidel *et al.* 2000 a.o.

Mustvee Town, the height of the bedrock surface decreases abruptly – 49 m within 1.2 km. On the bottom of a valley, 16 km in width and 40–50 m in depth, several "spots" of Devonian rocks have been established (Kiipli 1989). The broad and flat valley reaches Mustvee where its bottom is 10–15 m a.s.l. Another buried valley was discovered some kilometres south of Mustvee. The bottom of this southwesterly oriented valley is from 13 m up to 27 m b.s.l. (Fig. 16, Eltermann & Raukas 1963). It was supposed that the valley continues under the waters of L. Peipsi in a northeasterly direction and extends as far as the Sahmen Bank (Tavast & Raukas 1982).

On the southern slope of the Ahtme Eminence, the bedrock surface lowers towards the lake depression to 15–20 m altitude. The sinuosity of 35 m and 40 m contour lines is indicative of a slope which is dissected by hollows opening to the south and south-east. In the surroundings of the Permisküla Village on the upper reaches of the Narva River, a flat valley starts and widens towards the south. The bedrock surface lowers to an altitude less than 15 m, *e.g.* at Vasknarva it is 13.6 m a.s.l. Still, the bedrock topography seems to be alternating. According to Vellner (1928), in several places the bedrock lies rather high (26.0–26.5 m a.s.l.). A shallow and trough-shaped valley occurs beneath the Rannapungerja and Tagajõgi valleys. On the coast, it turns gradually wider, extends as far as Kauksi, where a higher area starts and spreads up to Uusküla. In the trough-shaped valley, the bedrock sinks to 15 m altitude at Rannapungerja and rises again up to 25 m altitude at Uusküla. Another negative feature has been established at Alajõe where the bedrock surface is 14 m a.s.l.

Seismic reflection profiling has imparted new information about the bedrock topography in the western part of L. Peipsi (Noormets *et al.* 1998, Miidel *et al.* 2000). It is evident that the main features of the bedrock topography in the bottom of L. Peipsi and in its surroundings are in good agreement.

The bedrock topography is rather flat in the area under consideration (Fig.16). The surface tends to lower towards the central part of the lake from north to south and from west to east. In the meridional direction, the inclination gradient is about 1 m km^{-1}. In the southern part of the studied area, the bedrock surface rises and is 4 m b. s. l.; the gradient is *ca* 0.75 m km^{-1}. Against the background of this general trend, the bedrock surface descends in the latitudinal direction, often very abruptly (Fig. 16). For example, at Kallaste the surface of the Devonian rocks lowers from an altitude of 28 m to sea level, the gradient being *ca* 9 m km^{-1}. It is likely that such a steep slope in the bedrock surface is connected with the existence of small scarps; in the central part of the lake, the bedrock surface is smoothed out and the differences in height do not usually exceed 5–10 m per 4–5 km.

A wedge-shaped and north-south trending eminence, narrowing and lowering in a southerly direction, starts from the north coast, between Kauksi and Alajõe (Fig. 16). The height of the bedrock surface ranges from 13 m a.s.l. to 7 m b.s.l. In the eastern part of the area, a 10-km-long eminence rises above the surrounding bedrock surface by 5–8 m.

An up-to-7-km-wide trough occurs west of the wedge-like eminence. At first, this trough runs in a meridional direction, but some 7 km northeast of Kodavere it splits up forming two branches. One branch continues to the south-west towards Kallaste, while the other turns into a broad central depression (Fig. 16). The depth of the bedrock surface changes from 1–5 m b.s.l. in the north to 22 m b.s.l. at the turning point. The bottom of the central depression is 31 m b. s. l. The longitudinal profile along the trough's bottom is uneven. There are several, 10–20 m deep hollows in the central part of the trough.

A wide bedrock plateau between the lake's north-west coast and the above-mentioned trough has an altitude of 12–18 m. In the east, the plateau is bordered by a gentle slope which gradually turns into the trough's floor.

South of the central depression, the uprising bedrock forms a plateau from Devonian terrigeneous rocks, with the surface lying 9–2 m b.s.l. In the surroundings of the Island of Piirissaar, the bedrock surface reaches near sea level. The plateau slopes slightly westwards. In its northern and central parts, some small hollows and knobs with an amplitude in height from 2 to 5 m can be followed.

The seismic data obtained suggest a commonly flattened bedrock topography, and do not support the idea of the existence of a pre-Quaternary network of valleys in the bottom of the lake basin. Considering the dense grid of seismic profiles between the Sahmen Bank and Mustvee, it is hardly possible that any buried valley, starting from the Sahmen Bank and running towards Mustvee, could have remained unnoticed.

According to Eltermann and Raukas (1963), the bedrock is 13–12 m b.s.l. in the area 3 km south of Mustvee and 1.5 km away from the lake. If indeed there is a NE-SW oriented buried valley in this area, it would start from the mainland, not from the basin area, occupied by the lake.

The seismic profiles do not show any signs of meridionally oriented valleys in the Peipsi Basin. The wide trough established in the northern part of the studied area does not reach the southern part of the lake basin where a relatively high bedrock plateau is located. The area between Ranna and Kallaste with the bedrock surface 2 to 5 m b.s.l., could serve as a tributary for the above-mentioned trough. The tributary seems to continue further to the west and have a link with the buried valley suggested by Miidel and Tavast (1978) and Tavast and Raukas (1982) in the adjacent coastal area. In a borehole, made into the suggested buried valleys 11 km west of Kallaste, the bedrock is 54 m b.s.l. (Tavast & Raukas 1982).

Anyhow, assuming a connection between the above-named bedrock relief forms in the lake and in the adjacent coastal area, it is rather striking that the buried valley is much wider within the lake than in the coastal area. This may relate to the effect of the ice masses which concentrated in the lake causing there more intensive erosion (Peipsi ice-lobe, after Tavast & Raukas 1982). Probably, the cavities along the buried valley were also formed as a result of glacial erosion. The Middle-Estonian Lowland with L. Võrtsjärv is situated west of the Peipsi Basin. In the eastern part of the lowland the height of the bedrock is much the same as in the Peipsi Basin (25–30 m a.s.l.). For that reason, the boundary between these large features of the bedrock topography is transitional; after Tavast and Raukas (1982) it runs along the Tartu–Torma line.

In comparison with the adjacent areas, the bedrock is higher at Kallaste and in the north-west (47 m a.s.l.), and particularly high (more than 60 m a.s.l.) in the Kokora insular eminence east of Kallaste (Tavast & Raukas 1978). The eastern boundary of the Kokora Eminence forms a distinct slope in the present topography between Kallaste and Alatskivi. The waves of the Peipsi Ice Lake have cut some scarps into the slope (Mieler 1926, Liblik 1969). At Kallaste, the waves of L. Peipsi have cut a 5–8-m-high cliff into the Devonian rocks. According to Helmersen (1864), this happened in 1844.

The boundary between the lake basin and the Middle-Devonian plateau is also indefinite, particularly in the south of the Kokora Elevation. Within the Emajõgi mire system there is a hollow which opens to the east. In this hollow, the bedrock lowers to 7 m a.s.l. in the surroundings of Kavastu. Although the bedrock rises in a westerly direction, it remains as formely relatively low (15–20 m a.s.l.). A somewhat higher and narrow area between Kallaste and Võnnu, bordered by the 30 m contour line and reaching the lake, separates the depression around the Emajõgi River mouth from a valley which starts from Võnnu. At Mehikoorma, the bottom of this a-few-kilometre-wide valley is 7 m b.s.l. The boundary between the Middle-Devonian Plateau and the Peipsi Basin is traceable as a gentle slope along the Võnnu–Ahja–Mooste line. It is marked by the 40 and 50 m contour lines. Between Mooste and Treski, the boundary is again unclear, but from there up to L. Pihkva it is rather distinct. The bedrock lowers towards the east (20–30 m a.s.l.), but by reason of the wide distribution of swamps on the west coast of L. Pihkva, very little is known about the bedrock topography there. Valleys running to the west dissect the presumably even bedrock surface. One of those valleys is situated on the Vööpsu–Põlva line; in this valley the bedrock is 1 m a.s.l. at Vööpsu, 7 m b.s.l. at Räpina and 43 m b.s.l. at Põlva. In the Piusa Valley, the bedrock is 13.6 m b.s.l. at Vymmorski (Apukhtin & Sammet 1967).

In the following, some data about the bedrock topography east of L. Peipsi will be presented. The area from the source of the Narva River up to the Plyussa River is occupied by a flat Middle-Devonian Plateau rising 30–40 m a.s.l. East of the Plyussa River in the vicinity of the Gostitsy Village, the bedrock reaches 40–60 m a.s.l. The Plyussa River separates the area with small relative heights in the west from the higher area with a complicated topography and southeasterly orientated buried valleys (Sammet 1961).

The above-mentioned flat Middle-Devonian Plateau continues southwards up to the Dobrutsi Village where it is replaced by a northwesterly oriented eminence rising 45–70 m a.s.l. In the Luga Elevation, south of the eminence, the bedrock rises higher than 100 m above sea level (Isachenkov *et al.* 1982). The west slope (20–60 m a.s.l.) serves as a boundary between the plateau and the lake basin. In the west, there is a vast paludified plain. The extremely scanty data available allow only to speculate that the bedrock surface is there approximately at the height of the sea level (Fig. 16; Isachenkov 1969, Malakhovskij & Bakanova 1971, Isachenkov *et al.* 1982, Tavast & Raukas 1982).

Borehole records demonstrate that the height of the bedrock surface ranges within wide limits (40–50 m) which is indicative of a complicated bedrock topography. In the lower course of the Zhelcha River, the bedrock is at least 29 m b.s.l. (Malakhovskij & Bakanova 1971). For this reason, it is supposed that the river is connected with a buried valley (Tavast & Raukas 1982). There is a ridge at Pnevo. In a low cliff, the Devonian is outcropping. The bedrock reaches 31 m a.s.l.

The slope of the Luga Elevation continues in the vicinity of L. Pihkva's east coast. In the south, there is a distinct escarpment treated as the northern boundary of the Devonian Cuesta (Isachenkov 1969a). The height of the Devonian or the Pihkva Escarpment (klint), bordering L. Pihkva from the south, is 15–20 m, occasionally 30 m. In the surroundings of the Petski Village, the escarpment withdraws from the lake and runs further to Vashina Gora wherefrom it continues to Pechory and from there to southwestern Estonia. The escarpment was formed in the zone marking transition from the terrigenous rocks of the Middle Devonian Gauja and Amata stages to the carbonate rocks of the Upper Devonian Pļaviņas Stage. Deep valleys dissect the Upper-Devonian Plateau (45–60 m a.s.l.). The zigzag course of these valleys suggests that their development was controlled by the direction of tectonic joints (Bekker 1924, Isachenkov 1969a). The floor of valleys is usually below sea level. Thus, the floor of the Optjok River is at Sokha 13.6 m and at Vashina Gora 21 m b.s.l. (Apukhtin & Sammet 1967).

2.4. QUATERNARY COVER

2.4.1. THICKNESS

The thickness of the Quaternary deposits in the Peipsi Basin is highly variable (Fig. 17). In general lines, it increases from the surrounding bedrock elevations towards the basin and along it from north to south, being usually 5–10 metres in the northern part of the basin and less than a metre in some localities on the southwestern slope of the Pandivere Elevation and on the southern slope of the Ahtme Eminence. The Quaternary cover is thin in several places in the Narva–Luga Lowland, *e.g.* west of Permisküla Village where the deposits are 2–4 metres thick. The area with a thin Quaternary cover extends along the Narva River as far as the Gorodenka Brook. Some kilometres west of Kuningaküla Village there is evidently a bedrock knob where the Quaternary deposits are only 0.5 m thick. At many points in the middle reaches of the Gorodenka Brook, the bedrock crops out on the surface. At Kondushi, Radoveli and Zagriv'e villages on the east bank of the Narva River, the Quaternary cover is also thin, commonly less than 5 m, seldom 5 m (Mozhaev 1973). In several places in the vicinity of Agusalu it is less than 2 m thick.

In most of the Narva–Luga Lowland, the Palaeozoic rocks are covered with Quaternary deposits, ten or even more metres in thickness. The exploration for oil-shale revealed that in several places in the Puhatu Mire, the Quaternary

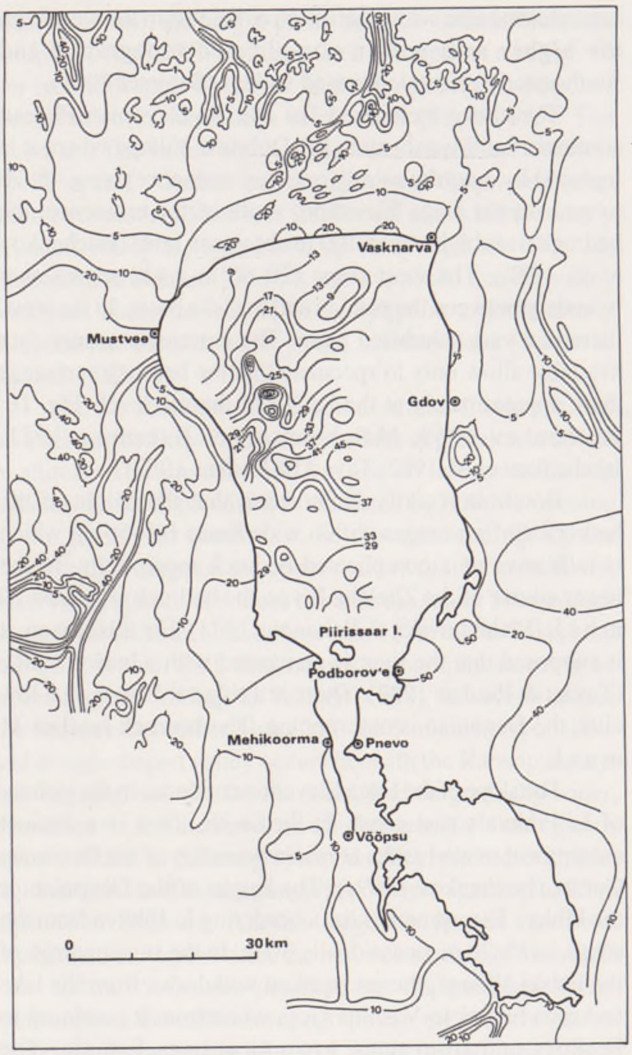

Fig. 17. Thickness of the Quaternary deposits. Based on the data by the Geological Survey of Estonia, Isachenkov 1969, Noormets *et al.* 1998.

deposits are 14–15, occasionally even 24 m thick. In a valley-like bedrock feature in the surroundings of the Poruni River and the Gorodenka Brook, the thickness of the Quaternary cover reaches 24 and 37 m, respectively; in several other locations it is only 5 m. The variations in the thickness of the deposits are connected with the changes in the bedrock topography.

The thickness of the Quaternary cover increases towards L. Peipsi's north coast, being *ca* 18 m at Vasknarva, 12 m at Smolnitsa, 16.8 m at Alajõe, 17.7 m at Kauksi, 22 m at Rannapungerja and 16.6 m at Lohusuu. Seismic reflection profiling in the western part of L. Peipsi has revealed (Noormets *et al.* 1998) that the Quaternary cover continues to thicken on the lake bottom (Fig. 17) towards the deeper, central part of the lake and also southwards along the lake. In the meridionally elongated glacially eroded trough-like valley, the thickness of the Quaternary deposits increases from 15 to 40 m. The data obtained by means of the only borehole made in the lake, show that the thickness of the Quaternary cover is 27.6 m at the Sahmen Bank (Kajak 1964). The greatest thicknesses have been established in the central part of the lake (50 m) and locally in hollows of the meridional trough (up to 60 m). In the lake, off the west and north coasts, the Quaternary deposits are 10–15 m thick. Along the west coast, the thickness of the Quaternary cover increases rapidly up to 30 m. From the central part, it starts to diminish remaining, nevertheless, greater than 30 m. The distribution of the thicknesses is in good accordance with the basic features of the bedrock topography (flat plateau in the west, meridional wide valley, central depression, *etc.*) (Fig. 16, Miidel *et al.* 2000). The bedrock topography is reflected even in the till surface topography (Fig. 18), in the thickness of glaciolacustrine and lake deposits and, to a certain extent, also in the present lake bottom (Figs. 19, 20). This suggests that the total amount of glacial erosion was not compensated by the glacial or post-glacial sedimentation (Noormets *et al.* 1998). The thickness of the till layer (Fig. 21) ranges from 5 to 17 m (maximum 31.5 m). The Quaternary cover is thicker east of Kallaste, in the central part of the lake and in the immediate vicinity of Piirissaar Island, forming wide and flat east-west trending ridges. Glaciolacustrine and post-glacial lacustrine deposits are at their thinnest east of Mustvee where in a large area their thickness is less than a metre and, in places, till crops out both in the coastal zone and at some distance off the shore. The thickness increases up to 25 m in the meridional trough and up to 36 m in the central depression. Southwards, the thickness of glaciolacustrine and postglacial lacustrine deposits decreases again, being only 4–10 m at the mouth of the Emajõgi River (Noormets *et al.* 1998).

The greatest thicknesses of the Quaternary cover in the northern part of the lake basin are associated with glacial accumulative landforms and buried valleys (Kajak 1964). Thus, in the Kuremäe End Moraine the deposits are more than 35 m and in the Iisaku–Illuka marginal glacial ridge more than 45 m thick (Kajak 1963). In radial eskers (Mäetaguse, *etc.*) and plateau-like kames (in the vicinity of Illuka), the deposits are usually 10–15 m thick. The greatest thickness of the Quaternary deposits (more than 75 m) has been recorded east of Kurtna in the area where the Kurtna Kame Field is situated above the Vasavere buried valley. In that valley, the average thickness of the deposits is 35–45 m.

Alternation of areas with thin and thick Quaternary cover is characteristic of the southern part of the Peipsi Basin (Fig. 17). On the southeastern slope of the Pandivere Elevation, the Quaternary cover shows significant thickening. In places, the thickness of the deposits, mostly of Pleistocene age, reaches 96 m, the average being 40–50 m (Kiipli 1989). The Quaternary cover is also thick on L. Peipsi's coast amounting to 45 m a few kilometres to the south from Mustvee and *ca* 70 m farther in the south-west (Eltermann & Raukas 1963). All the above-mentioned great thicknesses are related to the buried valleys in the bedrock relief and, in part, also to the marginal glacial formations in the surroundings of Töljase, Nautrase and Kaiu.

Immediately on L. Peipsi's coast, between Omedu and Alatskivi and between Mehikoorma and Värska, the average thickness of the Quaternary cover is 5 m which is less than elsewhere. In a narrow coastal belt between Mehikoorma and Molozhva, the deposits are again more than 5 m thick; in mires probably 10 or more metres. The usual thickness is still 5–10 m, and very rarely less than a metre, In the Emajõgi mire system, the thickness of the Quaternary cover supposedly exceeds 20 m, but authentic data are lacking.

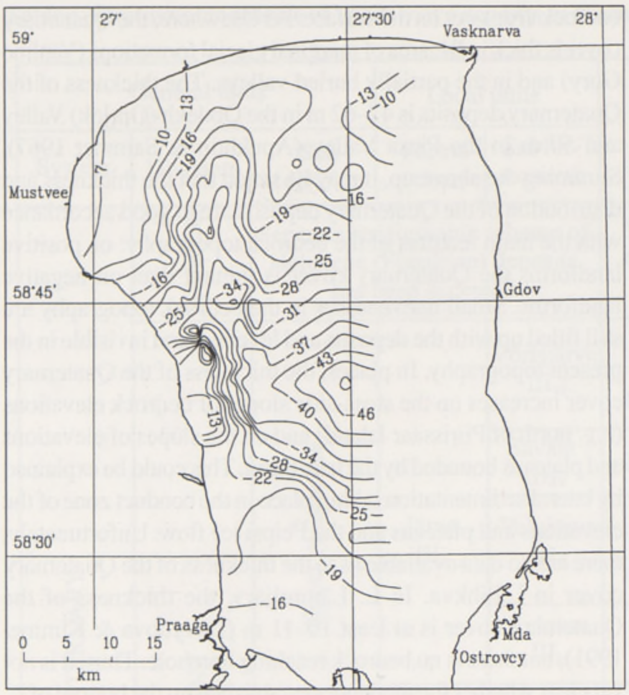

Fig. 18. Map of till topography (Noormets *et al.* 1998). Contours show the depth below lake level. The altitude of the lake-level is 30 m a.s.l.

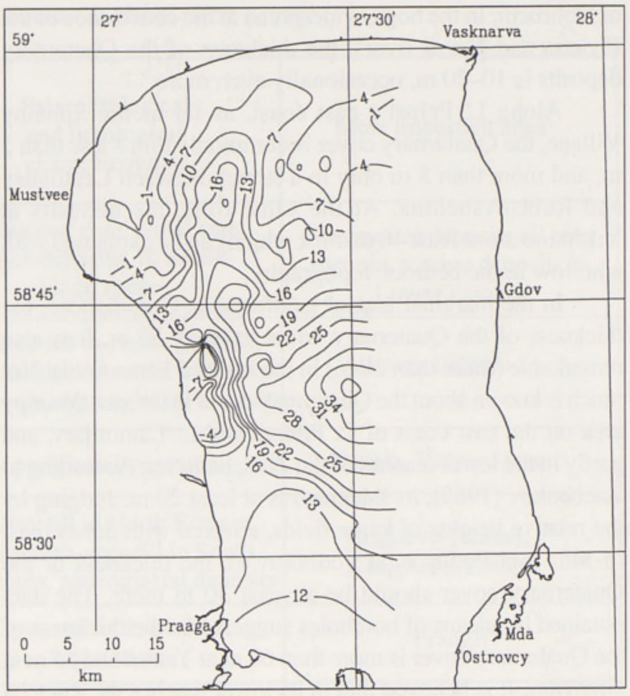

Fig. 19. Map of thickness of the glaciolacustrine and postglacial lacustrine deposits (Noormets *et al.* 1998).

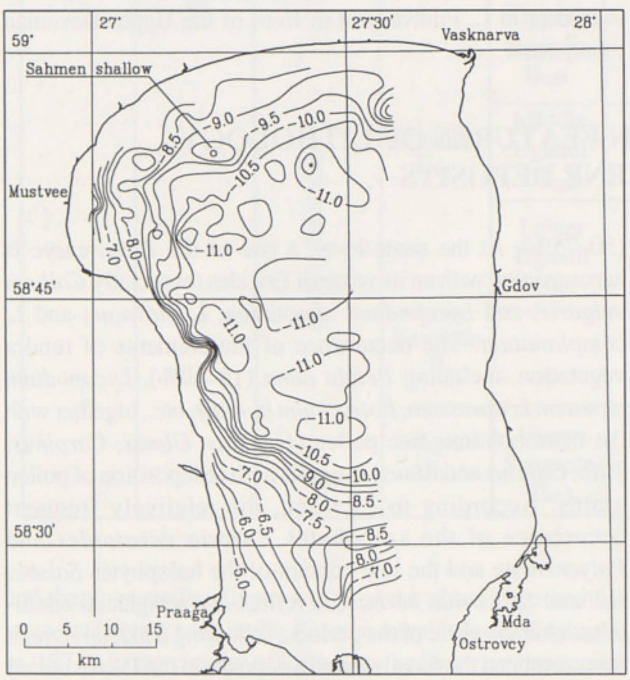

Fig. 20. Bathymetry map of L. Peipsi (Noormets *et al.* 1998).

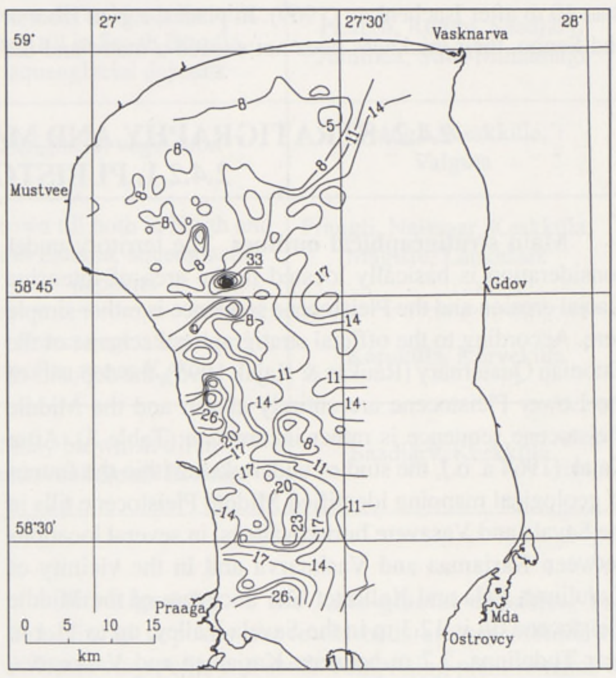

Fig. 21. Map of thickness of the till (Noormets *et al.* 1998).

On marginal glacial landforms, the Quaternary cover is 5–10 m thicker than on surrounding plains. At Lis'e it is at least 15 m thick. On the Mehikoorma–Aravu End Moraine, the Quaternary deposits are evidently of the same thickness, but the landform is situated partly in a buried valley. At Mehikoorma, the Quaternary cover is 39.5 m thick in the valley where almost the whole section consists of glaciolacustrine deposits. In the vicinity of Aravu within the same valley, the thickness of the Quaternary deposits reaches 46 m. The deposits are at their thickest just in the buried valleys (62 m near Räpina, 65 m at Savostvere, 54 m at Mikitamäe).

The thickness of the Quaternary cover in the eastern part of the Peipsi Basin is very poorly known, except the relatively well-studied oil shale mining area in the north. In a large area between Lake Peipsi and the Plyussa River, the thickness of the Quaternary deposits does not exceed 5 m (Fig. 17). The smallest thicknesses (average 5 m) occur east of the Plyussa River in a meridional belt running from the Slantsy Town to Gostitsy Village, but also in some northwesterly oriented belts on the left bank of the Cherma River, and west of the Plyussa River. In some localities, the Quaternary cover is less than a metre thick, *e.g.* in the vicinity of Dobruchi. In the bogs, widespread at the confluence of the Plyussa and Narva rivers, the thickness of the Quaternary deposits is 10–20 m, occasionally even more.

Along L. Peipsi's east coast, as far as the Spitsiny Village, the Quaternary cover is commonly thin – less than 5 m, and more than 5 m only in a stretch between Laptovitsy and Rubtsovshchina. At the same time, the deposits at Yushkino are at least 40 m thick which can be explained with a hollow in the bedrock topography.

In the marginal glacial landforms at Gorodishche, the thickness of the Quaternary cover exceeds 30 m. It is also remarkable (more than 20 m) in the isolated kame fields. Not much is known about the Quaternary cover in the vast swampy area on the east coast of L. Pihkva and L. Lämmijärv, and partly in the lower reaches of the Zhelcha River. According to Isachenkov (1969), its thickness is at least 20 m. Judging by the relative heights of kame fields, covered with dunes (east of Mda and Podlip'e, at Podborov'e), the thickness of the Quaternary cover should be at least 30 m there. The data obtained by means of boreholes suggest that the thickness of the Quaternary cover is more than 60 m at Yamm and 55 m at Nizovitsy. It is believed that in its lower reaches the Zhelcha River flows above a buried valley (Tavast & Raukas 1982, Isachenkov *et al.* 1982). In the area, bordering on the south tip of L. Pihkva, the Quaternary deposits are rather thin – less than 10 m after Isachenkov (1969). In places, *e.g.* at Izborsk and Senno, their thickness does not exceed a metre and the bedrock crops out on the surface. As elsewhere, the Quaternary cover is thick in the area of marginal glacial formations (Vaulino Gory) and in the partially buried valleys. The thickness of the Quaternary deposits is 47–62 m in the Obdekh (Optjok) Valley and 50 m in the Piusa Valley (Apukhtin & Sammet 1967). Summing the above up, it may be stated that the thickness and distribution of the Quaternary deposits are in good accordance with the main features of the bedrock topography: on positive landforms the Quaternary cover is thinner than on negative landforms. Small unevenesses in the bedrock topography are still filled up with the deposits and levelled, and invisible in the present topography. In places, the thickness of the Quaternary cover increases on the stoss-side slopes of bedrock elevations (*e.g.* north of Piirissaar Island) and on the slopes of elevations and plateaus bounded by the lake basin. This could be explained by lateral sedimentation taking place in the conduct zone of the elevations and plateaus and the Peipsi ice flow. Unfortunately, there are no data available as to the thickness of the Quaternary cover in L.Pihkva. In L. Lämmijärv, the thickness of the Quaternary cover is at least 10–11 m (Davydova & Kimmel 1991), but there is no bedrock reaching borehole. Thus, it is not yet clear whether the tendences established on the bottom of L. Peipsi, continue southwards. Considering the small thickness of the Quaternary cover on the Devonian plateau in the coastal zone of L. Peipsi, it could be supposed that the Quaternary cover is thicker in L. Pihkva, just in front of the Upper Devonian escarpment.

2.4.2. STRATIGRAPHY AND MAIN FEATURES OF LITHOLOGY

2.4.2.1. PLEISTOCENE DEPOSITS

Main stratigraphical outlines. The territory under consideration is basically located in the area of intensive glacial erosion and the Pleistocene sequence is rather simple here. According to the official stratigraphical scheme of the Estonian Quaternary (Raukas & Kajak 1995), the deposits of the Lower Pleistocene are entirely absent and the Middle Pleistocene sequence is rather incomplete (Table 3). After Kajak (1964 a. o.), the studies undertaken within the frames of geological mapping identified Middle Pleistocene tills in the Savala and Vasavere buried valleys, in several locations between Karjamaa and Vasknarva and in the vicinity of Tudulinna, Pala and Kallaste. The thickness of the Middle Pleistocene till is 12.3 m in the Savala Valley, up to 11.1 m near Tudulinna, 3.7 m between Karjamaa and Vasknarva, 1.4 m near Kallaste and 0.2–0.3 m near Pala. In the Savala Valley, different till beds at a depth of 25.8–30.2 m are separated with glaciolacustrine clays which contain spores and pollen and plant remains of unestablished age.

Palynological investigations by Liivrand (1991 a.o.) have demonstrated the dominance (40–60%) of arboreal pollen, the main part of which is represented by *Betula*, especially *B. nana*, and *Pinus* (up to 40%). The proportion of *Picea* pollen is 10–25%. The content of herbs is low (18–32%). Artemisia (10–55%), Chenopodiaceae (8–18%), Gramineae (up to 50%) and Cyperaceae (mainly 30–40%) are present in the whole intertill layer and also in the both till beds (Fig. 22). Of the *Betula* pollen in the depth interval 27.4–30.0 m, the majority belongs to *Betula sect. Albae* (50–75%). At the same level, a rise of the *Picea* curve is accompanied with an increase of Ericales (especially *Calluna vulgaris*) and *Lycopodium annotinum*, *L. clavatum* and *L. complanatum*. The occurrence of the elements of tundra vegetation, including *Betula nana* (18–40%), *Lycopodium alpinum*, *L. apressum*, *Botrychium boreale*, *etc*., together with the thermophilous tree pollen (*Quercus*, *Ulmus*, *Carpinus*, *Tilia*, *Corylus* and *Alnus*) suggests high redeposition of pollen grains. According to Liivrand, the relatively frequent occurrence of the xerophytes *Eurotia ceratoides* and *Polycnemum* and the low content of the halophytes *Salsola kali* and *Salicornia herbaceae* reflect dry periglacial conditions, characteristic of the periods, following a glacial retreat. She correlated the Savala intertill deposits in the Purtse buried valley with the Middle-Weichselian interstadial, when trees were present but did not form closed forests. According to that interpretation, the lower till bed is either Lower/Middle Weichselian or Warthian.

In several sections, two different till beds of the last glaciation cover each other (Roostoja, Tudulinna) or are separated with water-laid sediments of different thickness, in some places (*e.g.* at Lis'e) with annually laminated up-to-9-m-thick glaciolacustrine varved clays (Piipenberg 1935). In most cases, these till beds have a clearly different colour and lithological composition which means that the accumulation of tills was affected by different ice movements.

In the northern part of the lake basin, the different till beds are traditionally correlated with the older Sakala, Otepää

Table 3. 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, Vääna-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

and the younger Pandivere stades or the older Luga and the younger Neva stades in the Russian terminology (Raukas *et al.* 1971). In the vicinity of Roostoja, the lower carbonaceous till is grey, while the upper one is reddish-brown and entirely devoid of carbonate grains and pebbles (Raukas 1963). Distinctly different is also the mineral composition of both tills (Raukas 1961). The accumulation of the lower till was probably affected by a south-easterly oriented ice flow, while the accumulation of the upper till was controlled by a south-westerly oriented ice flow, which crossed the Devonian outcrops near the Narva River. This is in good agreement with the results of till fabric analysis, direction of glacial striae and orientation of glacial relief forms. In the southern part of the lake basin, lithological differences between the till beds of different age are not so clear. It is also possible that a two- or three-layered till sequence may consist of coeval basal and ablation till originating from a single glacial event (Raukas 1963).

Stratigraphy of the Late-glacial deposits. The Late-glacial deposits above till beds are represented by different clays, silts and sands comprising plant remains or interlayers, which can be, at least partly, redeposited. This hampers their radiocarbon dating. As those deposits often rest upon the limy till, the dating results are significantly affected by the hard water and reservoir effect. If the southernmost part of the lake basin was freed from the ice cover some 13,500 yr BP, then in Bølling time (13,200–12,200 yr BP) there were severe climatic conditions with fast accumulation of glacial and aqueoglacial sediments. Local vegetation just set in to develop and the role of redeposited pollen in sediments was high. That is why the Bølling sediments in this region do not reveal any clear palynological characteristics and the lower boundary of the Older Dryas cannot be palynologically defined (Pirrus & Raukas 1996). In the Remmeski Bog near Vastseliina (Fig. 23), there are

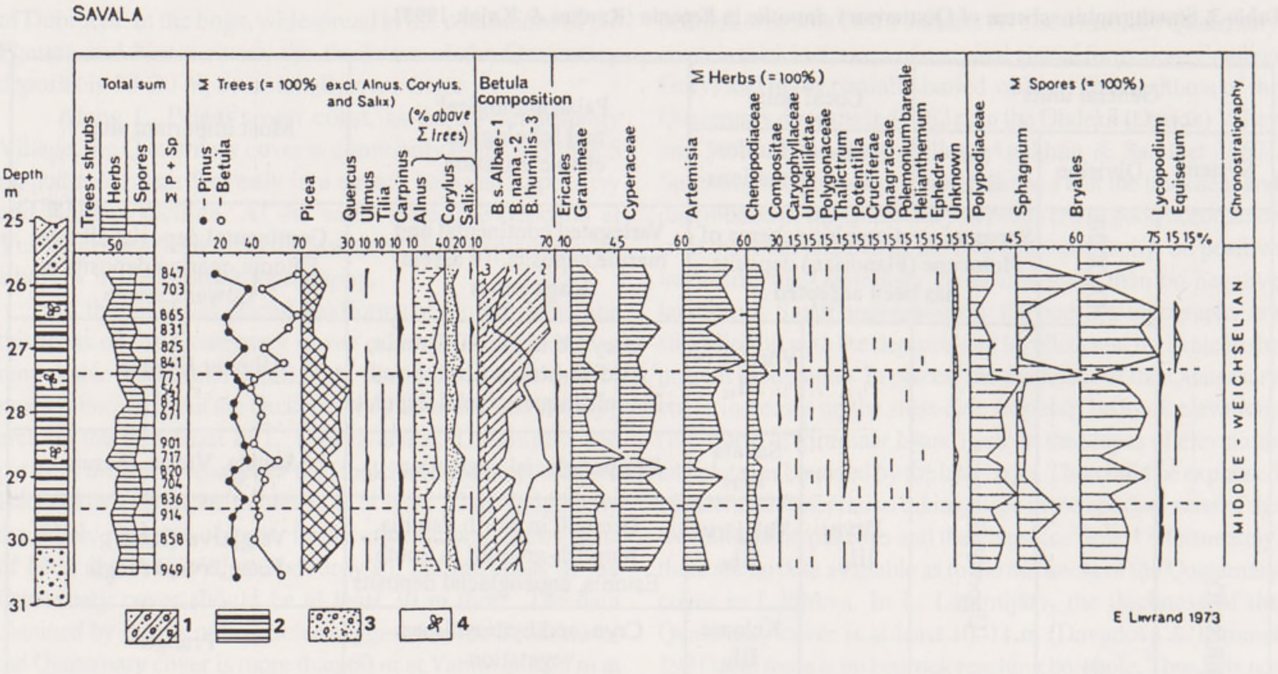

Fig. 22. Pollen diagram from the Middle-Weichselian intertill deposits in the Savala Valley (borehole 7854) after Liivrand (1985, 1991): 1 – till; 2 – varved clay; 3 – gravelly sand; 4 – plant remains.

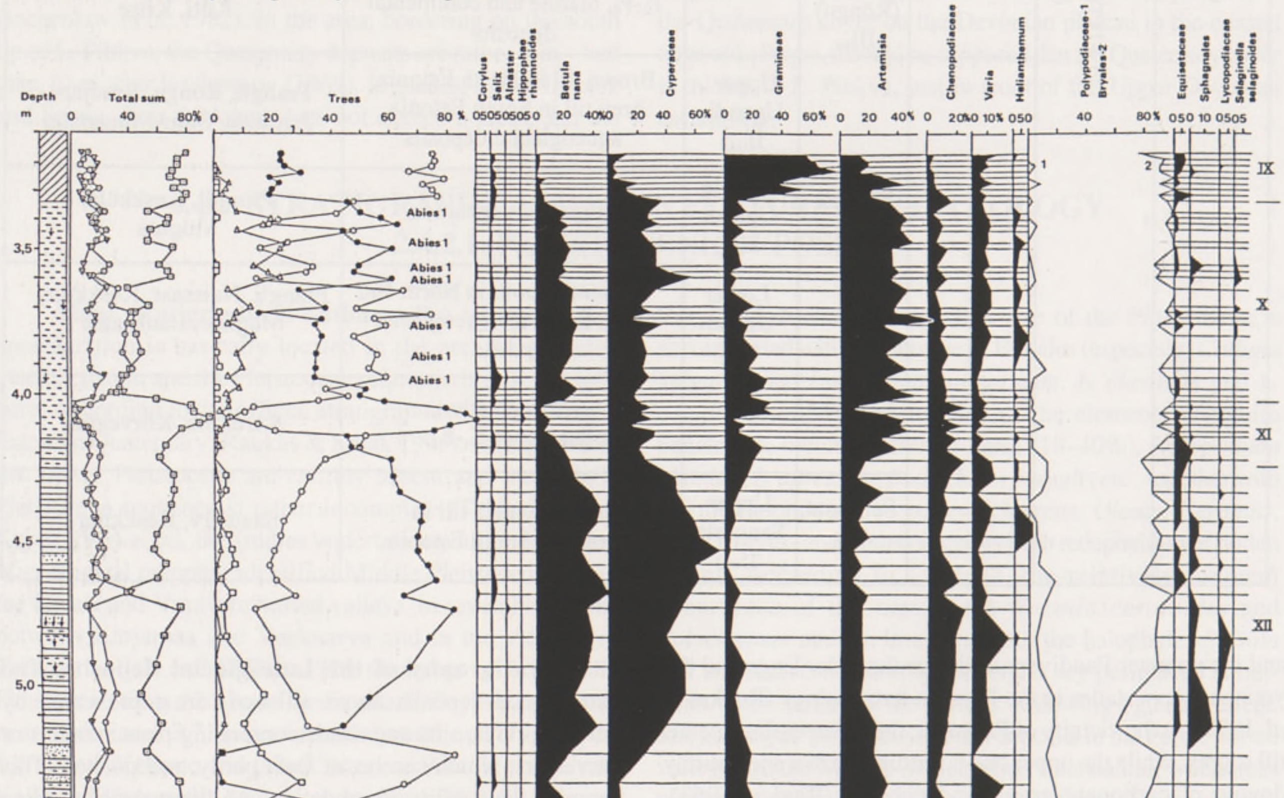

Fig. 23. Pollen diagram of Late-glacial deposits in the Remmeski Bog after R. Pirrus (1969).

some indirect evidence to indicate that the lower part of the section is Bølling (Pirrus 1969); however, this evidence is still very thin to allow any far-going conclusion to be drawn.

For the stratigraphical subdivision of the Late-glacial sediments we have used the official chart of Estonia (Pirrus & Raukas 1996), presented in Table 4.

The Older Dryas (12,200–11,800 yr BP) sequence is more or less complete only in the southern part of the lake basin. It consists of varved clays or rhytmically laminated silts and sands which may contain minor amounts of plant remains. The sediments have a high (20–30, in the southern part of the territory 30–40) herb pollen percentage (Artemisia, Chenopodiaceae, Helianthemum, Cyperaceae, Gramineae, and several other species of primary vegetation along with *Betula nana L.*). Artemisia dominates forming up to 40%. The proportion of Chenopodiaceae ranges from 5 to 10, Cyperacea form 10–20, and Gramineae up to 5%. Among tundra species, *Betula nana L.* has a high percentage (10%,

Table 4. Stratigraphic scheme of Estonian Late-glacial deposits (Pirrus & Raukas 1996).

Chronological scale, yr BP	Main subdivisions	Subformations	Chronozones		Index		Definition of boundaries, yr BP	Palynozone	Index	Baltic Sea stages
	Holocene	Lower Holocene	Preboreal		PB	PB_2	9 500	*Betula*	B	Ancylus Lake
10 000						PB_1	10 000	*Betula-Pinus*	B-P	Yoldia "sea"
10 500	Pleistocene	Võrtsjärve	Subarctic	Younger Dryas	DR_3		10 800	*Artemisia - Betula nana*	Ar-Bn	Baltic Ice Lake
11 000				Allerød	AL	AL_b	11 300	*Pinus*	P	
11 500						AL_a	11 800	*Pinus-Betula*	P-B	
12 000			Arctic	Older Dryas	DR_2		12 200	*Artemisia Chenepodiaceae*	Ar-Ch	Local ice Lakes
12 500 13 000				Bølling	BØ		13 200	*Betula-Cyperaceae*	B-Cy	
13 500				Oldest Dryas	DR_1					

max. 25–30%). Besides, the pollen of *Dryas octopetala L.*, *Rubus chamaemorus L.*, and spores of *Selaginella selaginoides* (L.) Link, *Botrychium boreale* (Fr.) Milde have also been found (Pirrus & Raukas 1996). The occurrence of xerophilous and halophilous pollen is remarkable. The pollen composition of the underlying till suggests that *Alnus, Corylus, Q. mixtum* and *Picea*, partly also *Betula* and *Pinus* are redeposited. In the constituent of trees, *Pinus* dominates over *Betula* (Pirrus 1969).

The Allerød Chronozone (11,800–10,800 yr BP) is represented by lacustrine clays and silts with dark-grey interlayers. Scattered plant remains, mostly leaves and stalks of *Bryales* moss, are common. The Allerød Chronozone in Estonia is subdivided into two parts: (a) *Pinus–Betula* Zone and (b) *Pinus Zone* (Pirrus & Raukas 1996). The lower boundary of the Allerød Chronozone is fixed with a rather distinct increase of AP pollen and decrease of herbs (Artemisia, Chenopodiaceae) and *Betula nana L.* The lower *Pinus–Betula* Palynozone is described only in the southern part of the area, the upper Pinus Zone is spread all over the territory. The boundary between the above two palynozones is fixed with a distinct increase of *Pinus* (80–90%) and a continuous decrease of herbs and *Betula nana L. Picea* is constantly present in low values (about 5%). The variety of species of mesophilous bog and meadow terrestrial herbs is already remarkable.

The Younger Dryas Chronozone (10,800–10,000 yr BP) is represented all over the lake basin by lacustrine silts and clays, often with fine sand interlayers. Plant remains (mainly *Bryales*) are scattered or occur as thin layers, occasionally abounding in hydrotroilite (Pirrus & Raukas 1996). The lower zone boundary is placed at the strong and rapid increase of the content of herb pollen and *Betula nana L.* The chronozone is characterized by a remarkably high frequency of herb pollen (*Artemisia* is the dominant) ranging from 40–60%. Among the pollen and spores of xerophilous herbs and tundra plants, the species present already in the Older Dryas are found in considerably increased amounts (Pirrus 1971). Among the trees, *Betula* dominates in the southern part and *Pinus* in the northern part of the area. *Picea* has a regular but rather different frequency: 26% at Remmeski and a few percents at Saviku. The upper boundary of the chronozone is placed at the rapid increase of tree pollen, prevailingly *Betula* (in the southern part of the territory up to 90%), and *Pinus* (about 20%). Among birches, *Betula sect Albae* prevails.

At the end of the Younger Dryas about 10,300–10,200 yr BP, organic sediments started to deposit in lakes instead of sand, silt and clay, the accumulation of which had clearly prevailed during the Late-glacial. The oldest organic sediments with an age of 10,200±90: TA-328 (Sarv & Ilves 1975) in the lake basin were dated in the Saviku section near the mouth of the Emajõgi River (Fig. 24).

Lithology of tills. In the Peipsi Basin, different genetic types of tills have been identified. Prevailing are glacial diamictons deposited by glaciers directly on the ground (subaerial tills), and among them lodgement, supraglacial (ablation) and proglacial (frontal) varieties are present. In order to clarify the role of subaquatic tills in Estonia, complex micropalaeontological, geochemical and structural investigations of tills were carried out. The results showed that tills in Estonia are predominantly subaerial in origin (Raukas 1973).

The occurrence of marine fauna is usually considered as a more reliable indication of the subaquatic formation of tills. None of the twenty seven sites studied in the lake basin, revealed any traces of marine fauna. Only the Lohusuu site yielded five foraminifers: one specimen of *Buccella frigida* (Cushman), one specimen of *Protelphidium orbiculare* (Brady) and three specimens of *Elphidium subclavatum Gudina*, probably redeposited from the Gulf of Finland's floor, where long-term marine conditions existed during the Eemian.

Both the geochemical and most of lithological evidence suggest subaerial genesis of tills. The evidence include the great similarity between the lithological composition of tills and underlying rocks, poorly sorted sediments, lack of authigenic marine inclusions, poor rounding of clasts and an

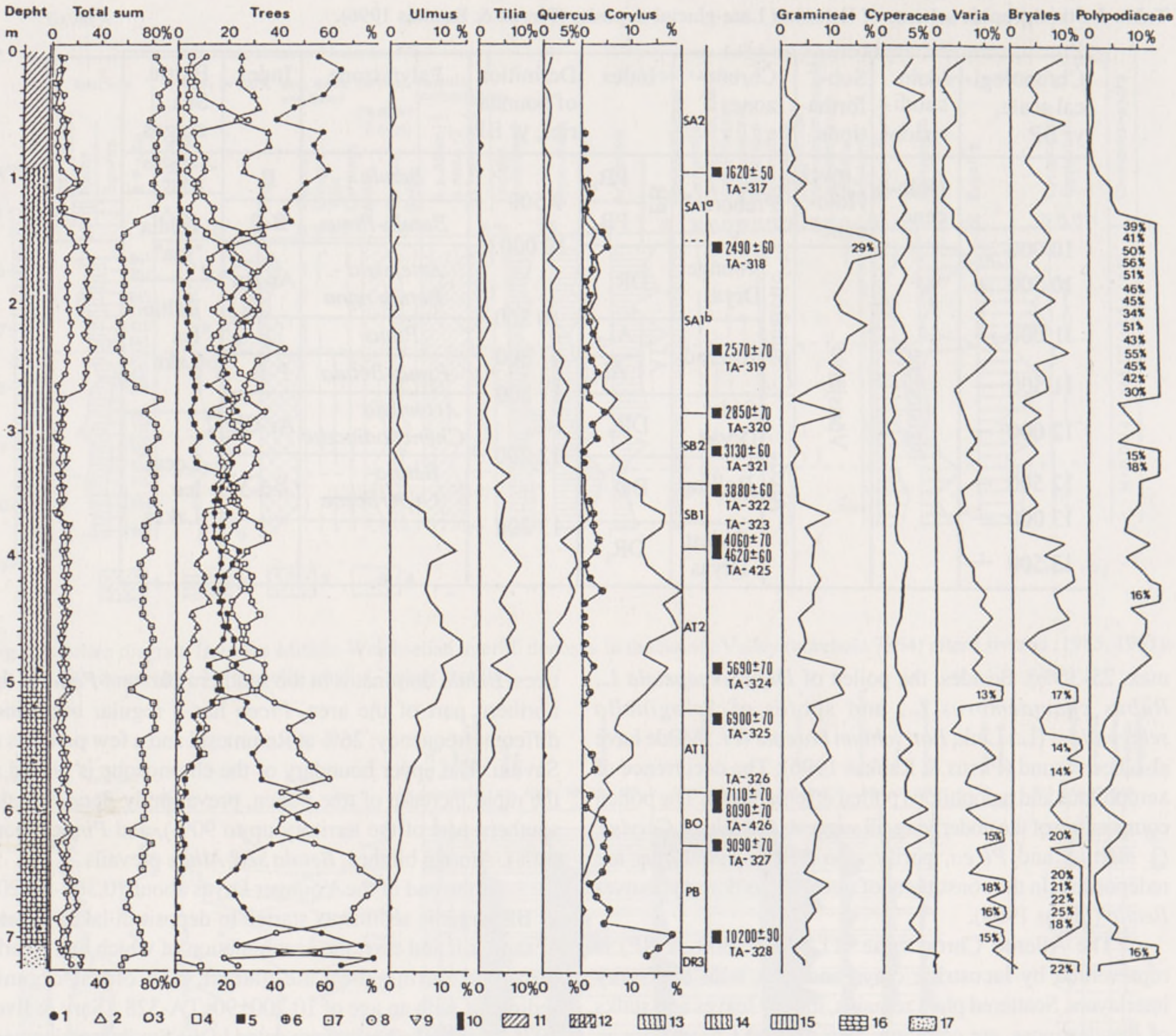

Fig. 24. Pollen diagram of Holocene deposits in the Saviku section near the mouth of the Emajõgi River after A. Sarv (Ilves & Sarv 1975): 1 – *Pinus*; 2 – *Picea*; 3 – *Betula*; 4 – *Alnus;* 5 – broad-leaved trees; 6 – *Salix*; 7 – trees and shrubs; 8 –herbs; 9 – spores; 10 – ^{14}C dates; 11 – sedge peat; 12 – reed-sedge peat; 13 – reed peat; 14 – wood-reed peat; 15 – woody peat; 16 – reed gyttja (sapropel); 17 – sand.

increase in roundness towards the south and southeast (*i.e.* in the direction of the movement of continental ice), orientation of clasts in the direction of ice advance (parallel to glacial striae and almost horizontal in position), occurrence of glaciotectonic structures (Photo 13), *etc.* Only in a very few places we have found laminated subaquatic varieties of tills (Photo 14), transitional between glacial and glaciolacustrine sediments, and iceberg tills, covering top parts of some kame hillocks (Photo 15). Iceberg-transported are probably also erratic boulders on top of aqueoglacial relief forms (Photo 16).

In all likelihood, subaerial tills were formed both beneath the advancing ice (lodgement till) and also as a result of bottom melting from passive ice during glacial downwasting (basal melt-out till). Deformation tills with different glaciodynamic structures are quite frequent (Kul'e, Lis'e). They are formed by drag beneath the sole of the moving glacier ice. They are more sandy and contain more pebbles and cobbles than typical lodgement tills. Small rafts are often packed into deformation till.

Proglacial or frontal tills are similar to deformation tills. They are related to push and dump end moraines. End moraines of complicated internal structure (moraines with traces of ice pressure overlain by glaciofluvial deposits, or *vice versa*) also occur, as well as "interior peripheral moraines" of push character, developed between dead and active ice (Raukas *et al.* 1971). Often frontal tills consist of lodgement, deformation and flow till facies squeezed and associated with injections of meltwater deposits and bedrock erratics.

The terrain in the lake basin is flat. As a result, the content of supraglacial material in tills is quite low. Ablation tills, resulting from surface melting and gravity flow of supraglacial and englacial debris are extremely rare. They occur only in some places on the top and slopes of the Iisaku–Illuka Esker System and in the hilly topography near Pechory Town. The thickness of the ablation till is 0.5–1.5 m only and it contains much less fine (silt and clay) fractions than a typical basal till.

As is well known, the lithological and mineral composition of tills depends on a number of factors, such as the composition and relief of the underlying bedrock, the dynamics of ice movement, the location of material within the glacial body,

Photo 13. Glaciotectonic structures in the Alatskivi Esker. Photo by A. Miidel.

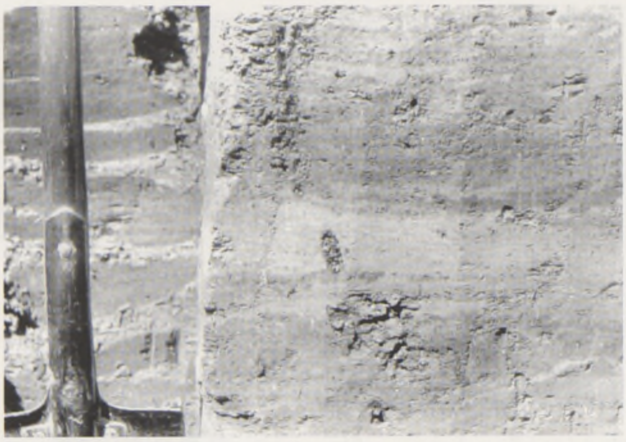

Photo 14. Laminated water-laid till at Kul'e. Photo by A. Raukas

Photo 15. Till, probably iceberg transported, on top of a kame hillock in the Illuka Kame Field. Photo by A. Raukas.

Photo 16. Erratic boulder on top of a kame hillock in the Illuka Kame Field. Photo by A. Raukas

the character of accumulation, the nature and intensity of the weathering of material, *etc*. (Raukas 1978, Raukas 1995).

Numerous investigations have demonstrated that during the flow the ice sheets incorporated quantities of local bedrock material. Moreover, the maximum content of rock particles (about 60–80%) from any particular bedrock stratigraphical unit is usually found close to the distal boundary of the outcrop of that unit. At a distance of 6–8 km from the bedrock unit boundary, the amount of rock particles from the corresponding unit does not exceed 20–30% of total (Raukas 1978).

Local tills, consisting entirely or almost entirely of local sedimentary material, are not widespread. They have been found near Kuremäe, in the middle course of the Narva River and to the west from Pskov. The tills of the last glaciation in the lake basin contain more crystalline rocks than those in the neighbouring areas. At the same time, basal tills composed more or less entirely of englacial debris (the so-called erratic tills), are rare. The tills, practically devoid of carbonate particles, have been found north of Räpina.

The vast majority of basal tills belong to the intermediate group between local and erratic types, containing local and allochthonous (far-transported) material in varying ratios. According to Gaigalas (1969), they should be termed transitional lodgement (or basal) tills; in these tills the local sedimentary material is the dominant component. However, in the southern part of the lake depression, coarse fractions are often prevailed (80–90% of all clasts) by the erratic component.

The tills in the Peipsi Basin demonstrate to what extent the granulometric, lithological, mineral and chemical composition, and the colouring of the tills is determined by the features of the underlying bedrock along which the continental ice moved. Thus, in the northern part of the lake basin, where the bedrock mainly consists of Ordovician and Silurian limestones and dolomites, the tills are grey with different shapes of colour. In a restricted area close to the Devonian outcrops, reddish-brown tills, enriched with the sandy-silty material from the reddish-brown Devonian sandstones are spread.

The granulometric composition of tills is determined by glacial ploughing of bed, crushing and abrasion of drift load, and by the mixture of the material during its transport and deposition. In brief, the granulometric composition of tills reflects the route covered by the glacier and the dynamics of its movement. In comparison with the till on the Devonian sand- and siltstones in the southern part of the lake basin, the grey-coloured tills on the carbonaceous bedrock are richer in boulders, pebbles (Fig. 25-A) and gravel (1–10 mm), whereas the underlying carbonaceous rocks contribute mainly to the coarse fractions of the till (Photo 17). As a result, the local carbonaceous material prevails in the coarse-grained fractions of the till. The Fennoscandian magmatic and metamorphic

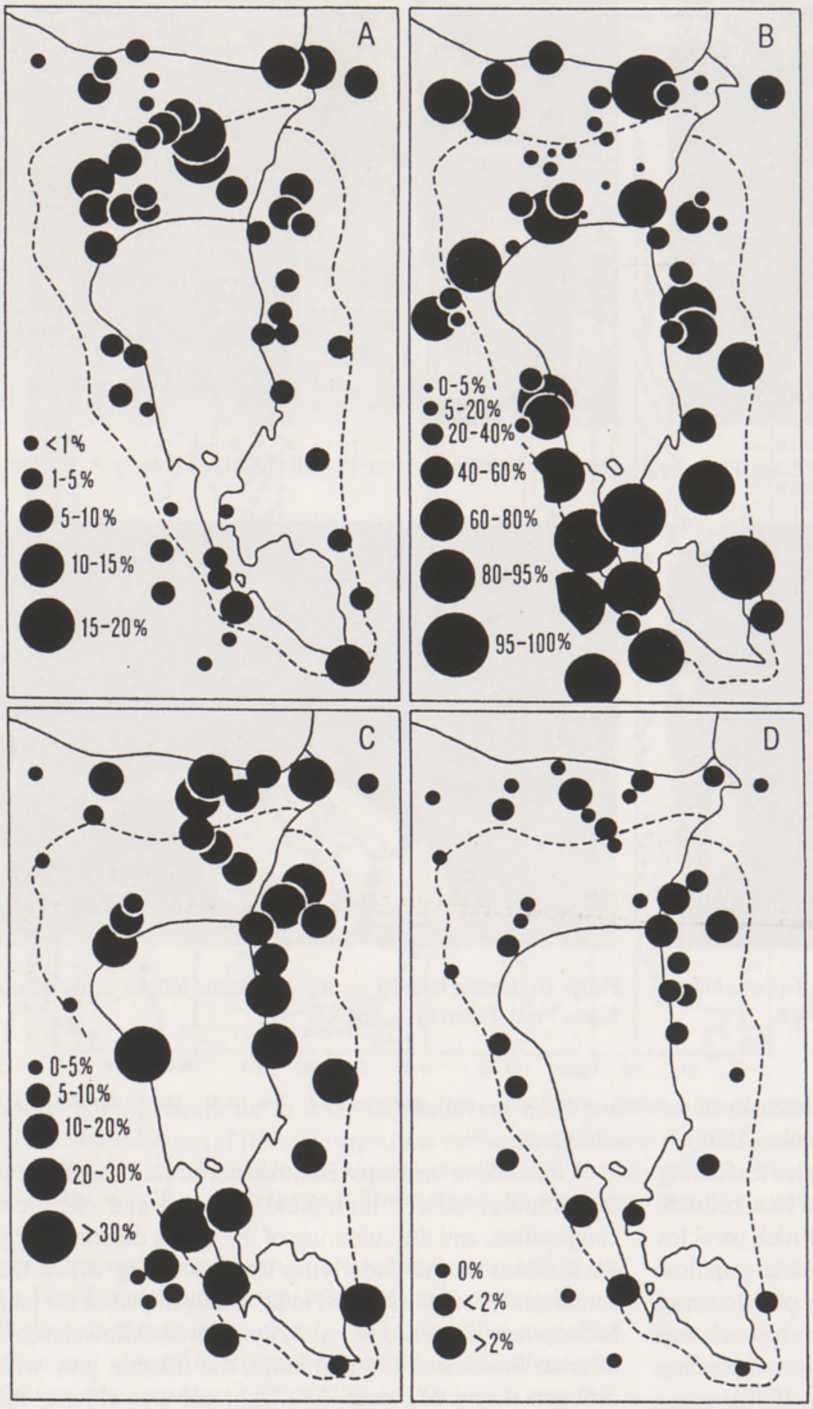

Fig. 25. Some examples on till lithology in the Peipsi Basin and surrounding areas. The content of: A – pebble fraction (10 to 100 mm); B – crystalline (magmatic and metamorphic) rocks; C – rapakivi; D – Suursaari (Hogland) quartz porphyries in the pebble fraction of crystalline rocks.

rocks, that predominate in the fore-klint till on the Vendian and Cambrian clays, silt- and sandstones, are present in smaller amounts (Fig. 25-B).

In the tills left by the last glacier on the carbonate bedrock in the northern part of the lake basin boulders make up 8–25%, pebbles and gravel 10–25% each, sand 7–20%, and silt and clay fractions 10-25% each, as an average. However, the composition of the tills overlying the Devonian sand- and siltstones in the southern portion is as follows: boulders 4–7%, pebbles 5–10%, gravel 5–20%, sand and silt 20–35% each, and clay particles 15–35%. Some typical grain-size examples of different tills are shown on Figure 26.The tills on the carbonate bedrock in the Peipsi area contain less coarse fractions than the tills elsewhere in Estonia (Raukas 1969, 1978). This can be explained with the narrow outcropping of carbonate rocks and the influence of Vendian, Cambrian and Devonian clastic rocks, mixed together in close-lying tills. The Devonian rocks in the southern part of the Peipsi Basin consist of reddish-brown and light-coloured sand- and siltstones, multicoloured clays and marls. As during the movement of the continental ice the unconsolidated, weakly cemented Devonian rocks enrich, in the first line, the sandy-silty fractions of the tills, the relative content of coarse fractions in those tills diminishes (Photo 18). In this part of the basin, the amount of carbonate matter decreases also due to ice-sheet erosion. The clasts in tills here have undergone long-term glacial transport and are, therefore, by far better rounded than their counterparts in the north. This indicates once again that during the movement of the ice, the erosion and rounding of the material caught by the glacier was much

Photo 17. Purplish-grey till, rich in clasts, on the carbonate bedrock at Tudulinna. Photo by A. Raukas.

Photo 18. Tills on the Devonian sand- and siltstones are poor in coarse fractions. Klenno. Photo by A. Raukas.

more intensive than its breaking up and crushing. Dolomite, as a much more resistant rock than limestone, plays an important role in the tills in the southern part of the lake basin. Since the crystalline rocks are more durable than the carbonate rocks, the tills in the southern part of the lake basin are enriched with the former. In the northern part of the lake basin the tills contain much more big boulders, accumulated in the coastal zone (Photo 19).

On the light-coloured Upper-Devonian dolomites in the southernmost part of the lake basin the local carbonate material prevails again in tills, and the till is rich in coarse fractions (Fig. 26-*4*).

The distance the clasts have covered with the glacier depends greatly upon the resistance of the source rocks (Raukas 1978,1995). The resistant crystalline rocks may have passed several hundred kilometres, fine-grained limestones and dolomites more than one hundred kilometres, and the weakly-cemented sand- and siltstones only some 10–15 kilometres. The sand- and siltstone clasts are rather rare in tills; however, there are also some exceptions. For example, in the reddish-brown till at Kallaste the proportion of Devonian sandstone clasts reaches 20%. In the northern part of the lake basin, *e.g.* at Tudulinna, the clasts of Cambrian sandstones account for 10–12% and soft shales for 5%.

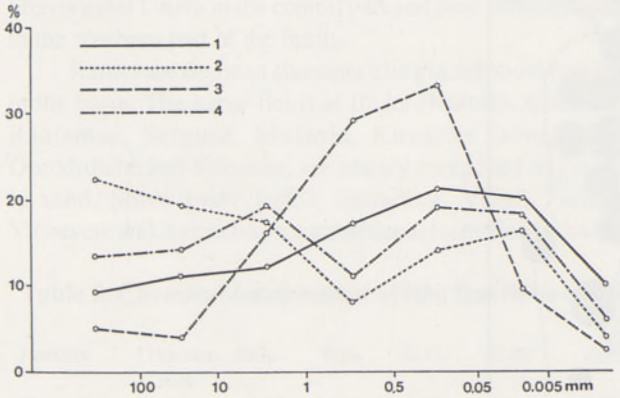

Fig. 26. Granulometric composition of main types of tills of the last glaciation in the Peipsi Basin: 1 – purplish-grey till at Tudulinna (on the outcrops of carbonate rocks); 2 – grey till near Kuremäe (on the outcrops of carbonate rocks); 3 – reddish-brown till at Mustvee (on the outcrops of Middle-Devonian sand- and siltstones); 4 – grey till near Izborsk (on the outcrops of Upper-Devonian dolomites and marls).

Among limestones and dolomites the most resistant types prevail. For instance, in the till of the last glaciation at Põlva, west of L. Pihkva, about 70% of all carbonate clasts are from the Raikküla, Juuru and Porkuni stages, some 12% from the Saunja Formation of the Nabala Stage and 5–8% from the Rakvere Stage. At the same time, the closer-lying Pirgu and Oandu stages have yielded only 5% of the clasts as have the Volkhov and Kunda stages, outcropping in the Baltic Klint.

The crystalline boulders are represented by a great variety of acid to ultramafic rocks, the source areas of which are eastern Finland and western Karelia. Due to the great Vyborg Rapakivi Massif, rapakivi is most widespread forming up to 42% of all crystalline rocks (Fig. 25-C). As to the pebbles, 78.2% are granites, 0.9% syenites and helsinkites, 2.1% mafic and ultramafic intrusive rocks, 2.0% acid effusives (mainly Suursaari quartz-porphyres, Fig. 25-D), 2.6% mafic effusives (mainly uralite porphyrites), 11.9% of metamorphic rocks, 2,1% quartzites and Jotnian quartzite-like sandstones and 0.2% other types of rocks.

The local bedrock has also exerted a remarkable influence on the mineral composition of tills. For example, outcrop distribution can easily be identified on Ordovician and Silurian carbonate rocks and Devonian sand- and siltstones. However, even smaller differences can be traced with respect to the differing composition of minerals, as already mentioned when speaking about the tills of different colour near the Devonian outcrops at the Narva River. As a

Photo 19. Big ice-pushed erratic boulders at Hitova on the eastern shore of L. Peipsi. Photo by A. Miidel.

rule, the content of carbonate minerals decreases abruptly and the content of quartz increases southwards from the outcrops of carbonate rocks (Fig. 27-A, B). Among the heavy mineral fraction, the content of amphiboles (Fig. 27-C), pyroxenes and other minerals typical of Precambrian rocks in Finland and Karelia, gradually decreases in a southerly and southeasterly direction. Correspondingly, the quantity of weathering-resistant minerals typical of the underlying Palaeozoic rocks, such as zircon (Fig. 27-D), garnet (Fig. 27-E), tourmaline, rutile, *etc.* increases. Altogether, more than 50 different minerals and mineral groups have been identified in the coarse silt (0.05–0.1 mm) and fine sand (0.1–0.25 mm) fractions. The dominating minerals are quartz, feldspars and carbonates which make up more than 95% of all the minerals recognized. The content of micas rarely exceeds 3%. The proportion of the heavy subfraction (more than 2.89 Mg/m^3) is usually small (about 1%), being a bit higher in the northern part of the basin. On the carbonate bedrock, the content of calcite and dolomite is mainly 20–60%, on the sand- and siltstones it is less than 10%; carbonates are often entirely lacking. It is of interest that in the northern part of the basin calcite is prevailing over dolomite, in the central and southern parts, *vice versa*, dolomite is more abundant. It may be explained with the influence of the bedrock in which dolomite is a more common rock type, and with the higher resistance of the dolomite during the transport. Characteristic of the Peipsi Basin is the high content of ore minerals (mainly ilmenite and magnetite), ranging from 20 to 50% in the fine sand heavy subfraction. On the Devonian outcrop, under the influence of the bedrock, the content of garnet decreases

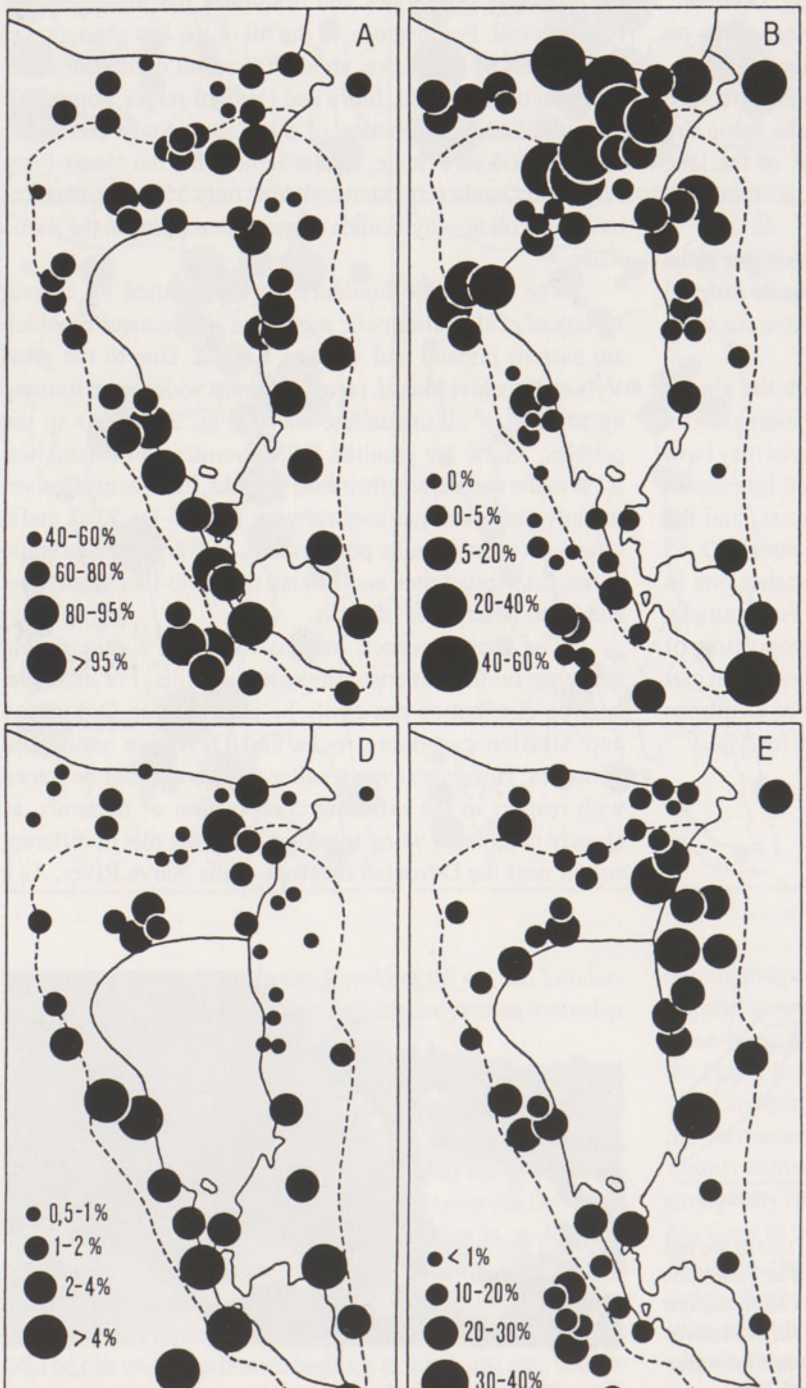

Fig. 27. The content of different minerals in the fine sand (0.1–0.25 mm) fraction of tills of the last glaciation in the Peipsi Basin. A – quartz, and B – carbonates (calcite and dolomite) in the light (<2.89 Mg/m^3) subfraction; C – amphibole, D – zircon, and E – garnet in the heavy (>2.89 Mg/m^3) subfraction.

slightly and that of zircon increases southwards (Raukas 1978).

A clear trend in the rounding of mineral grains is apparent. In the five-point-system the roundness of quartz grains in the northern part of the basin is 1.7–2.1, in the southern part it is 1.9–2.3. This is due to both the erosion during the transport and the influence of the Devonian bedrock, in which the mineral grains are originally better rounded.

The optical, thermal, X-ray and chemical analyses have shown that the main components of the clay fractions (<1 μm) of tills are hydromicas (illite) which make up 60–75%. On the basis of the X-ray analyses by K. Utsal from Tartu University, it is possible to identify in their composition some mixed-layered structures in which latticed elements of hydromicas interchange regularly with single swelling layers. Hydromicas of trioctahedral type and chlorite are also present but in small amounts. Well-crystallized kaolinite (15–25%) is a typical addition in all investigated sites. The results of chemical analyses show that Fe-rich hydromicas prevail. Montmorillonite was absent or its content was less than 5%. Calcite, dolomite, quartz, feldspars, amorphous silica and ferric oxides occurred as mineral impurities, but their total amount did not exceed 10%.

The chemical composition of fine fractions is rather variegated (Table 5). In the northern part of the basin (Kuremäe, Raudja, Agusalu) the tills contain more carbonates and less K_2O, Fe_2O_3, P_2O_5, SiO_2, TiO_2 and Al_2O_3 than the tills on the Devonian outcrop (Lämmijärv, Kallaste, Räpina, Värska). The content of free carbonates in the fraction less than 0.1 mm is between 0–32%; in most cases it is up to 20% in the northern and up to 5% in the southern part of the basin.

Lithology of glaciofluvial deposits. Both englacial (radial eskers, fluviokames) and periglacial (glaciofluvial deltas, outwash deltas, marginal eskers) genetical types of glaciofluvial deposits are represented in the Peipsi Basin and its surroundings. Sand, gravel and pebble-cobble material of glaciofluvial origin occurs also in the hilly topography. It has been described in the environs of Illuka–Raudi, Ohakvere, Mäetaguse, Assikvere in the northern part of the depression, near Dobruchi–Hvostrogo, Sidorovshchina, Zaborov'e, Shavora and Usova in the central part and near Novo-Izborsk in the southern part of the basin.

Kames are the most characteristic glaciofluvial features in the basin. The kame fields at Illuka (Kurtna), Koidmaa–Räätsmaa, Selguse, Mustoja, Knyazya Gora, Kaiu, Gorodishche and Kamarna, are mainly composed of gravelly sand. Some kame fields, including, Välgi, Töljase, Vassevere and Avinurme, are transitional formations between kames and hilly morainic topography. They comprise frequently hills with till cover, or hills which consist of weakly washed gravel and pebble-cobble material. In the small kame fields at Runga, Pilli and Hommiku, glaciofluvial kames are covered with varved clays or nonstratified glaciolacustrine deposits.

Marginal eskers are rare in the lake basin. The biggest eskers, located at Torkhovo–Zamogil'e, consist of sandy-pebble gravel which is prevailed by crystalline rocks (70%). Both carbonate (2.2) and crystalline (2.0–2.5) clasts are well-rounded. The marginal eskers at Buyokski, Kõnnu and Torma are also well preserved.

Radial eskers, which are more common in the lake basin, form several esker systems: Jaastrova–Beres'e, Rostitsa–Kolpino and Põdrassaare. Of the single comet-like eskers occurring in some places, the Slantsy esker is the most prominent. It is 1500 m long, up to 300 m wide and consists of rougher poorly outwashed material (pebbles-cobbles with sand and gravel, Photo 20) in the proximal part. Among the clasts, carbonate material is prevailing (in the pebble fraction 70–85%).

The most characteristic glaciofluvial deltas are Piirissaar and Talabanets. The Piirissaar delta consists mainly of sand and the Talabanets delta mostly from gravelly material.

Glaciofluvial deposits have a highly variable granulometric composition and structure. Their lithological and mineral composition is also diverse and closely related to the composition of the adjacent till and bedrock (Raukas 1978).

During 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 the gravel-and-pebble fractions of glaciofluvial deposits is by 10–15% higher than in tills. The content of carbonaceous rocks is accordingly lower. The content of metamorphic (predominantly gneisses) and coarse-grained magmatic rocks (mostly rapakivi and coarse-grained granites) is 5–10% lower than in initial tills (Raukas 1978).

Lithology of glaciolacustrine deposits. Glaciolacustrine deposits are also divided into englacial and periglacial genetical types. Englacial deposits form plateau-like limnoglacial kames (limnokames) near Iisaku and Illuka (Iisaku, Pootsiku, Varesemetsa, *etc.*), Saki - Parapalu and Kallijärve–Leegu. Kames consist of some 10-m-thick fine silty material with horizontal lamination. Frequently, varved-like rhythmical lamination exists, where sandy-silty material with a thickness of some centimetres regularly alternates with thinner silty-pelitic layers. The above-

Table 5. Chemical composition of fine fractions of tills

Locality	Fraction, mm	Compounds												
		SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	CaO	MgO	K_2O	Na_2O	P_2O_5	L.O.I	H_2O^-	CO_2
Kuremäe	< 0.1	40.13	0.45	9.84	2.04	0.99	19.66	3.69	2.24	1.07	0.14	19.92	0.87	17.88
Raudja	< 0.1	44.67	0.25	8.22	0.95	0.96	18.61	3.84	1.78	0.84	0.13	19.02	0.48	17.38
Agusalu	< 0.1	47.13	0.53	11.50	1.91	1.68	11.05	5.77	3.41	0.64	0.19	16.24	0.78	14.12
Lämmijärve	< 0.1	68.45	0.69	14.49	3.14	1.13	0.70	1.58	3.49	1.20	0.22	4.15	0.86	1.37
Kallaste	< 0.1	53.96	0.95	19.69	4.31	2.72	1.93	2.90	4.22	0.71	0.28	8.54	2.13	3.25
Räpina	< 0.01	53.72	1.00	21.15	6.90	1.11	0.83	1.96	4.27	0.53	0.26	9.07	2.43	1.38
Värska	< 0.01	55.58	1.00	18.82	6.53	1.13	0.95	1.97	4.46	0.65	0.33	7.84	1.81	1.66
Tudulinna	< 0.002	48.12	0.80	15.10		12.30	7.40	3.75	3.80	3.02		4.58	2.43	3.86

Analyst V. Säga

Photo 20. Poorly outwashed coarse-grained glaciofluvial deposits in the Slantsy comet-like esker. Photo by A. Miidel.

Photo 21. Glaciotectonically disturbed glaciolacustrine deposits in the Lis'e end-moraine. Photo by A. Raukas.

mentioned deposits are rich in micas (up to 15%) and contain quartz (60–90%) and feldspars (9–22%) in different amounts. The content of carbonates is low; in most cases it is less than 5%.

Varved clays are widely distributed in the Alutaguse and Narva lowlands with outcrops on the banks of the Narva, Pungerja and Alajõe rivers, in the mouth of the Emajõgi River and in several places near L. Pihkva. The major part of the varved clays was formed during the recession of the last glacier in ice-dammed lakes that existed at different times in the Peipsi Basin (Raukas & Rähni 1969). Occasionally, *e.g.* at Lis'e (Piipenberg 1935), intermorainic varved clay deposits have been found. These clays accumulated during the intervals between the stadial or oscillatory intrusions of the ice margin (Photo 21). Varved clays older than the last glaciation can be found only in ancient valleys.

E.Pirrus (1968) has pointed out the predominance of illite (dioctahedral hydromicas) and mixed-layered formations (55–75%) in clay fractions (<1μm) in the varved clays of the area. Also kaolinite (15–25%), various terrigenous minerals (up to 10%), amorphous silica (0.5–1.5%) and several hydrous ferric oxides (1.5–3 %) occur. After Saarse and Utsal (1974), chlorites (5–10%) are distributed in southern Estonia, including the Võru–Piusa Lowland. In the mineral composition of the sand and silt fractions quartz, feldspars, carbonates and micas (90–95%) prevail, their ratio varies considerably from fraction to fraction. Within the limits of one and the same section, the mineral composition is rather stable. 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 tills into glaciolacustrine clay (Pirrus & Raukas 1963). The ratio of single minerals and groups of minerals in one or another fraction from different parts of the profile is rather stable.

2.4.2.2. HOLOCENE DEPOSITS

Material. During the course of several decades, numerous pollen analytical studies have been carried out in the Peipsi Basin. The first studies of the kind were undertaken by Thomson (1929). Several overview-diagrams were compiled in the 1960s for the purposes of geological mapping and mineral prospecting of the region (Zirna & Pirrus 1961, Kajak *et al.* 1961, Väärsi *et al.* 1964, Tassa *et al.* 1967). Complex studies aimed at elucidating the geology of the lake were carried out at the end of the 60s and in the beginning of the 70s. During these investigations, more than ten peat cores were studied palynologically by R. Pirrus and the vegetational history of the region was described (Miidel *et al.* 1972).

Several sequences were analysed in the region of the Emajõgi River mouth. Sarv and Ilves (1975) studied the reference site at Saviku (Fig. 24). As a result, the first series of ^{14}C dates for the region was obtained. The natural conditions in the vicinity of the Akali neolithic settlement were studied in 1977–78 and two pollen diagrams were compiled by R. Pirrus (Moora *et al.* 1988).

With the focus on the features of the regional stratigraphy, additional cores of several middle-sized mires in the area of the basin were analysed (Kimmel 1994). Recently, the deposits of the Obdekh (Optjok) River floodplain mire were studied in order to clarify the development of the southern part of L. Peipsi in the Holocene (Miidel *et al.* 1995).

The number of pollen diagrams, describing organic sediments in the lake basin and constructed for different purposes, exceeds 30. For this overview, more complete and modern pollen diagrams, partly dated by ^{14}C were selected (Table 6; Fig. 28).

The description of the selected sites. The palynologically studied sequences of peat and gyttja provide material for the stratigraphical subdivision and correlations. The cores from middle-sized mires were mostly studied, because from a pollen-analytical point of view they reflect the vegetational development on a regional scale. Half of the diagrams cover the whole Holocene, in the rest of the cores the Holocene

Table 6. Palynologically investigated sites in the area of the Peipsi Basin which served as a main source for our survey. The chronozones and PAZ-s are given according to the authors' division.

No.	Site	Publication or manuscript	^{14}C dates	Chronozones and PAZ-s	Sediment type
1.	Puhatu	Miidel *et al.* 1972	-	PB-SA2	peat
2.	Muraka	Miidel *et al.* 1972	-	PB-SA2	peat
3.	Chistyj Mokh	Miidel *et al.* 1972	-	PB-SA2	peat
4.	Rannapungerja	Miidel *et al.* 1972	-	AT2-SA2	peat/gyttja
5.	Vasknarva	Miidel *et al.* 1972	-	AT2-SA2	peat/gyttja
6.	Kunest´	Miidel *et al.* 1972	-	AT1-SA2	peat
7.	Gorodishche	Miidel *et al.* 1972	-	AT2-SA2	peat
8.	Samolva	Miidel *et al.* 1972	-	SB1-SA2	peat
9.	Rov'ya	Miidel *et al.* 1972	-	AT2-SA2	peat/gyttja
10.	Alasoo	Miidel *et al.* 1972	4	DR3-SA2	peat/gyttja
11.	Saviku	Sarv & Ilves 1975,			
		Saarse *et al.* 1995	13	DR3-SA2	
			11	PAZ	peat/gyttja
12.	Akali	Moora *et al.*1988	4	AT-SA3	peat
13.	Võhmajärve	Kimmel 1990	14	6 PAZ	peat
14.	Imatu	Kimmel 1990	3	9 PAZ	peat/gyttja
15.	Mannjärv	Kimmel 1990	4	9 PAZ	peat/gyttja
16.	Kalsa	Kimmel 1994	13	10 PAZ	peat
17.	Toolamaa	Kimmel 1994	14	10 PAZ	peat
18.	Obdekh (Optjok)	Miidel *et al.* 1995	10	8 PAZ	peat

pollen sequence is not complete (the cores from the areas were influenced by the water-level rise in L. Peipsi). During the years, analysts have used different zone systems. The earlier schemes served as a combination of climato-, bio- and chronostratigraphy. The biostratigraphical units (pollen zones) were directly used to estimate the age of sediments. The terms Pre-Boreal, Boreal, *etc.* were used because there were not enough radiocarbon dates. In recent studies, the pollen assemblage zones (PAZ) have been used, which are correlated with the chronological scale by means of radiocarbon dates.

Muraka. The Muraka Mire (12,793 ha), a complicated system of bog massifs with mineral islands, is situated 15 km north of L. Peipsi. In the core, taken from the middle section of the eastern part of the mire, the basal fine-grained sand is covered by different kinds of peat with the total thickness of 6.20 m. The pollen diagram (Miidel *et al.* 1972) is zoned on the basis of the pollen (no radiocarbon dates), and it is concluded that the peat has accumulated here since the Pre-Boreal.

Puhatu. Puhatu (57,079 ha), the biggest mire system in Estonia, consists of many different bog complexes. It is situated 12 km north of L. Peipsi. The sequence was studied

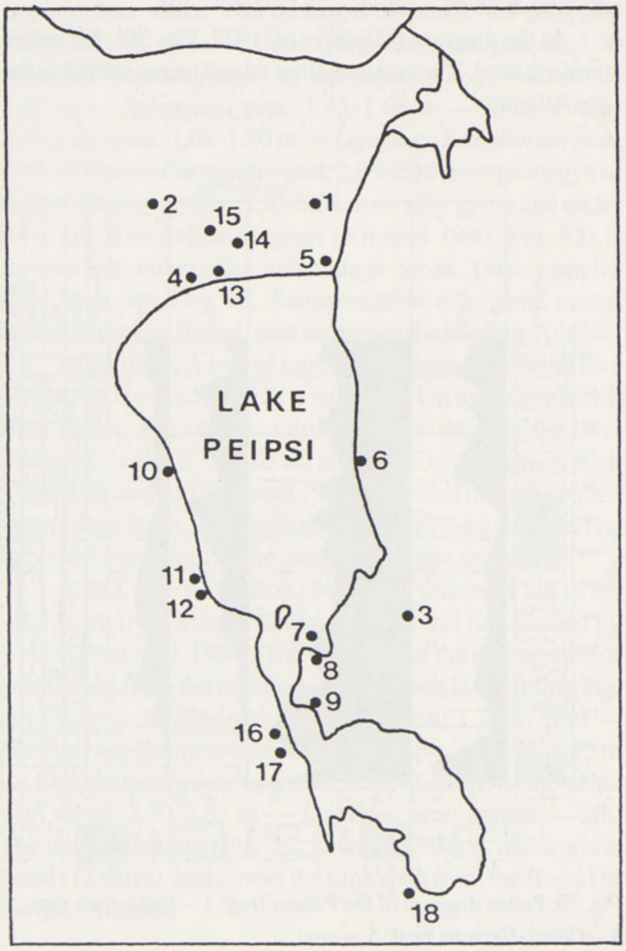

Fig. 28. Location of biostratigraphically studied sites of the Peipsi Basin mentioned in the text: 1. Puhatu; 2 – Muraka; 3 – Chistyi Mokh; 4 – Rannapungerja; 5 – Vasknarva; 6 – Kunest'; 7 – Gorodishche; 8 – Samolva; 9 – Rov'ya; 10 – Alasoo; 11 – Saviku; 12 – Akali; 13 – Võhmajärve; 14 – Imatu; 15 – Mannjärv; 16 – Kalsa; 17 – Toolamaa; 18 – Obdekh (Optjok).

in the central part of the system, west of L. Puhatu. Hear the basal sand (7.35–7.50 m) is covered by *Carex–Hypnum* peat (6.60–7.30 m), *Phragmites* peat (5.80–6.60 m), *Lignetum-Phragmites* peat (3.75–5.80 m), *Carex* peat (2.70–3.75 m) and *Sphagnum* peat (0–2.70 m). The pollen diagram (Miidel *et al.* 1972, Fig. 29) is divided into pollen zones. The peat deposits display a sediment record since the Pre-Boreal.

Chistyj Mokh. The mire is located in the eastern part of the depression, 12 km to the east from the mouth of the Zhelcha River. The lithology is as follows: 0–1.20m — *Eriophorum-Sphagnum* peat; 1.20–1.90 m — *Sphagnum* peat; 1.90–2.35 m — *Sphagnum–Eriophorum* peat; 2.35–2.70 m — *Pinus-Eriophorum* peat; 2.70–3.0 m — *Lignetum* peat; 3.0–3.10 m — sand. The diagram is subdivided into pollen zones. The fen peat started to accumulate in the Pre-Boreal.

Rannapungerja. The transition mire, an overgrown oxbow of the Rannapungerja River with a diameter of 600 m is located 2 km north-east of the mouth of the river. The main part of the sequence is formed by the fen peat of 3 m thickness. The lithology is as follows: 0–0.75 m — *Carex-Phragmites* peat; 0.75–1.25 m — *Carex* peat; 1.25–2.90 m — *Lignetum* peat; 2.90–3.0 m — peaty gyttja; 3.0–3.25 m — gyttja with plant remains; 3.25–3.35 m — grey sandy silt; 3.35–3.50 m — sand.

According to the zonation (Miidel *et al.* 1972), the accumulation of sediments started in the Late Atlantic.

Vasknarva. The transition mire formed in an oxbow of the Narva River is situated 3 km west of Vasknarva, 1.3 km to the north from L. Peipsi, 3 km east of the river. The lithostratigraphy is as follows: 0–0.20 m — *Carex–Sphagnum* peat; 0.20–2.10 m — *Phragmites* peat; 2.10–4.0 m — gyttja; 4.0–4.20 m — silty sand with gyttja; 4.20–4.50 m — sandy silt; 4.50–6.75 m — sand; 6.75–7.0 m — till.

In the diagram (Miidel *et al.* 1972, Fig. 30), the pollen zones are used. The accumulation of sediments started in the Late Atlantic.

Kunest´. The studied oxbow mire is located on L. Peipsi's east coast, 300 m to the east from the mouth of the Kunest´ River. The lithology is as follows: 0–0.60 m — *Carex* peat with silt; 0.60–1.40 m — *Carex–Phragmites* peat with silt; 1.40–1.55 m — *Phragmites–Hypnum* peat with silt; 1.55–2.50 m — *Lignetum–Phragmites* peat with silt; 2.50–4.90 m — *Phragmites* peat with silt; 4.90–5.30 m — muddy peat with shell fragments; 5.30–5.50 m — sand with organic matter and shell fragments; 5.50–6.60 m — sand. The diagram (Miidel *et al.* 1972) represents the stratigraphical sequence from the Early Atlantic to the Sub-Atlantic.

Gorodishche. Gorodishche Island (2.5 × 0.3 km) is situated in the southern part of L. Peipsi, in the bay, into which the Samolva and Zhelcha rivers flow. The island is covered by *Carex–Phragmites* fen. The sequence is from the central part of the island. The lithology is as follows: 0–0.55 m — *Phragmites* peat; 0.55–0.80 m — sand; 0.80–2.85 m — *Phragmites* peat; 2.85–3.30 m — *Lignetum-Phragmites* peat; 3.30–3.35 m — black silty sand with organic matter; 3.35–3.50 m — sand with organic matter. The diagram (Miidel *et al.* 1972) is divided into pollen zones. The sequence starts in the Late Atlantic.

Samolva. The site is 1 km to the south from the mouth of the Samolva River, in the southeastern part of the Peipsi Basin. The sequence is as follows: 0–0.20 m — *Carex peat*; 0.20–2.20 m — *Carex-Phragmites* peat; 2.20–2.50 m — *Lignetum* peat; 2.50–2.60 m — silt with organic matter; 2.60–2.75 m — sand. The pollen diagram (Miidel *et al.* 1972) is divided into pollen zones. The upper part of the Holocene (SB1 - SA) is represented.

Rov'ya. The site is situated on the left bank of the Rov'ya River, on the east coast of L. Lämmijärv. The lithology is as follows: 0–0.50 m — *Carex* peat; 0.50–1.25 m — *Phragmites* peat; 1.25–2.15 m — *Lignetum-Phragmites* peat; 2.15–3.50 m — *Lignetum* peat; 3.50–4.25 m — gyttja; 4.25–

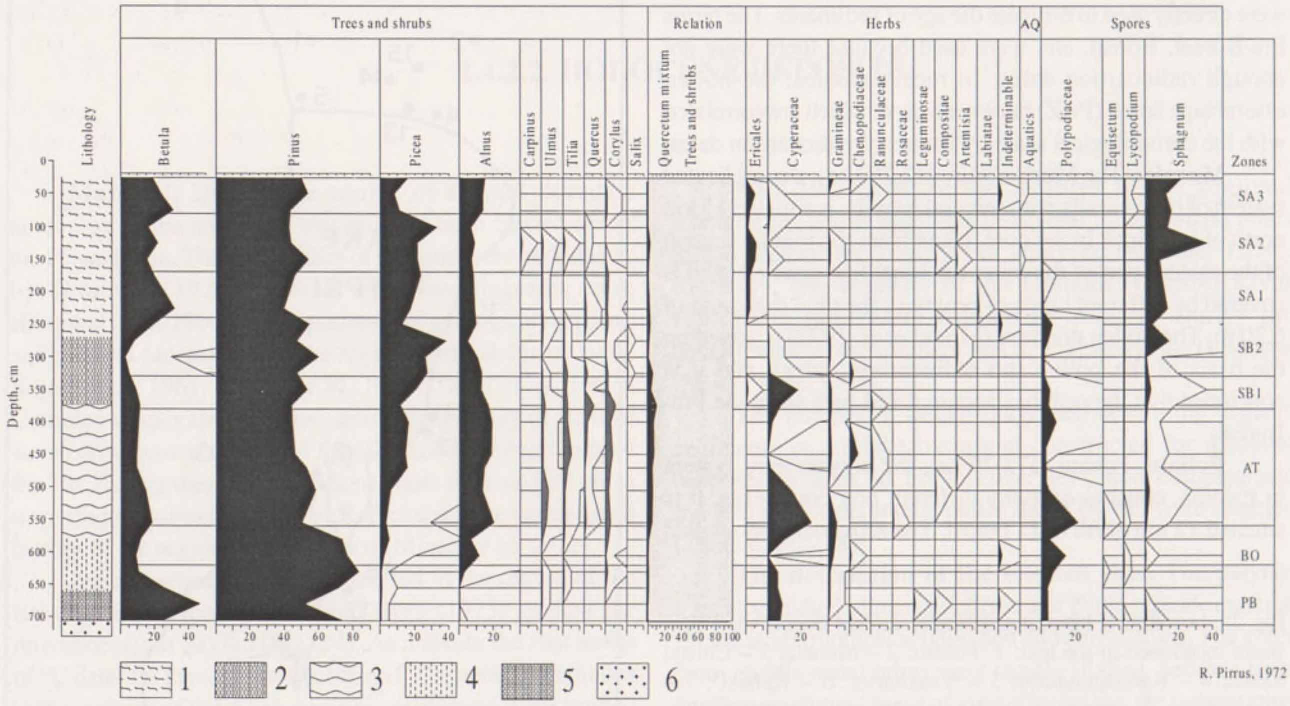

Fig. 29. Pollen diagram of the Puhatu Bog: 1 – *Sphagnum* peat; 2– *Carex* peat; 3 – *Lignetum-Pharagmites* peat; 4 – *Pharagmites* peat; 4 – *Carex-Hypnum* peat; 5 – sand.

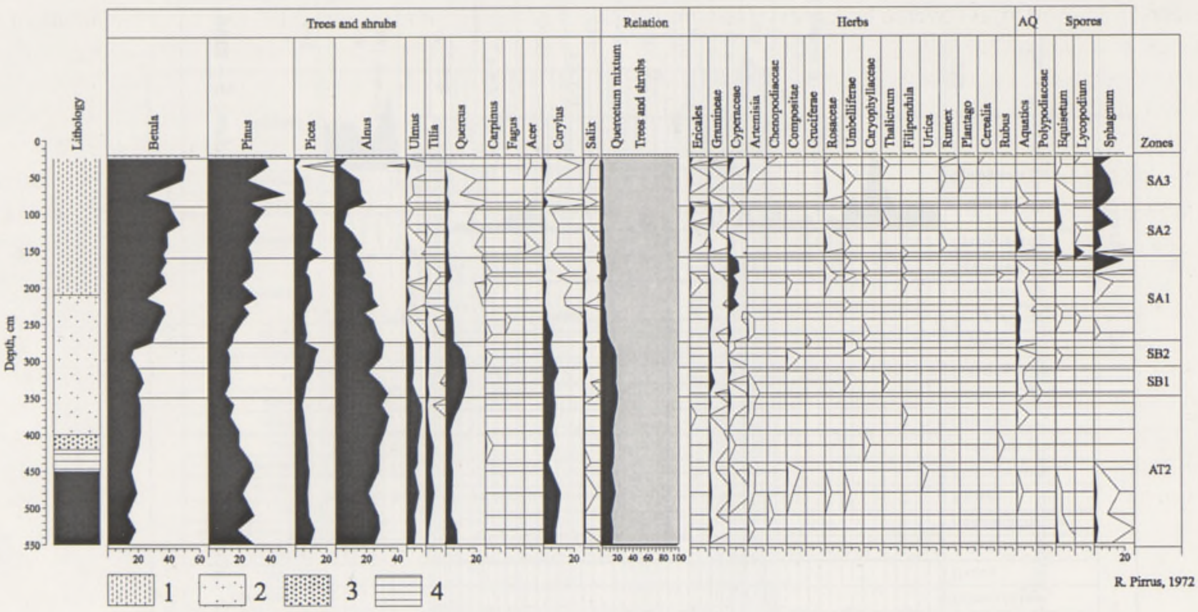

Fig. 30. Pollen diagram of the Vasknarva Bog (oxbow deposits): *1 – Pharagmites* peat; 2 – gyttja; 3 – sand; 4 – silt.

4.95 m — sand with organic matter.The diagram is divided into pollen zones (Miidel *et al.* 1972). The sand and the lower part of the gyttja accumulated in the Late Atlantic.

Alasoo. Alasoo Mire (498 ha) is situated 1.5 km south of Kallaste Town, in the region of the mouth of the Alatskivi Valley. The pollen diagram (Miidel *et al.* 1972) is subdivided into pollen zones. The bottommost silt (4.40–4.70 m) of DR3 is overlain by peat and gyttja (3.50–4.40 m) of PB and BO. *Phragmites peat* (1.05–3.50 m) has formed since AT and is covered by the sand of L. Peipsi's beach formations which accumulated in SA2.

Akali. Two sequences have been studied from the paludified mouth region of the Emajõgi River (L. Peipsi's west coast), on the southwestern bank of the small Akali River that links the Emajõgi with the Kallijõgi. The pollen zones are used for the zonation (Moora *et al.* 1988). *Carex* and *Carex-Phragmites* peat has formed here on the sand since the Late Atlantic. Two radiocarbon dates from both of the sequences are available.

Saviku. The bio- and chronostratigraphically studied sequence of peat deposits is situated on the left bank of the Emajõgi River. The pollen diagram was first published by Sarv and Ilves (1975) and re-interpreted later (Saarse *et al.* 1995). The 4.9-m-thick peat deposits are underlain by gyttja, 3.15 m in thickness. The transition between peat and gyttja is at an elevation of 25.1 m and is dated to 5800 BP. The diagram is divided into eleven local pollen assemblage zones. Thirteen radiocarbon dates are available. The diagram covers the time span from the Late Dryas to the Sub-Atlantic.

Võhmajärve. The Võhmajärve Bog (872 ha) is situated 4 km to the north from L. Peipsi, 2 km south of Sälliku. The core was drilled in the central part of the bog. The sediment stratigraphy is as follows: 0–2.95 m — *Sphagnum* peat; 2.95–5.50 m — *Eriophorum-Sphagnum* peat; 5.50–5.58 m — transition peat (*Eriophorum-Spagnum* peat with *Lycopodium*); 5.58–5.70 m — fen peat (*Eriophorum* peat with *Lycopodium*); 5.70 m — sand. The diagram (Kimmel 1990, Fig. 31) is divided into six local pollen assemblage zones. Fourteen radiocarbon dates have been obtained. The peat started to form in the Early Atlantic.

Imatu. Lake Imatu (27.8 ha, 43.8 m a.s.l.) is situated 8 km to the east from Iisaku and is surrounded by the Järvesoo transitional mire. The samples were collected near the southern lake shore. The thickness of peat at the sampling site is 2.05 m, while the underlying lake sediments are 3.50 m thick. The stratigraphy of the deposits is as follows: 0–1.45 m — *Sphagnum* peat; 1.45–1.68 m — *Eriophorum-Sphagnum* peat; 1.68–1.80 m — *Lignetum-Eriophorum* peat; 1.94–2.05 m — *Phragmites* peat; 2.05–2.10 m — peaty gyttja; 2.10–5.50 m — gyttja; 5.50–5.54 m — silty gyttja and under it — silt. The pollen diagram (Kimmel 1990, Fig. 32) is divided into nine pollen assemblage zones. Three samples have been dated by ^{14}C. Sedimentation of organic matter started in the Pre-Boreal, peat formation the late Sub-Atlantic.

Mannjärv. A belt of raised bog surrounds a small lake (4.4 ha, 59.0 m a.s.l), which is situated 6 km to the north-east from Iisaku. The core was drilled on the shore of the lake. The gyttja layer (3.75–6.65 m) is covered by *Sphagnum* peat. The pollen diagram (Kimmel 1990) is divided into nine pollen assemblage zones, four radiocarbon dates are available.The sequence characterizes the whole Holocene sequence.

Kalsa. The Kalsa Bog (245 ha) is situated 4 km to the south-west from Meeksi Settlement and was first studied by Veber (Truu *et al.* 1964). The lithology of the new section of peat drilled from the middle part of the bog is the following: 0–1.30 m — *Eriophorum–Sphagnum* peat; 1.30–1.75 m — *Eriophorum-Sphagnum* peat with wood pieces; 1.75–1.85 m — *Eriophorum-Sphagnum* peat; 1.85–2.70 m — the same with wood; 2.70–2.85 m — *Lignetum* peat; bottom — silt. The pollen diagram (Fig. 33, Kimmel 1994) is radiocarbon dated (13 dates), and covers the time span from the Boreal to the Sub-Atlantic.

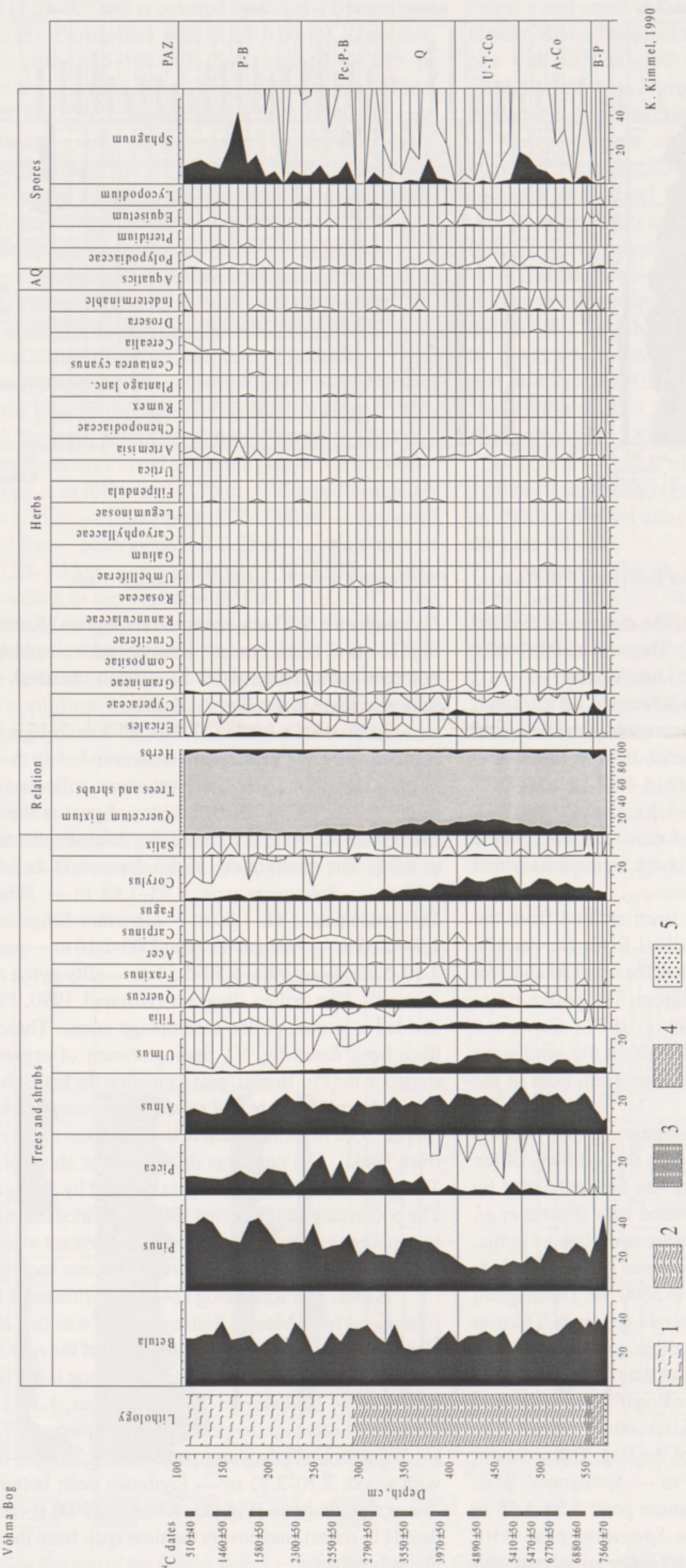

Fig. 31. Pollen diagram of the Võhmajärve Bog: 1 – *Sphagnum* peat; 2 – *Eriphorum-Sphagnum* peat; 3 – *Eriphorum-Sphagnum* peat; 4 – *Eriphorum-Sphagnum* peat with *Lycopedum*; 5 – sand.

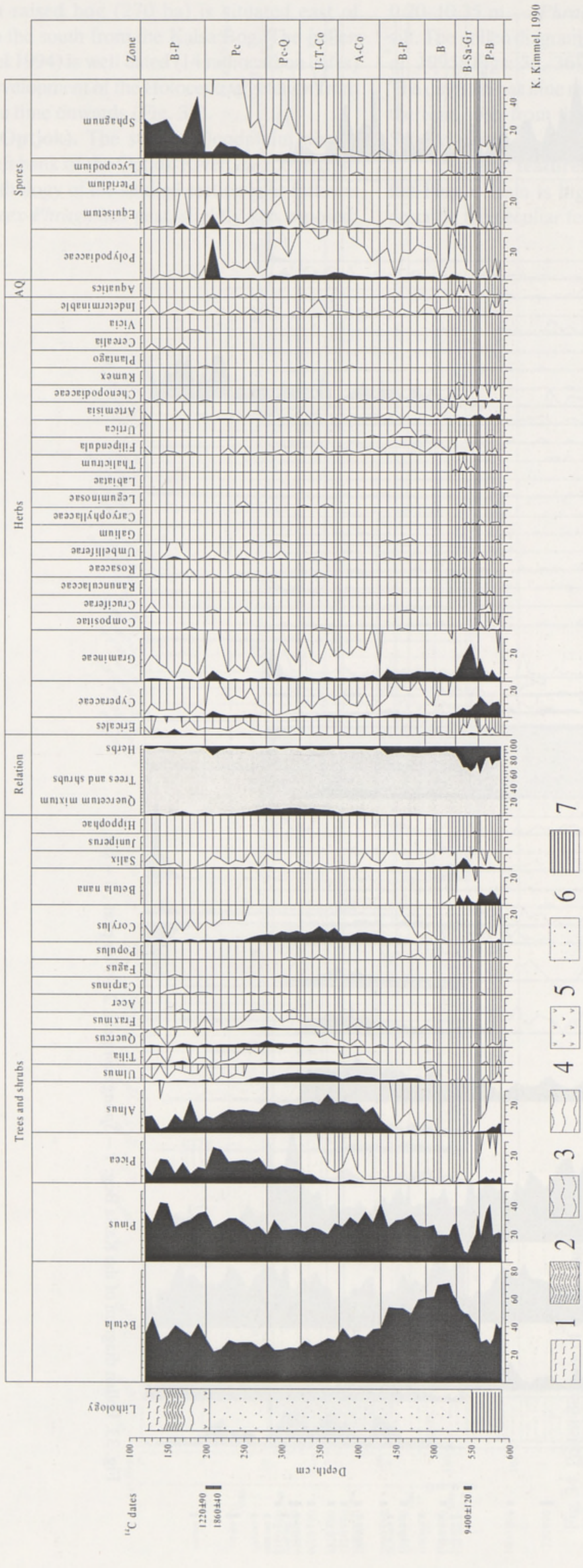

Fig. 32. Pollen diagram of the Imatu: 1 – *Sphagnum* peat; 2 – *Eriophorum-Sphagnum* peat; 3 – *Lignetum-Eriphorum* peat; 4 – *Lignetum-Phragmites* peat; 5 – *Pharagmites* peat: 6 – gytja; 7 – silt.

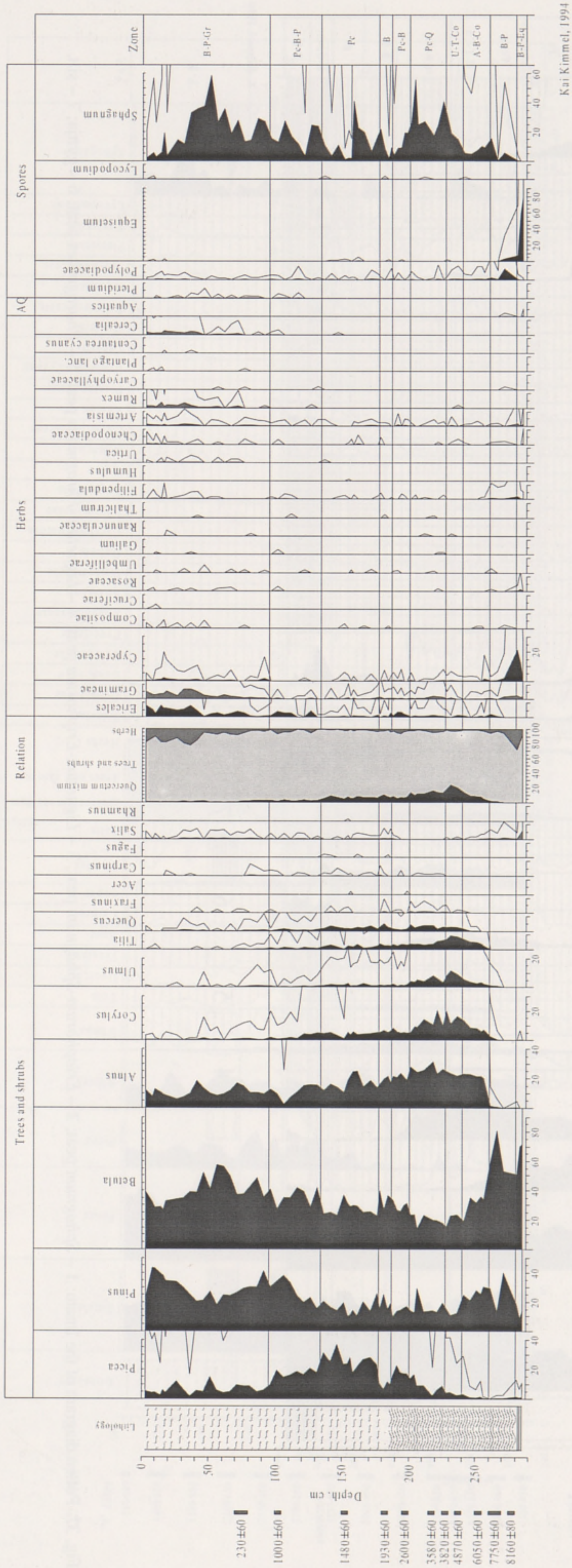

Fig. 33. Pollen diagram of the Kalsa Bog: 1 – *Sphagnum* peat; 2 – *Fen* peat ; 3 – silt.

Toolamaa raised bog (270 ha) is situated east of Meelva, 8 km to the south from the Kalsa Bog. The pollen diagram (Kimmel 1994) is well dated (14 radiocarbon dates) and shows the development of the Holocene vegetation from the Early Atlantic time onwards (Fig. 34).

Obdekh (Optjok). The studied floodplain mire is situated on the left bank of the Optjok River, *ca* 100 m from L. Peipsi. The lithology of the studied section is as follows: 0–0.20 m — *Carex-Phragmites* peat, slightly decomposed; 0.20–10.35 m — *Phragmites* peat; 10.35–10.50 m — grey silt. The pollen diagram was compiled by R. Pirrus (Miidel *et al.* 1995; Figs. 35, 36), eight local PAZ are distinguished. The diagram has nine radiocarbon dates. The diagram covers the time span from the beginning of the Pre-Boreal to the Sub-Atlantic.

Peculiar features of the local biostratigraphy. The flat Peipsi Basin is highly paludified. The first attempt to describe the peculiar features of the local biostratigraphy in

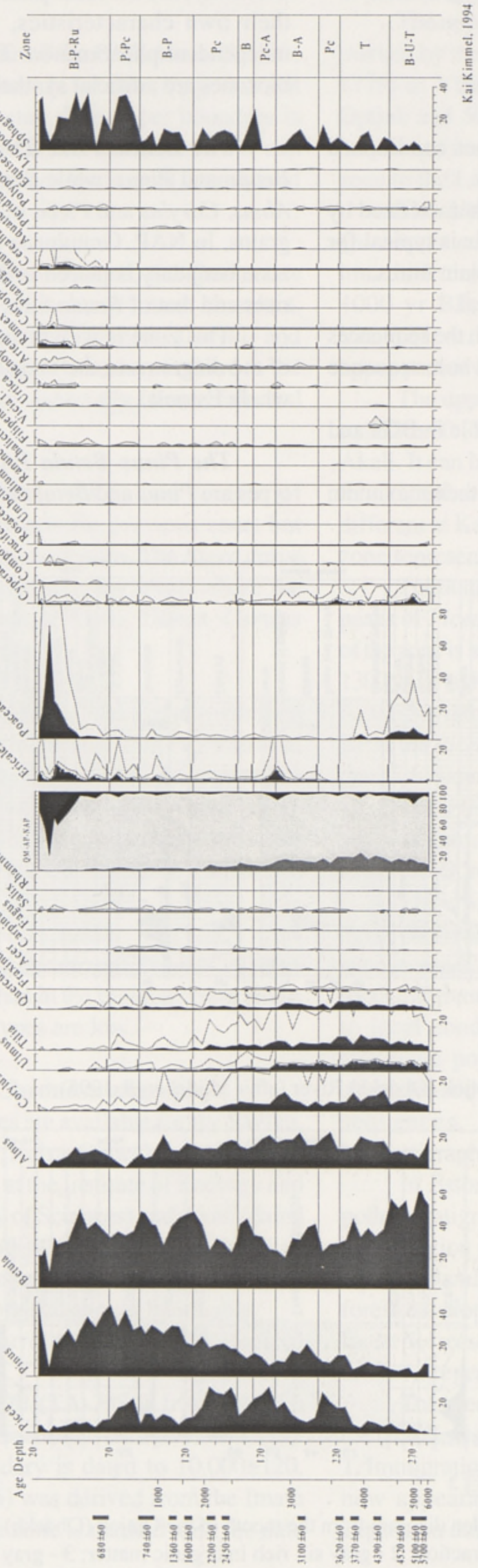

Fig. 34. Pollen diagram of the Toolamaa Bog.

more detail was made by R. Pirrus on the basis of the comparison of the pollen diagrams available at that time (Miidel *et al.* 1972). The following pollen zones were estimated; the zones were also treated as climatic periods.
The pollen zone DR3;
The pollen zone PB — *Betula–Pinus;*
The pollen zone BO1 — *Pinus–Betula;*
The pollen zone BO2 — *Pinus–Betula– Alnus;*
The pollen zone AT1 — *Alnus–Ulmus–Corylus;*
The pollen zone AT2 — *QM–Alnus–Corylus;*
The pollen zone SB1 — *Pinus–Betula–Alnus–Quercus*;
The pollen zone SB2 — *Pinus–Picea*;
The pollen zone SA1b — *Pinus–Betula;*
The pollen zone SA1a — *Pinus–Betula–Picea;*
The pollen zone SA2 — *Pinus–Betula.*

The peculiar features of the pollen zones conditioned by regional vegetation history were outlined:

1. The beginning of the Pre-Boreal is characterized by a high content of the *Salix* pollen. Later *Salix* is typical for the sequences connected to rivers and floodplain mires.

2. *Betula* is dominating in the Pre-Boreal.

3. In BO2, *Pinus* has the highest values in the sequences where its content remains high through the whole sequence (the sequences in the areas of sandy soils).

4. *Picea* appears earlier (BO1) and its role in BO2 and AT1 is higher in the eastern part of the basin.

5. *Quercetum mixtum* and *Corylus* reach their maximum in AT2 and form usually 20–25%. The value of *Quercetum mixtum* is very low (6%) in the regions of sandy soils.

6. *Quercus* maximum appears in SB1; in SB2 it is present in some sequences only.

7. The upper maximum of *Alnus,* characteristic of the Sub-Atlantic, is not found in the Peipsi Basin.

On the basis of the pollen-stratigraphical data available up to now, the succession of pollen zones describing the main tendencies in the regional vegetational development in the Peipsi Basin is as follows below. It must be stressed that each site, with its own local pollen assemblage zones all having their own characteristics, is an effectively unique and independent phenomenon. The regional pollen assemblage biozones are artificial syntheses.

The Betula PAZ. The tree pollen is represented by *Betula* and *Pinus*, while *Betula* is mostly dominant. *Ulmus, Alnus, Corylus* and *Picea* may be present with single pollen grains. In NAP, Gramineae and Cyperaceae dominate. The zone boundary is placed at the level, where the *Pinus* curve rises and that of *Betula* falls.

This zone is well distinguishable and distinct in most of the diagrams in the Peipsi Basin and also throughout the whole Estonia.

The Pinus–Betula PAZ. As before, the main pollen types are *Pinus* and *Betula* having often equal values. *Betula*

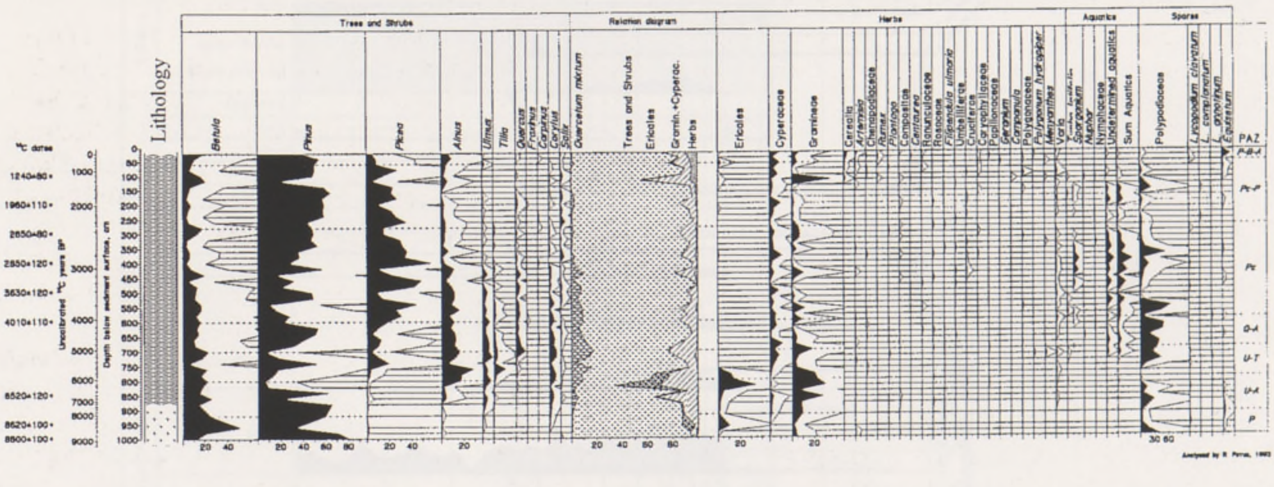

Fig. 35. Pollen diagram from the mouth of the Optjok (Obdekh) River (after Miidel *et al.* 1995).

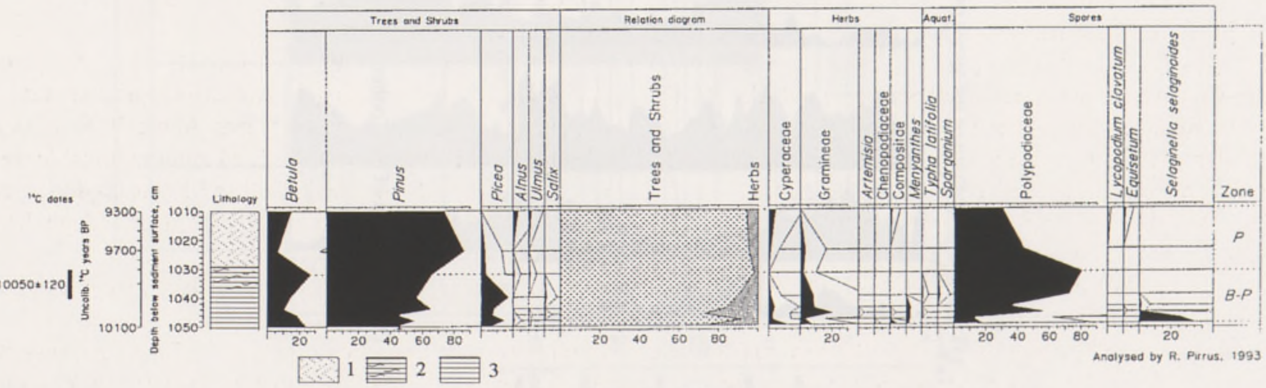

Fig. 36. Fragment of the lowermost part of the pollen diagram from the mouth of the Optjok (Obdekh) River (after Miidel *et al.* 1995): 1- Pharagmites peat containing a mineranogenous fraction; 2 - gray silt rich in organic matter; 3 - gray silt.

may also dominate but the content of *Pinus* is higher than in the previous zone. *Ulmus* and *Corylus* (sometimes also *Alnus*) have very low values. *Picea* pollen may be represented only with single pollen grains. The upper boundary is placed at the level, where *Ulmus, Corylus* and *Alnus* curves rise.

The Alnus–Ulmus–Corylus PAZ. *Betula* and *Pinus* dominate, *Alnus, Ulmus* and *Corylus* are higher than in the previous zone and their curves are rising. The upper boundary is fixed at the level of a sharp rise of the *Alnus* curve, also of the rise of *Ulmus* and *Corylus.*

The Alnus–Corylus–Ulmus–Tilia PAZ. *Alnus, Corylus, Ulmus* and *Tilia* have higher values. In the zone, the permanent curves of *Picea* and *Quercus* start. The upper boundary is placed at the level, where the curves of *Ulmus, Tilia* and *Corylus* rise. The zone is not always distinct and is not found in all diagrams.

The Ulmus–Tilia PAZ. In this zone, the *Quercetum mixtum* curve culminates. *Ulmus* is usually most abundant, *Tilia* has also maximum values. The curves of *Corylus* and *Alnus* are high, too. The upper zone boundary is fixed at the level, where *Quercetum mixtum*, especially *Ulmus*, falls and *Picea* rises.

The Picea–Quercus–Alnus PAZ. The content of *Quercetum mixtum* is lower than in the previous zone, but still remarkable. *Quercus* has its maximum. The *Picea* curve rises. The zone boundary is placed at the level, where the *Picea* curve rises and the curves of *Alnus, Ulmus, Corylus* and *Quercus* decrease.

The Picea PAZ. In this zone, *Picea* has maximum values. Characteristic is the fluctuating curve. *Pinus* and *Betula* are essential again. *Alnus* is lower than in the previous zone. Low values of *Quercetum mixtum* and *Corylus* pollen are observed. The upper zone boundary is placed at the level, where the *Picea* curve decreases and the *Pinus* and *Betula* curves increase.

The Betula–Pinus PAZ. *Betula* and *Pinus* dominate. The content of *Picea* is lower than in the previous zone. *Alnus* forms 10–20%. Other pollen types are low.

Radiocarbon datings. Chronostratigraphy. Altogether 79 radiocarbon dates are available for the Saviku, Võhmajärve, Imatu, Mannjärv, Kalsa, Obdekh and Toolamaa (dated by A. Liiva and E. Ilves at the Institute of Zoology and Botany, the Estonian Academy of Sciences) and Akali (dated by R. Rajamäe at the Institute of Geology, the Estonian Academy of Sciences) sequences. The datings are made mainly on peat, so the ages obtained should be reliable.

The *Betula* PAZ is correlated to the Pre-Boreal Chronozone with no problems.

The ^{14}C date 10,050±120 (TA-2380) from Obdekh refers to the lower part of the Pre-Boreal (Miidel *et al.* 1995). In Mannjärv, the lower boundary is dated to 10,000±120. The date 9400±120 (TA-1805) was derived from the Imatu section. The upper limit of the zone is marked with the date 9090±70 (TA-327) from the Saviku sequence. In the Optjok section, it is approximately 9000 yr BP.

The upper limit of *Pinus–Betula* PAZ is dated to 7800 yr BP at Obdekh and approximately to 7700 yr BP at Kalsa. As the section is compressed, the zonation resolution here is low. The zone may be correlated more or less with the Boreal Chronozone.

The *Alnus-Ulmus-Corylus* PAZ is not distinct in all diagrams. It is well represented at Saviku and Võhma. The upper limit is marked by the date 6900±70 (TA-325) at Saviku and is between the dates of 7560±70 (TA-1782) and 6880±60 at Võhma. The PAZ may be correlated with the Early Atlantic.

The upper limit of *Alnus–Ulmus–Corylus–Tilia* PAZ is marked by the dates 5410±50 (TA-1778) and 5470±50 (TA-1779) at Võhmajärve. It is approximately 5800 yr BP at Optjok and 5000 yr BP at Kalsa. The duration of the zone may range from the upper part of the Early Atlantic Chronozone to the Late Atlantic.

The *Ulmus–Tilia* PAZ with its maximum values of *Quercetum mixtum* is distinct in several diagrams. The upper limit is determined to 4800 yr BP at Optjok, approximately 4000 yr BP at Kalsa and Saviku, and 4200 yr BP at Võhmajärve. The zone may be fitted to the interval from the upper part of Late Atlantic to the Early Sub-Boreal.

The upper limit of *Picea–Quercus–Alnus* PAZ is dated to 3620±60 (TA-1912) at the Toolamaa site and 3480±60 at Akali. It can be fixed at 4000 yr BP at Optjok, 3500 yr BP at Saviku and at 3300 yr BP at Võhma. The limit is somewhat different at Kalsa — 2700 yr BP. In the rough estimation, the zone represents the Sub-Boreal Chronozone.

The duration of *Picea* PAZ, characterized by the numerous peaks of *Picea*, varies also with the localities. The upper limit of the zone is asynchronous being fixed at 2300 yr BP at Optjok, 1300 yr BP at Kalsa, and approximately 2000 yr BP at Toolamaa. At the Saviku site, *Pinus* starts to rise approximately 2200 yr BP as the *Picea* values stay high. The date 1220±90 (TA-1803) marks the boundary at the Imatu sequence.

The *Picea* PAZ and the uppermost *Pinus-Betula* PAZ are asynchronous and correspond to the upper part of the Late Sub-Boreal and the Sub-Atlantic.

The lowermost pollen zones (*Betula* PAZ, *Pinus–Betula* PAZ) are well represented in the diagrams, and proved to be synchronous. As the Holocene plant communities became more differentiated towards the climatic optimum in response to local conditions, the occurrence and transition of the following pollen zones is often asynchronous as certain characteristic features appear at each site. In the upper part of sequences, *Picea* fluctuates and complicates the biostratigraphic correlations.

In Estonia as a whole (Saarse & Rajamäe 1997), the pollen stratigraphy reflecting the Holocene vegetation history of the region shows two major events. About 7000-6800 yr BP, the *Betula–Pinus* forest declined and the broad-leaved forest developed. Some 3800-3200 yr BP, the broad-leaved forest decreased and the *Pinus–Betula* forest with a high share of *Picea* regenerated.

The specific features of the biostratigraphy of the region are as follows:

1. Immigration of *Picea* to the area in the Younger Dryas, its new appearance approximately 7000–6500 yr BP, and expansion during the Late Atlantic and Early Sub-Boreal. In

the Sub-Boreal and Sub-Atlantic, the *Picea* curve has a fluctuating character and cannot, therefore, serve as a stratigraphical marker in biostratigraphy. This has been stressed also on the occasion of the regional stratigraphy of the Haanja area (Saarse & Rajamäe 1997).
2. The Atlantic Chronozone is represented commonly by three PAZ. This is also showed by Saarse *et al.* (1995).
3. *Ulmus* immigrated relatively late and distributed frequently even in the Sub-Boreal.
4. *Quercetum mixtum* culminated in the Late Atlantic and early Sub-Boreal (from 6000 yr BP to 4500–3500 yr BP) and played an important role until 3000 yr BP. *Ulmus* and *Tilia* culminated mostly simultaneously, followed by the *Quercus* maximum.
5. The age of the significant stratigraphical level of the *Ulmus* decline ranges from 4800 yr BP to 3600 yr BP.

2.5. LATE- AND POSTGLACIAL CRUSTAL MOVEMENTS

Already in the early 20th century it was postulated that the northern Lake Peipsi area had been rising faster than the southern part of the lake (Hausen 1913, Mieler 1926, Ramsay 1929). This phenomenon has clearly been affecting the hypsometry in the area and, following from this, lake-level changes and the evolution of the lake. Since there is no direct geological or geomorphological data allowing to evaluate the rate of the uplift and the amount of the tilting within the lake basin, use was made of the extrapolated ancient shoreline data from coastal Estonia. L. Peipsi isolated finally from the Baltic soon after the Weichselian ice had retreated to the Gulf of Finland. For this reason, the above-mentioned data can only be used in drawing conclusions about the amount of crustal tilting, but not about lake level changes. Ramsay (1929) obtained the following values of tilting: more than 20 m on the basis of the Baltic Ice Lake (BIL) B_{III} shoreline data, and 35 and 13 m on the basis of the transgressive shorelines of the Ancylus Lake and Litorina Sea, respectively. Svensson (1989, 1991) demonstrates that according to B_{III} shoreline, the uplift in the northern part of the lake has been 20 m greater than in the southern part.

The currently proposed distance diagram of crustal tilting in the lake basin (Fig. 37), based on multiple data starting from Orviku (1960, 1969) and Pärna (1962) and ending with Kessel and Raukas (1979), still leaves the topic open for discussion. Following the diagram, the extrapolated data of the Ancylus transgressive shoreline points to a 28 m difference, and the same data on the Litorina transgressive shoreline to a 15 m difference in the crustal movements within the lake basin. In order to reconstruct the crustal tilting in the area since the BIL B_{III} phase, the gradient of 18 cm km^{-1} was used. According to Pärna (1962), such a gradient of crustal tilting was characteristic of the area south-east of the hinge line (Pärnu–Navesti River–Narva) and gives a 24 m difference in the land uplift within the lake basin. The presented data cannot be used to characterize lake-level changes, because neither during the BIL B_{III}, Ancylus Lake nor the Litorina Stage was L. Peipsi connected with the above-mentioned bodies of water. Hence, this data shows only the crustal tilting within the lake basin which, as mentioned above, must be considered when drawing conclusions about the evolution of the lake.

The data about the gradients of shoreline tilting in the Peipsi Basin are scattered and have caused contradictory conclusions as to the rate of the crustal uplift (Miidel 1981, Hang *et al.* 1995). Raukas and Rähni (1969) maintain that the highest shorelines in the lake basin have a gradient of 65–55 cm km^{-1}. According to Pärna (1962), the lower shorelines have a gradient of 40–34 cm km^{-1} and the gradient of the BIL B_{III} shoreline is changing from 45–34 cm km^{-1}. At the same time, Liblik (1969) suggests a gradient of only 4–9 cm km^{-1} for raised Peipsi shorelines (Fig. 38). Judging from the data obtained by Kessel and Miidel (1973), Miidel and Vaher (1997) such gradients and, hence, the rate of land uplift, are comparable with the shoreline gradients and land uplift at the end of the Litorina Sea and in the beginning of the Limnea Sea. From this multiple data it may be concluded that the rate of crustal tilting increased during the BIL. In other words, the crustal uplift in Estonia was at its greatest during the BIL. This conclusion is in good agreement with the data by Svensson (1989, 1991), according to whom the maximum change in the shoreline tilting took place not at the time of deglaciation but somewhat later.

As shown above, the gradient of shoreline tilting has continuously decreased since the BIL stage. But the zero gradient has never been recorded. Miidel (1981), arguing that the river terraces at the same altitude in different valleys are of the same age, concluded that the southern part of L.Peipsi Basin remained stable during the Late Weichselian. Later Hang *et al.* (1995), correlating shorelines and river terraces, demonstrated that the crustal uplift took place in the southern part of the lake as well (Fig. 37). It is supported by the altitude of the mineral bottom of the valleys. As long as the Emajõgi and Optjok rivers in the southern part of the Peipsi Basin flowed into the same body of water, their base-levels must have been situated at the same altitude. Accordingly, the bottom of the Emajõgi River must have been at least four metres lower than nowadays. Based on this, the gradient of uplift 5 cm km^{-1} was calculated (Hang 1993). This correlates well with the gradients of the raised shorelines (4–9 cm km^{-1}), established by Liblik (1969 – see Fig. 38). Thus, the area with a zero gradient of uplift and with the shorelines running parallel to each other must be situated south of the mouth of the Optjok River. Evidently, basing on this and referring to the studies by Raukas and Rähni (1969), Mörner (1979, 1980) reached the conclusion that the onset of tilting in the Scandinavian uplift region started right after the Low Baltic (Luga = Haanja) Stadial which he dated at 13,250–12,900 BP.

The Holocene crustal movements are also complicated to record, while the shorelines of that period, if formed at all, are flooded or buried under recent sediments. It is possible that the low water-level shoreline is represented by geomorphologically young coastal formations close to

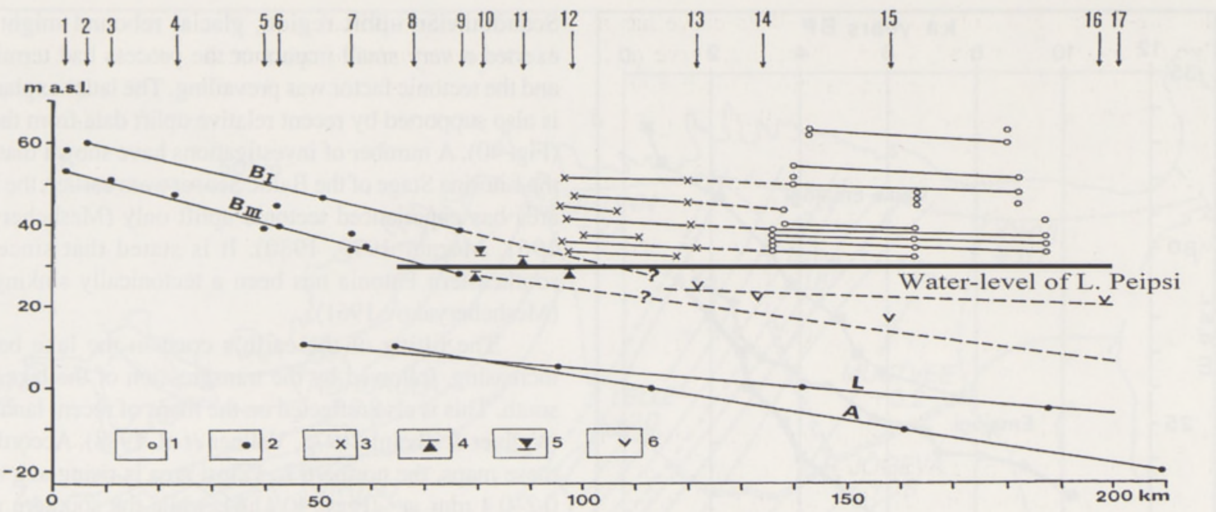

Fig. 37. Distance diagram demonstrating the relationship of the shorelines of the Baltic Ice Lake (BI, BIII), Ancylus Lake (A), and Litorina Sea (L) in the Peipsi Basin (azimuth of the projection line 326°). The diagram is based on the data from Pärna 1962, Sammet 1965, Kessel & Raukas 1979. Correlation of Late Weichselian shorelines of L. Peipsi and river terraces after Hang 1995: 1 – river terraces; 2 – coastal formations of the Baltic Sea; 3 – coastal formations of L. Peipsi after Liblik 1969; 4 – probable thresholds in the source of the Narva R. (10) and in the valley of the Emajõgi R. (11 and 12); 5 – surface in the source of the Narva R.; 6 – valley bottoms. Sites: 1 – Kunda, 2 – Iila, 3 – Kalvi, 4 – Purtse Hiiemägi, 5 – Toila, 6 – Voka, 7 – Laagna, 8 – Rannapungerja, 9 – Kingissepp, 10 – Jaama, 11 – Kaabe, 12 – Kallaste, Kärevere, 13 – Saviku (in the mouth of the Emajõgi R.), 14 – Lääniste (Ahja R.), 15 – Räpina (Võhandu R.), 16 – mouth of the Obdekh (Optjok) R., 17 – mouth of the Velikaya R.

Raadna, southwest of Rannapungerja. A narrow terrace, 5–6 m higher than the present lake level, has formed in front of these formations. The pollen data suggest that a small meadow started to develop on that terrace in the second half of the Atlantic Chronozone and, accordingly, the Raadna coastal formations might have accumulated at the beginning of the same chronozone (Miidel *et al.* 1975). If this assumption is correct, the northern part of the Peipsi Basin has risen 5–6 m since the end of the Atlantic Chronozone. The corresponding transition from the Atlantic to the Sub-Boreal Chronozone in the sediments in the mouth of the Emajõgi River has been recorded at an altitude 4 m lower than the present lake level (Sarv & Ilves 1975). Correlating the altitude of this transition in both of the above-mentioned sites, Miidel (1981) reports the tilting gradient of 14–15 cm km^{-1} (direction of the tilting 236°). This gradient decreases towards the south.

Although the presented data support the idea of continuous southward tilting of the Peipsi Basin, the main proof is the accumulation of peat in the southern part of the basin. The 10-m-thick layer of peat in the mouth of the Optjok River shows that in the Pre-Boreal the water-level was 10 m lower than at present (Miidel *et al.* 1995, Hang *et al.* 1995). The lake level started to rise in the beginning of the Boreal. Somewhat to the north – in Värska Bay – a slow rise took place in the second half of the Boreal (Pirrus & Tassa 1981). Peat deposits at Laane were flooded during the Sub-Boreal (Pirrus *et al.* 1981) and at Alasoo during the Sub-Atlantic (Paap *et al.* 1981). It has been assumed that the water level

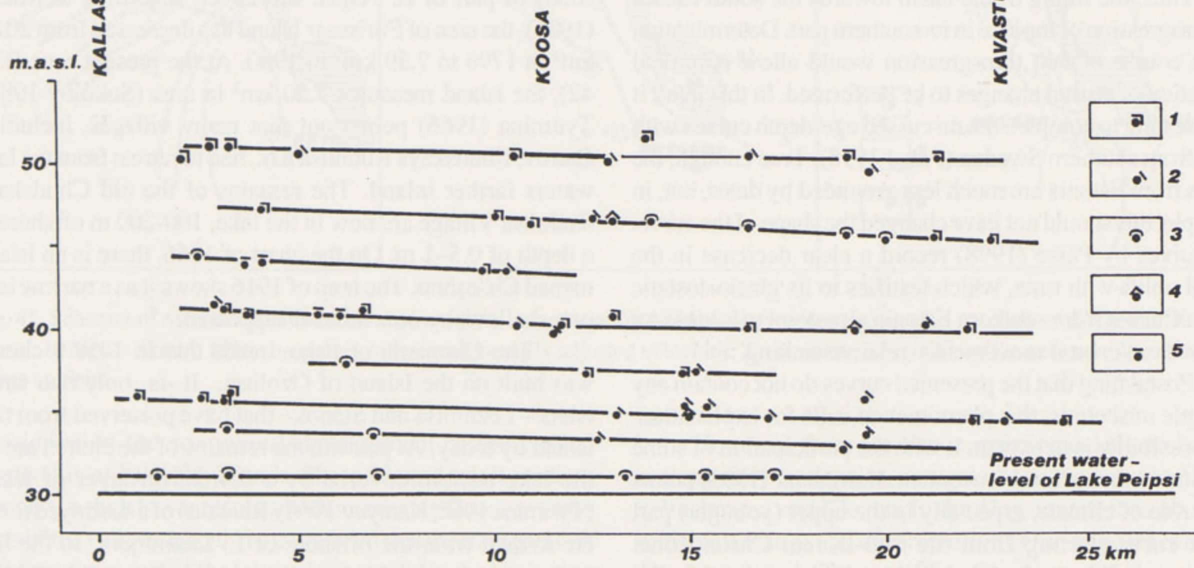

Fig. 38. Late-Weichselian shorelines of L. Peipsi (Liblik 1969): 1 – coastal scarp; 2 – coastal slope; 3 – beach ridge; 4 – stone field; 5 – coastal terrace.

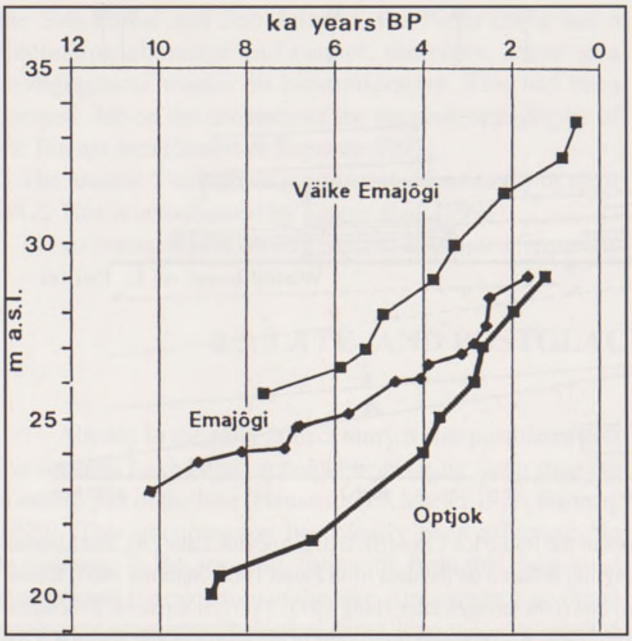

Fig. 39. Holocene lake level changes in the mouths of the Obdekh (Optjok), Emajõgi and Väike Emajõgi rivers, compiled on the basis of ^{14}C datings and on the assumption that the water-level rise and the growth of fen peat were balanced (Hang *et al.* 1995).

rise in L. Peipsi and the growth of fen peat at its shores and mouths of rivers was more or less in equilibrium. Proceeding from this assumption, Holocene lake level changes in the mouths of the Optjok, Emajõgi and Väike Emajõgi rivers were compared (Fig. 39) (Hang *et al.* 1995). The suitability of the Saviku section on the Emajõgi River, which consists evidently of oxbow deposits, for this kind of comparison has also been discussed (Moora *et al.* 1988). Nevertheless, in our mind the comparison is motivated by the similar features of the depth-age curves from the Saviku section and the Optjok River. The presented curves show that the water-level rise was faster in the mouth of the Optjok and Väike-Emajõgi rivers than in the mouth of the Emajõgi River which, taking into consideration the greater effect of crustal uplift in the south, seems reliable.

Thus, the tilting of the basin towards the south causes the transgression of the lake in its southern part. Determination of the course of that transgression would allow empirical estimation of crustal changes to be performed. In this light, it is interesting to compare the discussed age-depth curves with those from southern Sweden (Påsse 1998). True enough, the curves from Estonia are much less grounded by dates, but, in principle, this should not have changed the shape of the curve. The curves by Påsse (1998) record a clear decrease in the crustal uplift with time, which testifies to its glacioisostatic origin. Curves from southern Estonia also point to intense, or at least even crustal movement – relative sinking.

Postulating that the presented curves do not contain any principle miscounts, this phenomenon calls for explanation. One possibility is to explain it with the participation of some factor other than crustal movement. Hang *et al.* (1995) points to the role of climate, especially in the upper (younger) part of the curve starting from the Sub-Boreal Chronozone. Another possible explanation is that since, geographically, southern Estonia is located in the peripheral area of the Scandinavian uplift region, glacial rebound might have exerted a very small impact or the process had terminated and the tectonic factor was prevailing. The latter explanation is also supported by recent relative uplift data from the area (Fig. 40). A number of investigations have shown that since the Litorina Stage of the Baltic Sea, or even earlier, the Baltic area has experienced tectonic uplift only (Meshcheryakov 1961, Mörner 1979, 1980). It is stated that since then southeastern Estonia has been a tectonically sinking area (Meshcheryakov 1961).

The tilting of the earth's crust in the lake basin is increasing, followed by the transgression of the lake to the south. This is also reflected on the maps of recent land uplift (Vallner & Zhelnin 1975, Vallner *et al.* 1988). According to these maps, the northern L. Peipsi area is rising at a rate of 0.2–0.4 mm yr^{-1} (Figs. 40,41A), while the southern part is sinking at a rate of up to 0.8 mm yr^{-1}. A hinge line of southwest – northeast orientation is situated to the southeast from Mustvee Town.

However, in many points, the charts present contradictory data (Miidel 1981, Miidel 1992). For instance, according to the map of recent tectonic movements in Eastern Europe (Karta... 1973) the eastern coast of L. Peipsi belongs to the Ingermanland anticline and is uplifting with at a rate of 2.4 mm yr^{-1} (Fig. 41B). Taking into consideration also the land uplift in the western coast of L. Pihkva and slow sinking in the northwestern coast of L. Peipsi recorded from the map, an unexpected transgression in the northern part of the lake could be concluded. Later, another map (Karta...1986, Fig. 41 C) shows the lake basin as a sinking area with the lake tilting eastwards.

Both these charts are in contradiction with the data presented by Vallner and colleagues (Vallner & Zhelnin 1975, Vallner *et al.* 1988) and multiple historical data from the area. According to "Tartumaa" (1925), a flooded area between Mäksa and Räpina has extended more than 4 km within 35 years. The southeast coast of Salu Island is sinking and gradually disappearing under the water and the area of both L. Koosa and L. Leegu at the mouth of the Emajõgi River, is increasing. All this is indicative of the transgression in the southern part of L. Peipsi. Moreover, according to Mieler (1926), the area of Piirissaar Island has decreased from 20.08 km^2 in 1796 to 7.59 km^2 in 1900. At the present time (Fig. 42), the island measures 7.50 km^2 in area (Sokolov 1983). Tyumina (1966) points out that many villages, including Ostrov, Chudskaya Rudnitsa a.o., had to retreat from the lake waters farther inland. The remains of the old Chudskaya Rudnitsa Village are now in the lake, 100-200 m offshore at a depth of 0.5–1 m. On the chart of 1866, there is an island named Chaeshno. The map of 1916 shows it as a narrow islet or a shallow; by now it has disappeared.

The Chronicle of Pskov reads that in 1458 a church was built on the Island of Ozolitsa. It is only two small islets – Lezhnitsa and Stanok– that have preserved from that island by today. At present, the remains of the church are on the lake bottom covered by a 0.4–2.5 m layer of water (Tyumina 1966, Karayev 1966). Remains of a destroyed forest are known from the offshore of L. Lämmijärv. In the late 19th century, the islands in the southern part of L. Pihkva merged with the mainland (Setumaa 1928). Taking into

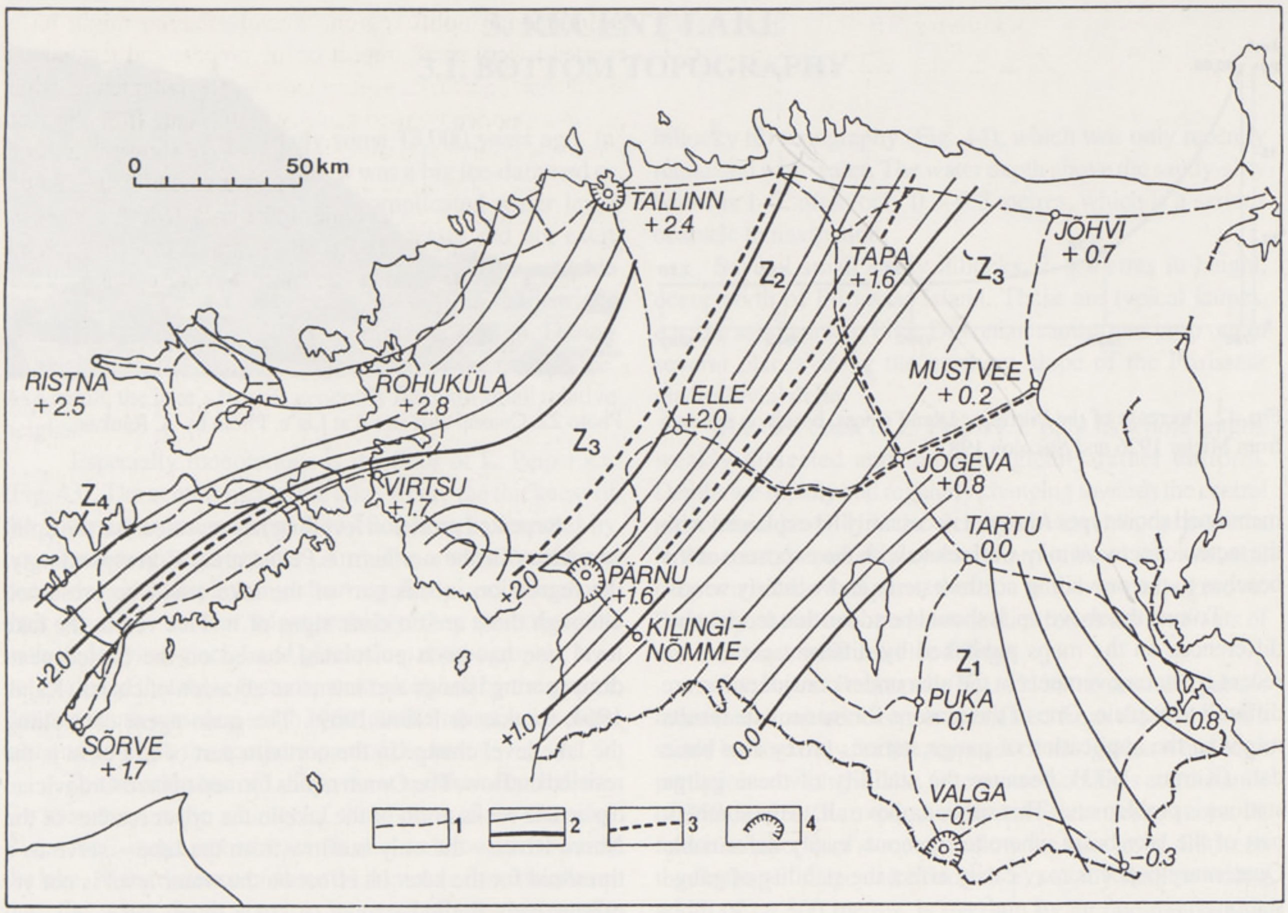

Fig. 40. Scheme of recent vertical movements (mm yr^{-1}) in Estonia (Vallner *et al.* 1988): 1 – levelling network; 2 – isobases; 3 – boundaries of planes of annual velocities; 4 – area of subsidence.

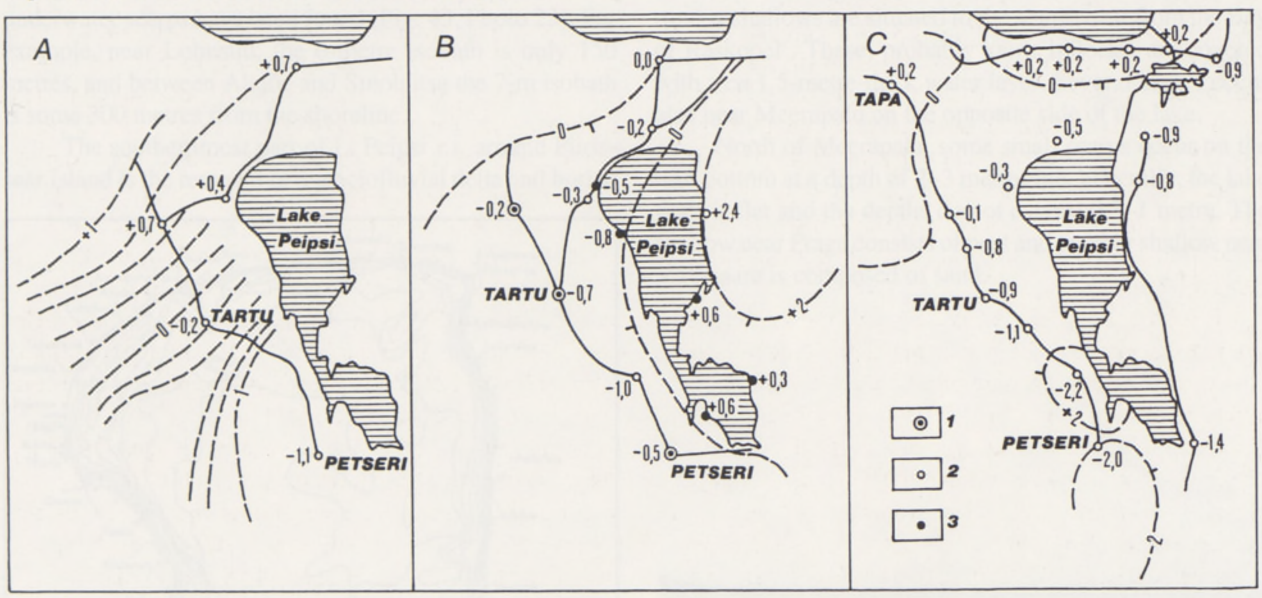

Fig. 41. Schemes of recent crustal movements (mm yr^{-1}), demonstrating different approaches and results: A - Vallner & Zhelnin 1975, B - Karta...1973, C - Karta... 1986: 1 – points with simultaneously adjusted velocities; 2 – characteristic points of the network; 3 – water-level gauge stations.

consideration the continuous development of the coastal scarps (Lis'e, Talabskij Islands, Photo 22) and small gradient of rivers in their lower reaches with increasing floods in those parts of the valleys, it could be concluded that the transgression in the southern part of the lake, which started in the beginning of the Holocene, is ongoing.

However, there are also morphological signs showing contradictory tendencies, particularly on the east coast, close to Vasknarva and Klenno. In drawing conclusions about the land uplift, scarps and the proportion of sandy beaches among the shore types (Tavast 1984, Ch. 4) are of the greatest significance. On the other hand, the prevalence of the above-

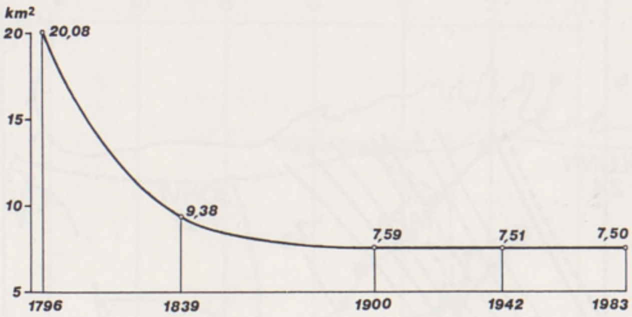

Fig. 42. Decrease of the Piirissaar Island's area, based on the data from Mieler 1926 and Sokolov 1983.

Photo 22. Coastal cliff in till at Lis'e. Photo by A. Raukas.

mentioned shore types must not necessarily be explained with the tectonic factor; it may well relate with the exposure of the beaches to the prevailing northwesterly and westerly winds.

To sum the above up, it should be added that remarkable differences on the maps published by different authors on recent crustal movements in the area under consideration are difficult to explain. One of the reasons for inadequate results might be the application of gauge stations proxy as a basic data (Karta... 1973), because the stability of these gauge stations is problematic. This applies, above all, to the southern part of the lake basin where terrigenous, easily deformable Quaternary deposits may easily affect the stability of gauge stations.

Repeated precision levelling has revealed that the uplift was faster in the northern L.Peipsi area. Correspondingly, the regression in this part of the lake could be presumed although there are no clear signs of it. *Vice versa*, the lake level rise has been postulated, based on the buried peat, disappearing islands and intensive abrasion of coasts (Kajak 1964, Raukas & Rähni 1969). The main agent controlling the lake level change in the northern part of L. Peipsi is the resisted outflow. The Omuti rapids formed of hard Ordovician carbonate rocks north of the lake in the upper reaches of the Narva River – the only outflow from the lake – serve as a threshold for the lake. Its effect on the water level is not yet exhaustively studied.

3. RECENT LAKE

3.1. BOTTOM TOPOGRAPHY

Lake Peipsi formed only some 13,000 years ago. In the early stages of development, it was a big ice-dammed or ice-influenced water basin with complicated water level fluctuations. Short-term erosional processes did not exert much influence on the variegated glacial topography outside the nowadays lake. The present-day lake is shallow; its average depth is 8 m and maximum depth 15.3 m. During all glaciations, the lake basin was eroded by the moving ice. As a result, the lake's floor is generally flat with small relative heights.

Especially monotonous is the floor of L. Peipsi *s.s.* (Fig. 43). The central part of the lake, where the thickness of the water layer ranges from 8 to 10 metres, is featured by occasional oval deeps (up to 12.4 m). The orientation of these deeps does not display any clear regularity and, therefore, their genesis is unclear. Most probably, they are connected with underlying glacial and glaciofluvial topography. It is not excluded that these features were ploughed by the glacier flowing south-west and south in the northern part, and south-east and south in the southern part of the basin.

The coastal zone of L. Peipsi *s.s.* is extremely shallow. For instance, in the vicinity of the mouth of the Emajõgi River, the 2-metre isobath extends about a kilometre towards the lake. At the distance of about 2 kilometres from the shoreline, the water depth starts to increase rapidly practically everywhere until reaching 6–8 metres. The most pronounced changes occur at the northern and eastern coasts, where underwater scarps have been found (Fig. 43, Photo 23). For example, near Lohusuu, the 6-metre isobath is only 150 metres, and between Alajõe and Smolnitsa the 7-m isobath is some 300 metres from the shoreline.

The southernmost part of L. Peipsi *s.s.* around Piirissaar Island is the remnant of a glaciofluvial delta and buried

Fig. 43. Bottom topography of L. Peipsi *s.s.*

hillocky till topography (Fig. 44), which was only recently inundated with water. The water depth above the sandy-silty banks is 1–2, often only 0.3–0.5 metres, which is a serious obstacle to navigation.

Several small sandy hillocks, 2–4 metres in height, occur north of Piirissaar Island. These are typical kames, *e.g.* the sand bank at Piiri. Devonian sandstones crop out in several places along the northern slope of the Piirissaar glaciofluvial delta.

The east coast of L. Peipsi *s.s.* is, at its whole length, weakly dissected and morphologically rather uniform. Depths are rapidly and regularly changing towards the central part of the lake. Only in the vicinity of Vetvenik, the isobaths in the coastal zone display two sharp turns to the west and between shallower parts there is a 9.4-m-deep groove. Probably, the above-mentioned relief forms are remnants of end moraines with dead ice fillings.

The Bay of Raskopel' (Fig. 45) in the northern part, is more than 7 metres deep. It is isolated from the lake with a sandy bar and only for the purposes of navigation it has been partly joined with the lake. One-metre isobath is at a distance of 250 metres from the shoreline where a scarp starts. Its foot is at a depth of 4 metres below the lake level. Between the Raskopel' Cape and Podolesh'e, a sandy shallow with a width of *ca* 800 metres, is overlain by an one-metre-thick layer of water.

The flat-topped and steep-sloped Raskopel' and Ostrovtsy shallows are situated to the south-west from the Bay of Raskopel'. These, probably kame hillocks, are covered with a *ca* 1.5-metre-thick water layer. Several kames occur also near Meerapalu on the opposite side of the lake.

North of Meerapalu, some small scarps occur on the lake bottom at a depth of 2–3 metres but, generally, the lake floor is flat and the depths do not exceed 0.3–1 metre. The shallow near Praga consists of peat and another shallow near Virvisaare is composed of sand.

Photo 23. The mouth of Rannapungerja River at low water level. White breaking waves in the background mark underwater scarp or slope. Photo by A. Miidel

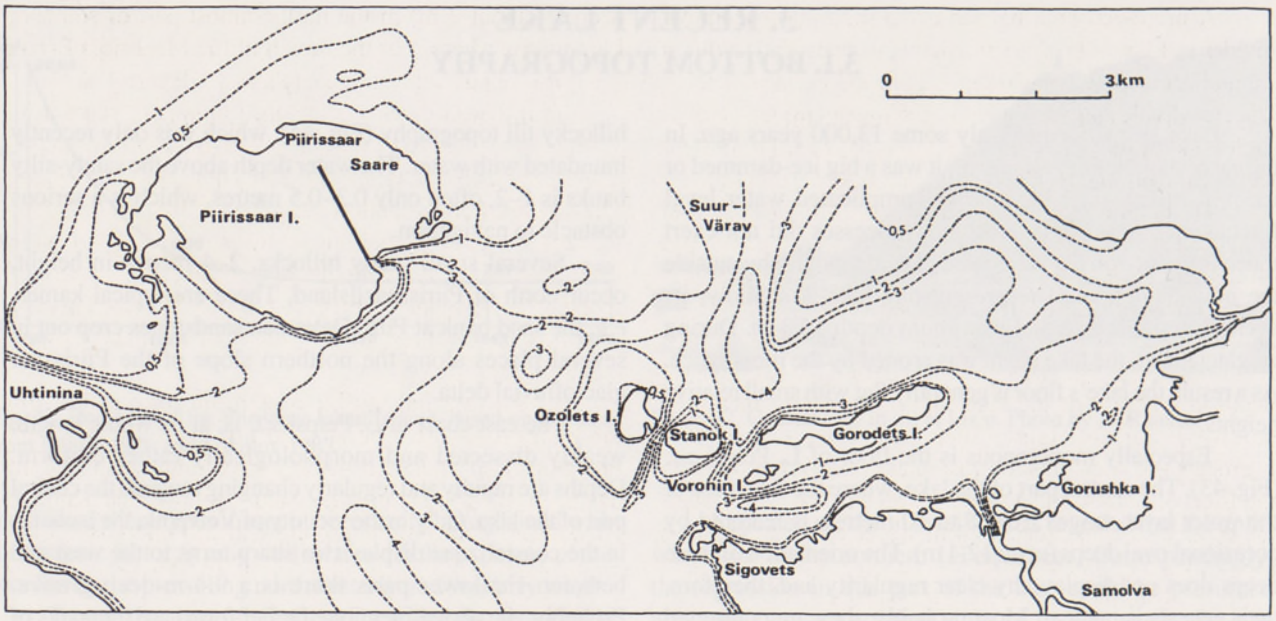

Fig. 44. Bottom topography in the surroundings of Piirissaar Island.

Between Koosa and Kallaste, the shoaly area turns narrower and the scarps at a depth of 2–3 metres near Lahepera conflow. Near the Kallaste Town, the underwater topography is at its steepest, reaching 6 metre depth close to the shore. A steep scarp occurs also near the Kodavere Bank and near Ranna, where the width of the shallow zone does not exceed 100 metres. Northwards, it broadens again and attains its maximum width between Omedu and Lohusuu. In this part of the lake, the 5-metre isobath is 2–3 kilometres from the shoreline. The shallow zone holds many gently sloping hillocks of unknown genesis.

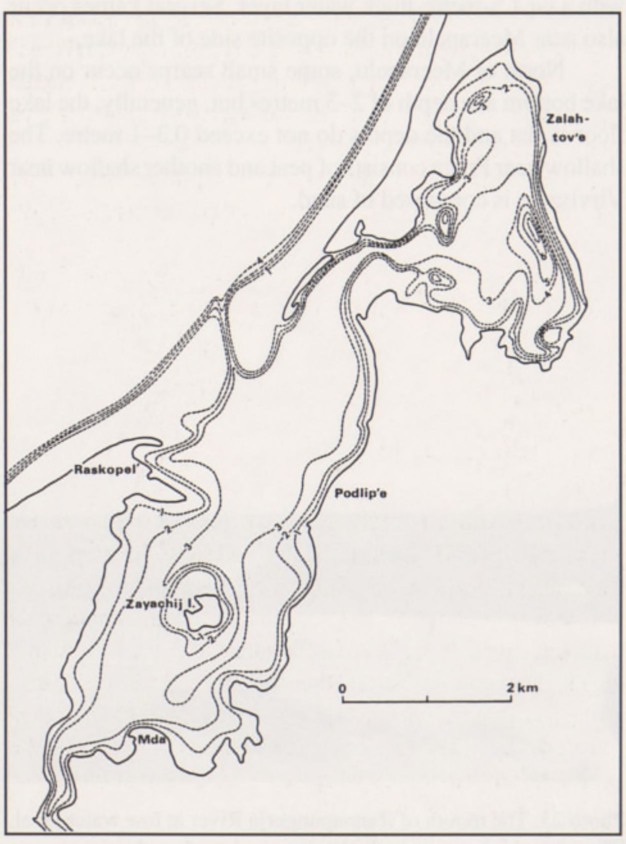

Fig. 45. Bottom topography in Raskopel' Bay.

The bottom topography of L. Lämmijärv is much more complicated (Fig. 46), especially in the southern part. On the lake bottom, the remnants of the ancient river system — the Velikaya with tributaries (Võhandu, a.o.) — are clearly observable. Ancient valleys are filled with Late-glacial and Holocene sediments with different grain-size. The greatest depth — 15.3 metres (Kullus 1969), has been registered here, between Mehikoorma and Pnevo.

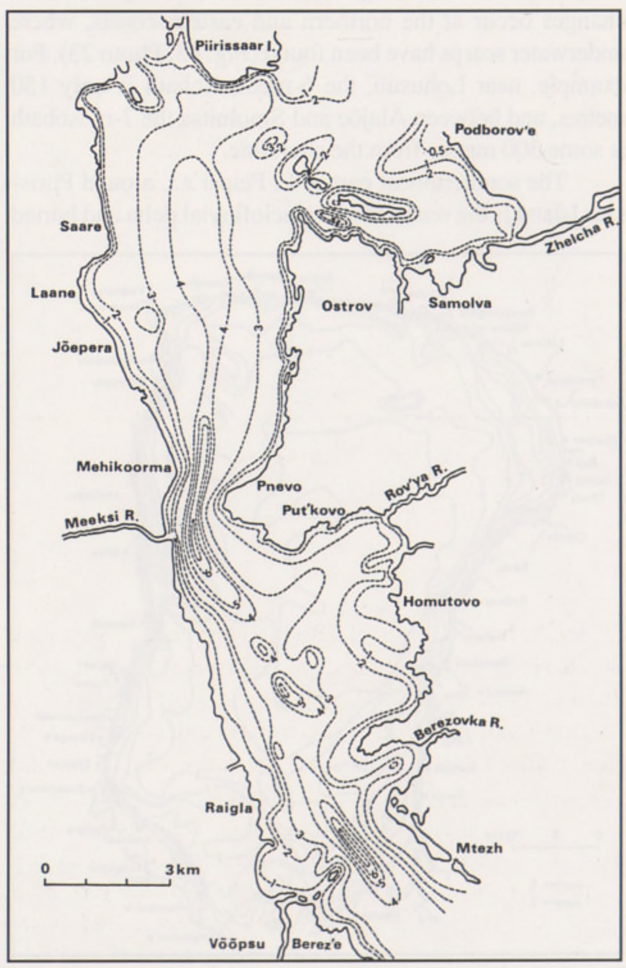

Fig. 46. Bottom topography of L. Lämmijärv.

A narrowed area between Mehikoorma and Pnevo divides L. Lämmijärv into two parts: northern and southern Lämmijärv. About four kilometres to the north and south from the divide there prolongs a deep (over 5 m) and narrow (at Mehikoorma 350 m) hollow (wash-out rill). The eastern bank of the hollow is steeper than the western one (Fig.46).

The bottom topography of northern Lämmijärv is most complicated in the area south-east of Piirissaar Island (Fig. 44), where several shallows and hollows (in some places deeper than 4 m) are situated. The shallows near Podborov'e and the islets of Ozolets, Stanok and Gorodets are the remnants of the Piirissaar glaciofluvial delta. The greatest depth in this part of the lake (4.9 m) has been measured south of Voronin Islet. Such deeps are erosionally overdeepened. Features of erosion are observable between the islets of Stanok and Voronin, where in the "Big Gates" the depth of the lake reaches 4 m. Some erosional hollows are not visible in contemporary topography, because they are filled with up-to-8-m-thick Late-glacial and Holocene deposits, *e.g.* the hollow south of Piirissaar Island.

In southern Lämmijärv, the Salu Hollow is almost as important as the Mehikoorma Hollow. It is 11 km long (8 km in L. Lämmijärv), about 1 km wide and up to 9.6 m deep (Fig. 47), partly filled with Late-glacial and Holocene deposits. The slopes of the Salu Hollow are at their steepest (at least 15 degrees) at the site where it joins the ancient Velikaya Valley. Both valleys, leaving aside filling material, are at least 15 metres deep. The Mehikoorma Hollow is at its deepest (about 7 m) in the northernmost part of southern Lämmijärv, where near the mouth of the Meeksioja Brook the slopes are also the steepest (about 15 degrees).

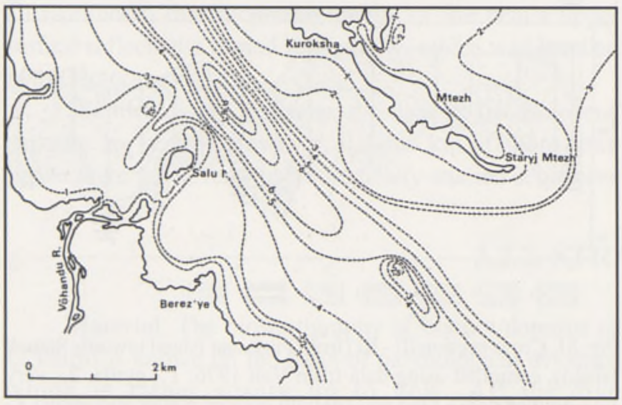

Fig. 47. Bottom topography in the southern part of L. Lämmijärv.

In different parts of southern Lämmijärv some 1.5–2.0-metre-deep grooves (*e.g.* near Raigna) and 2–3-metre-high subwater banks (*e.g.* Salu Bank) have been registered. The eastern shore is shallower and better dissected than the western shore. Here, the one-metre isobath is often more than 1 kilometre from the shoreline, whis causes serious problems in navigation. There is an subwater scarp at a depth of 2 metres at the west shore; from its level the depths are rapidly rising.

L. Pihkva has a flat and monotonous bottom topography. Relative heights do not exceed 0.5–0.7 m. Prevailing depths are 3.5 –4.5 m, maximum depth (5.8 m) has been measured in the Salu Hollow near Kolpino Island. This gently sloping hollow in L. Pihkva is 3.5 km long and up to 1 km wide, narrowing in some places up to 200–300 m.

Depths up to 5.1 m, occur in an area of 12 km^2 in the central northern part of the lake. The waters are shallow around the Talabskij islands, which are the remnants of the glaciofluvial delta and end moraine of the Otepää Stade. Water is deepest (more than 4 metres) between Belov and Talabenets islands, where end moraine was cut through with a rivulet.

As is known, the water level in L. Pihkva is rising, but in most places the accumulation of sediments is more rapid than the water level rise. For example, old navigation charts give *ca* 5 metres for the water depth southeast of Zemsk Island, where nowadays it is not more than 3.5 metres. Near the mouth of the Chernaya River, half a metre of sediments has accumulated during the last 50 years.

The coastal topography is most variable between the Lis'e Peninsula and the Rozhitsy Bank, where several small hollows and a scarp are traceable at a depth of about 3 metres. At the same depth, scarps occur also near the Talabskij islands and the Island of Zemsk.

Some 40 sandy islets have formed in the delta of the Velikaya River. Their number and configurations are rapidly changing.

Numerous 2–3-metre-high underwater stony banks (Kolpino, Kolontsy, Rozhitsy) occur in the western part of the lake.

3.2. LAKE DEPOSITS
3.2.1. THICKNESS AND STRUCTURE

The hitherto research into L. Peipsi has focused on the surface layer of the bottom deposits (Raukas & Rähni 1981, Raukas 1981, Pirrus 1981, Davydova 1981). Data on the thickness and structure of the Holocene deposits is scanty and available only on its southern part — L. Lämmijärv (Pirrus *et al.* 1985, Davydova & Kimmel 1991). The northern part of the lake has been poorly studied. Data have been basically derived from the applied study of sand resources at a water depth of 1–2.5 m, 1–1.5 km offshore from Remniku, between Kodavere and Lahepera and in the surroundings of the mouth of the Emajõgi River (Valt 1976). The information obtained within this study records the thickness and structure of lake deposits also in the surroundings of Piirissaar Island and at some sites in L. Lämmijärv. On the basis of seismic reflection profiling data (Noormets *et al.* 1998), the distribution and thickness of the Holocene and Upper Weichselian lake and glaciolacustrine deposits were established. The maximum thickness (34 m) of those deposits was recorded in the middle part of the lake (see 2.4.1). In the topmost part of that complex an unstable and weak reflector appears in places. An up-to-5m-thick layer upon that reflector was interpreted as loose and gas-rich lacustrine deposit of Holocene age.

In 1996–98, coring was performed from the lake ice at ten sites (Hang *et al.* 1999). These sites were located 2–15 km offshore and had a water depth of 8–10 m. The brownish loamy till under Upper Weichselian and Holocene deposits was reached in a borehole, 2.5 km offshore from Kodavere. Upper Weichselian deposits are represented by glaciolacustrine homogenous or varved clay. At several sites the varved clay was covered by a thin (0.2–0.5 m) layer of sand containing remains of subfossil molluscs. Shells also occurred in the next layer upwards which consisted of lacustrine lime or calcareous gyttja and was present in further offshore sequences. The lacustrine lime is covered by greenish gyttja. The subfossil molluscs bearing layer occurred at a depth of 10.8–10.95 m in the sequences closer to the shore, and at a depth of 12.26–13.85 m in the sequences further out to the lake. The total thickness of the Holocene complex starting with that layer is 0.2–6.5 m. Close to the northern coast, offshore from Alajõe, the Holocene lake deposits are absent or represented by a very thin layer of gyttja (Hang *et al.* 1999). At the same time, in close proximity to the coast, 0.1–0.2 km offshore from Remniku, the fine-grained sand with greyish organic matter forms a 0.8–2.4-m-thick layer (Valt 1976). This may relate to the sediment drift, which took place along the northern coast of L. Peipsi and was orientated from west to east.

The Holocene deposits in L. Lämmijärv have been studied in more detail. More than 170 boreholes (Fig. 48) were drilled for sand prospecting around Piirissaar Island and in an area between Piirissaar and Gorodets islands (Valt 1976). According to borehole records, the thickness of the Holocene deposits ranges from 0.2 to 9.4 m (Figs. 49–51). The smallest thicknesses occur west of the Sigovets Cape, where varved clays are spread (Fig. 50). The greatest thicknesses (7–9 m) have been recorded between the Uhtinina Cape and Piirissaar Island (Fig. 49), south-east of Piirissaar Island and near the southern coast. Holocene deposits are underlain by till, varved clay or bluish-grey clay, probably of glaciolacustrine origin.

The upper surface of the till is 3–11 m below lake level. Till has been found close to the northeastern and southern coasts, between Piirissaar and Ozolets islands and in several other places. The pronounced fluctuation of the till's upper

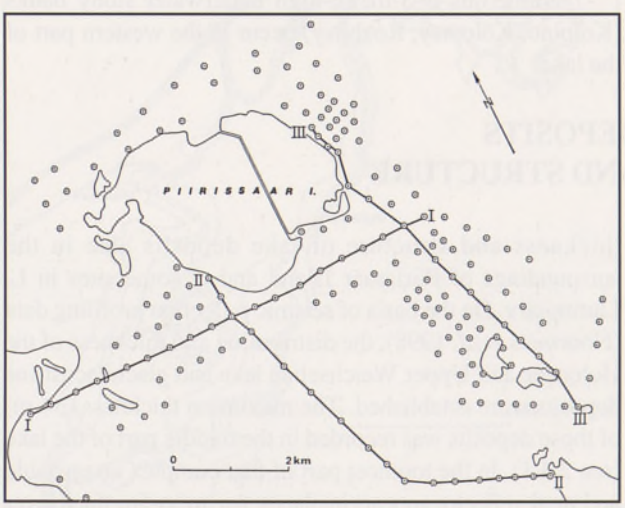

Fig. 48. Location of prospecting boreholes around Piirissaar Island (after Valt 1976): I-I, II-II, III-III and IV-IV cross sections in figs 49–51.

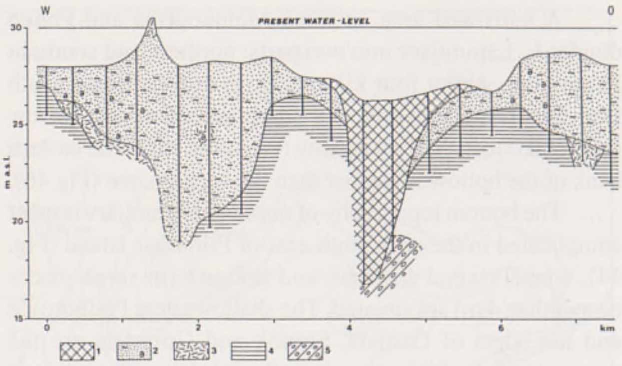

Fig. 49. Cross-section I - I , compiled using data from Valt 1976: 1 - gyttja; 2 - very fine-grained dense sand and silt with organic matter, in places with peat interlayers and shells; 3 - peat; 4 - glaciolacustrine homogenous clay; 5 - till.

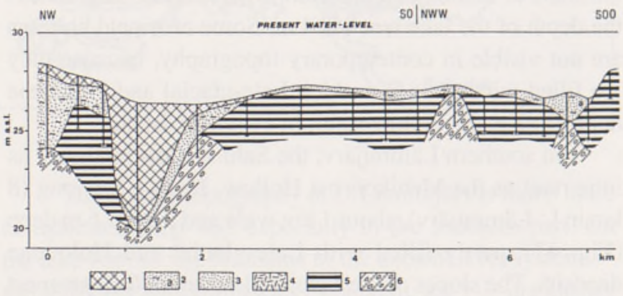

Fig. 50. Cross-section II - II (from Piirissaar Island up to the Sigovets Cape), compiled using data from Valt 1976: 1 - gyttja; 2 - very fine-grained dense sand and silt with organic matter and shells; 3 - very fine and dense sand and silt; 4 - peat; 5 - varved clay; 6 - till.

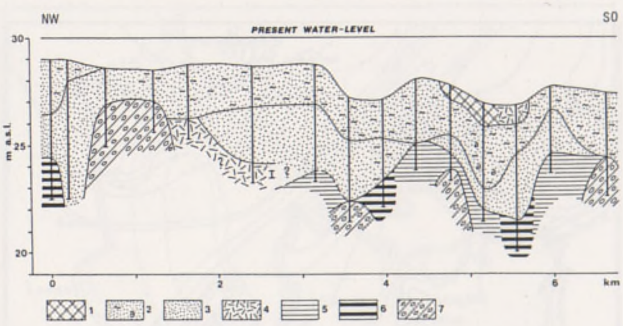

Fig. 51. Cross-section III - III (from Piirissaar Island towards Stanok Island), compiled using data from Valt 1976: 1 - gyttja; 2 - very fine-grained dense sand and silt with organic matter and shells; 3 - very fine-grained sand and silt; 4 - peat; 5 - glaciolacustrine homogenous clay; 6 - varved clay; 7 - till.

surface suggests that in the area under consideration, the Holocene deposits rest upon hummocky morainic topography or a morainic ridge (Figs. 50, 51). It may be that the islands of Ozolets and Stanok, and even Piirissaar have a till core and form the part of subwater morainic topography, which rises above water and was partially levelled during the accumulation of lake sediments in the Holocene period. Piirisaar has also been interpreted as a fluvioglacial delta (Raukas & Rähni 1969).

The Holocene sequence consists of grey, greenish- or yellowish-green compact fine-grained sand and silt, in which organic matter often forms interlayers. The sediments contain mollusc shells and detritus, macroremains of plants, particularly pieces of reed. Peat layers and interlayers occur typically in the area which lies between Piirissaar and Gorodets islands and the Uhtinina Cape. Peat (thickness 0.4–2.0 m) may lie on the lake floor at a depth of 3.7–9.4 m from lake level (Figs. 49,51). The thickness of the peat and the underlying layer is not known. Peat layers of this kind have been recorded north-east and south-east of Piirissaar Island, close to Uhtinina. In several cases, mostly west and south of Piirissaar Island, the blackish-brown well-decomposed peat (thickness 0.3–1.2 m) overlies bluish-grey compact clay, probably of glaciolacustrine origin. Occasionally, it may also rest upon sand or till. Such peat layer is at a depth of 3.8–8.9 m from the water level. Frequently, peat interlayers with a thickness of 0.1–0.5, seldom 0.8–1.1. m, occur between sand and silt layers. Their depth from the water level is variable ranging from 2.9 to 7.4 m. Peat layers in lake sediments have been recorded in the vicinity of Ozolets and Gorodets islands, east of Piirissaar Island and close to the Uhtinina Cape. It is of interest that two peat layers have never been found in one and the same borehole. This suggests that the formation of relatively thin peat interlayers was not controlled by the water level fluctuations in the lake, but pieces of peat were washed into the lake from the coasts and from the adjoining marshy land. At the same time, the wide distribution of peat on the clays, till *etc.* on the bottom of the boreholes shows that during a certain stage in the lake basin development the surroundings of Piirissaar Island underwent terrestrialization. It is not excluded that this portion of land broke the connection between L. Pihkva and L. Peipsi *s.s.* and gave rise to two independent water basins. Afterwards these lakes joined again to form a unitary body of water. In all likelihood, the fluctuating height of the peat's upper surface reflects the buried topography, which was levelled in the Holocene.

Southwards, up to Laane, the data on the Holocene deposits are scanty. However, 0.3–0.6 km offshore from Laane there are several palynologically studied sequences (Pirrus *et al.* 1985, see Ch. 3.2.2) where the Holocene sediments form a complex. It is at least 2.8 m thick; of that 2.5 m is the peat resting on silt. The deeper-lying silt forms a 4.4-m-thick layer. In the coastal zone at Mehikoorma, the Holocene deposits on till are only 0.3 m thick, except in one borehole at a distance of 0.6 km from the coast where the silt with occasional subfossil mollusc shells attains a thickness of 5 m. In the Meeksioja sequence (Pirrus *et al.* 1985, see Ch. 3.2.2) the thickness of lake deposits is more than 2.7 m and it increases up to 5 m near the east coast. In all likelihood, glaciolacustrine sediments are overlain by a 2-m-thick layer of well-decomposed peat (Valt 1976).

In the boreholes, one kilometre north of the mouth of the Võhandu River and 3 km off the shore, the silt overlying till or varved clay and containing subfossil mollusc shells has a thickness of 0.2–0.5 m, but in the boreholes east of the river mouth and at the east coast silt and extremely fine sand with peat interlayers are more than 8 m thick. Also in the vicinity of Salu Island, the thickness of silt exceeds 8 m (Davydova & Kimmel 1991). At the entrance to Värska Bay, Holocene deposits form a 9–12-m-thick layer (Pirrus & Tassa 1981). It is prevailingly sapropel, which occasionally rests upon an up-to-2-m-thick layer of peat (see Ch. 3.2.2).

Little is known about the Holocene deposits in L. Pihkva. According to Raukas and Rähni (1981), these are only 1–2, seldom more than 3 m thick.

The variations in the thickness of the Holocene sediments and the character of the sequence is evidently related to hydrodynamical conditions (waves, currents, depth, *etc.*), which were different in the different parts of the lake at the different stages of its development. Undoubtedly, the isolation of L. Pihkva and L. Lämmijärv from the northern part of L. Peipsi played a significant role. The general sedimentation trend was also controlled by tectonical movements which were different in the northern and southern parts of the lake basin (Raukas & Rähni 1981, Hang *et al.* 1999).

3.2.2. STRATIGRAPHY

Material. The biostratigraphy of bottom deposits of L. Peipsi is known on the basis of palynological evidence which is rather scanty up to now. All the five biostratigraphically studied sequences come from the southern part of the lake. In order to clarify the geological development of the lake basin during the Holocene several sections of bottom deposits were drilled in the 1970s. The results of the analyses of three sections — Meeksioja, Laane 1 and Laane 2 — were published (Pirrus *et al.* 1985). The geological investigations of Värska Bay in 1975 with the aim to specify the stratigraphy of the bottom deposits (Pirrus & Tassa 1981) added new information about the geological development of L. Peipsi, especially L. Pihkva. Two sequences were drilled from the ice in winter 1987 in the southern part of L. Lämmijärv (Davydova & Kimmel 1991, Kimmel 1994).

Lithostratigraphy. The Värska Bay sequence. The bottom deposits in the mouth of Värska Bay, in the northwestern part of L. Peipsi were drilled in 1975. The lithostratigraphy of the deposits is as follows:

1.80– 9.05 m — gyttja;
9.05–9.30 m — peaty gyttja;
9.30–10.00 m — silty *Phragmites* peat with gyttja;
10.00–10.40 m — well decomposed *Lignetum-Phragmites* peat
10.40–10.95 m — moderately decomposed *Phragmites* peat
10.95–11.40 m — silt

The sequence characterizes the development of sedimentation practically through the whole Holocene.

The Laane 1 sequence. The water-covered peat deposit was drilled at a site 5 km north of Mehikoorma Settlement, 300 m to the east from the western shore of the lake. The depth of water at that location was 1.6 m.

The sediment lithology of the Laane sequence is as follows:
1.60–4.05 m — *Lignetum* peat;
4.05–4.10 m — *Hypnum* peat;
4.10–4.30 m — silt with detritus of *Hypnum* peat.

According to the subdivision used, the sequence covers DR3 and the lower and middle parts of the Holocene.

The Laane 2 sequence was taken from deeper-water conditions, 300 m to the east from the Laane 1 sequence, where the water depth was 4.0 m.

The lithology of the sequence is as follows:
4.00–7.45 m — gyttja-rich silt;
7.45–7.75 m — reddish-brown hard silt with traces of aquatics or *Phragmites* roots;
7.75–8.30 m — silt with peat detritus;
8.30–8.80 m — glaciolacustrine deposits.

The sequence was described, but the pollen diagram has not been published.

The Meeksioja sequence is located 3 km to the south from Mehikoorma Settlement. Samples were taken at a distance of 500 m from the western shore of L. Lämmijärv, where the water depth was 9 m.

The lithology of the sequence is as follows:
9.00–10.85 m — clayey silt;
10.85–11.50 m — silt with peat detritus;
11.50–11.70 m — sand.

The deposits represent the lower part of the Holocene.

The Lämmijärv 1 sequence. A 9 m core was taken from the southern part of L. Lämmijärv, north of the Island of Salu (on the line Räpina — Mtezh), 1.3 km from the western shore of the lake. The depth of water was 7.50 m.

The lithology of the Lämmijärv 1 section is as follows:
7.50–15.77 m — dark greenish-brown clayey silt, with pieces of subfossil molluscs in the depth interval 15.50–15.77 m;
15.77–15.82 m — sand with silt containing pieces of subfossil molluscs;
15.82–16.00 m — silt with thin sand layers;
16.00–16.50 m — fine-grained sand.

The Lämmijärve 1 sequence is the most complete profile of the bottom deposits in the deep-water area of L. Peipsi studied so far. Nevertheless, a *hiatus* is expected in the lower part of the Boreal.

The Lämmijärv 2 sequence was taken 200 m east of the Lämmijärv 1 sequence, where the depth of water was 6.5 m. The lithology of the Lämmijärv 2 sequence is as follows:
6.50–8.57 m – dark brownish-grey clayey silt;
8.57–8.75 m — fine-grained sand;
8.75–8.85 m — blackish fine-grained silty sand;
8.85–9.00 m — fine-grained sand.

An extensive *hiatus* in the biostratigraphy is characteristic of the sequence.

Biostratigraphy. The formation of the pollen spectra in the bottom deposits of L. Peipsi has been influenced by several factors.

1. The large area of the lake, due to which the source area of the pollen deposition is vast and the pollen spectra reflect the development of the vegetation in a very big region. According to Pirrus (1981), the pollen and spore spectra of the upper layer of the bottom deposits reflect well the characteristics of the vegetation zones and subzones, in which the lake is situated. The differencies between the geobotanical regions of the basin are also noticeable.

2. The variable character of bottom deposits, especially in the shallow coastal area. From the stratigraphical point of view, the sections from deep-water conditions are more reliable.

3. The complicated hydrodynamical regime. It is particularly characteristic to the narrow L. Lämmijärv in which the bottom topography is more undulating and streams exert greater influence on the bottom sediments.

The biostratigraphy of the studied sequences is based mainly on the changes in the pollen curves of trees and on the ratio between the pollen of trees (AP) and herbs (NAP).

The Värska Bay sequence.The subdivision of the pollen diagram (Fig. 52) presented by R. Pirrus is based on the stratigraphical chart by Kajak *et al.* (1976). Six pollen zones have been distinguished (Pirrus & Tassa 1981).

1. Pollen zone PB (10.95–11.40 m). The pollen spectra are totally dominated by tree pollen, with *Betula* forming 61–73% and *Pinus* 24–38%. The proportion of *Salix* is permanently 1–2%. The zone boundary is placed at the sharp rise of *Pinus* and decrease of *Betula*.

2. Pollen zone BO (9.05–10.95 m). The pollen of *Pinus* predominates (63–82%), *Betula* forms 10–34%. Some *Ulmus, Alnus* and *Corylus* pollen grains are found. The zone boundary marks the decrease of *Pinus* and the beginning of the *Tilia* and *Quercus* curves.

3. Pollen zone AT (7.40–9.05 m). The content of *Quercetum mixtum (QM)* pollen (around 20%), and also *Alnus* and *Corylus* (15–24%) are relatively high, still *Betula* and *Pinus* dominate. The zone boundary is fixed at the *Ulmus* decline and decrease of the *Corylus* curve.

4. Pollen zone SB1 (6.0–7.40 m) is characterized by lower *QM* values than the previous zone. The content of *Quercus* is relatively high but unstable. *Alnus* reaches its maximum. *Picea* makes up about 20%. The zone boundary is marked by the falling *Ulmus* and *Tilia,* and increasing *Picea*.

5. Pollen zone SB2 (4.60–6.0 m) is distinguished by the lower *Picea* maximum. The zone boundary is placed at the decrease of *Picea*.

6. Pollen zone SA (1.80–4.60 m) is dominated by *Pinus* and *Betula*. More detailed stratigraphic division of the zone on the basis of pollen data is considered to be impossible. The first traces of human impact (*Cerealia* pollen) are found at 2.70 m. Also *Rumex*, *Artemisia* and *Plantago* at 2.90 m can be intepreted as indirect evidence of human activity.

The Meeksioja sequence. The pollen diagram (Fig. 53) has been divided by R. Pirrus on the basis of the stratigraphical chart of Kajak *et al.* (1976) into three pollen zones (Pirrus *et al.* 1985).

1. Pollen zone PB (11.30–11.50 m) has been determined by one pollen spectrum only. *Betula* and *Pinus* predominate. Remarkable is the presence of *Picea* pollen (about 10%). In the summary diagram, NAP (mainly Gramineae and Cyperaceae) forms 40%.

2. Pollen zone BO (10.85–11.30 m). *Betula* dominates throughout the zone and reaches its maximum (up to 70%). *Pinus* ranges around 30–50%. Compared to the previous zone, *Picea* is very low. The zone boundary is placed at the sharp rise of *Alnus*, *Corylus* and the QM curve.

3. Pollen zone AT (9.50–10.85 m). The main tree pollen curves are stable. *QM* ranges from 10 to 20%, *Alnus* forms 20–25%. *Picea* is up to 10% and begins to rise only near the upper limit of the zone.

The Laane 1 sequence. The pollen diagram (Fig. 54) analysed by R. Pirrus (Pirrus *et al.* 1985) is divided into six pollen zones on the basis of the stratigraphical chart of Kajak *et al.* (1976).

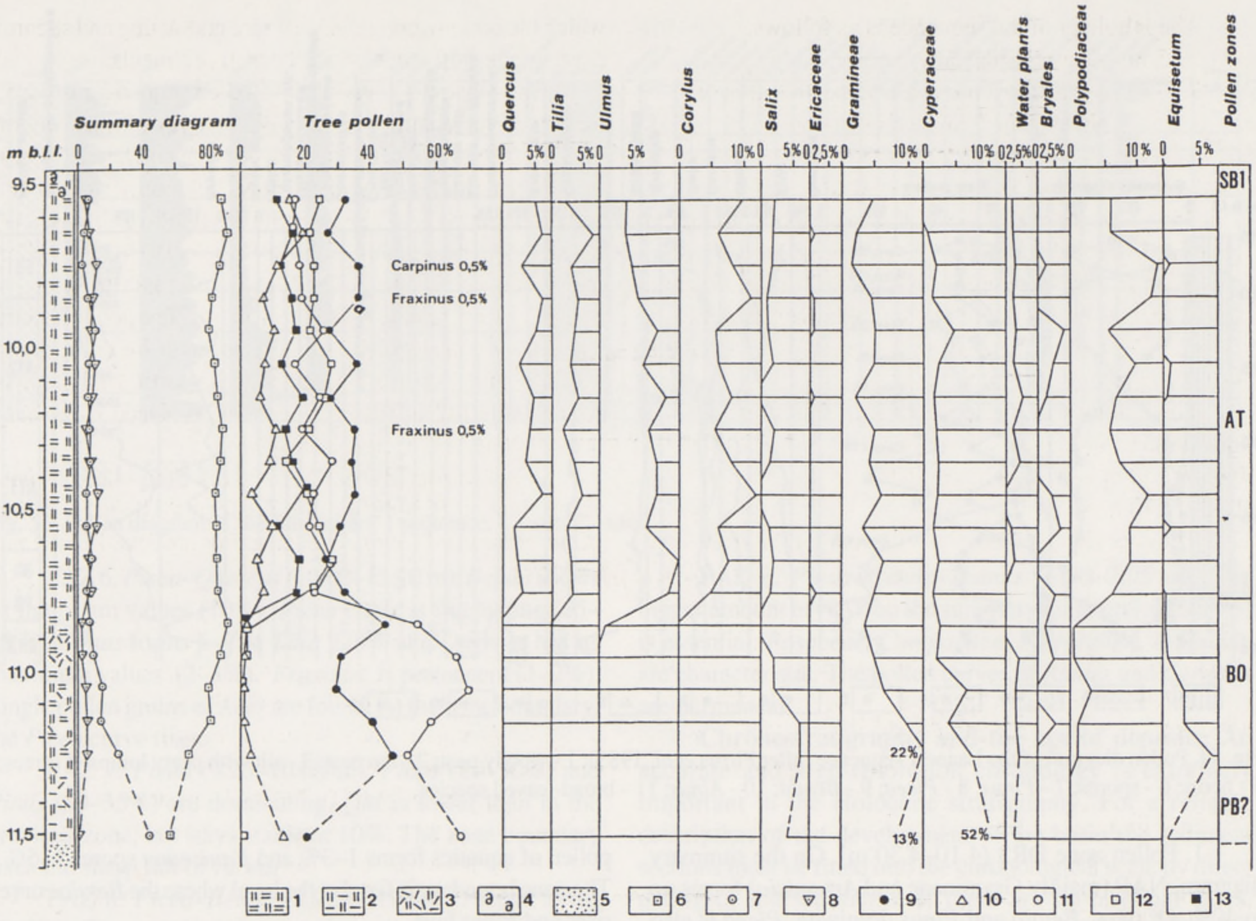

Fig. 52. Pollen diagram of the Värska Bay sequence (after Pirrus & Tassa 1981): 1 - gyttja; 2 - *Phragmites* peat; 3 - *Phragmites* peat with gyttja; 4 - *Phragmites* peat with silt; 5 - clayey silt; 6 - shells; 7 - plant macroremains; 8 - *Pinus*; 9 - *Picea*; 10 - *Betula*; 11 - *Alnus*; 12 - broad-leaved species; 13 - terrestrial herbs; 14 - trees; 15 - spores.

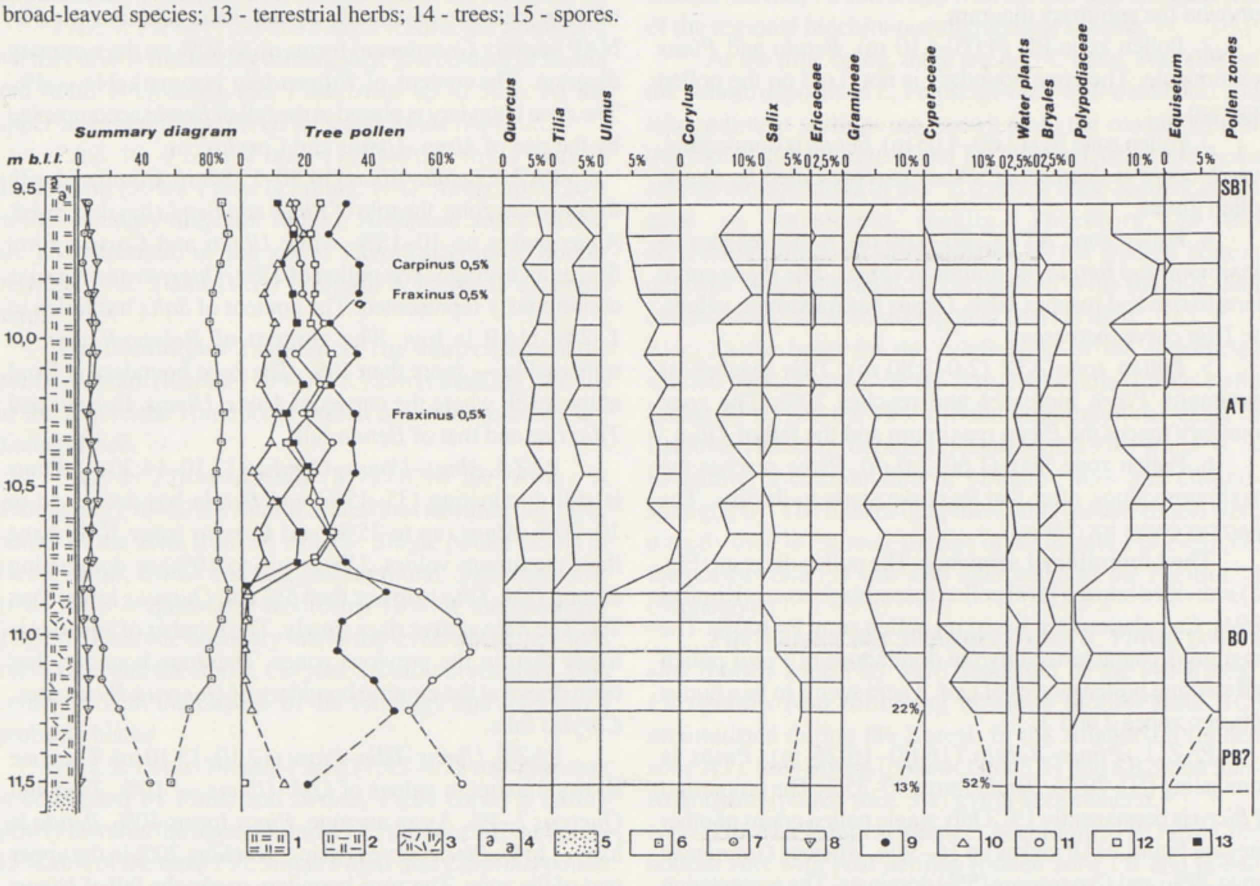

Fig. 53. Pollen diagram of the Meeksioja sequence (after Pirrus *et al.* 1985): 1 - clayey silt; 2 - somewhat clayey silt; 3 - silt with peat detritus; 4 - shells; 5 - sand; 6 - trees; 7 - herbs; 8 - spores; 9 - *Pinus*; 10 - *Picea*; 11 - *Betula*; 12 - *Alnus*; 13 - broad-leaved species.

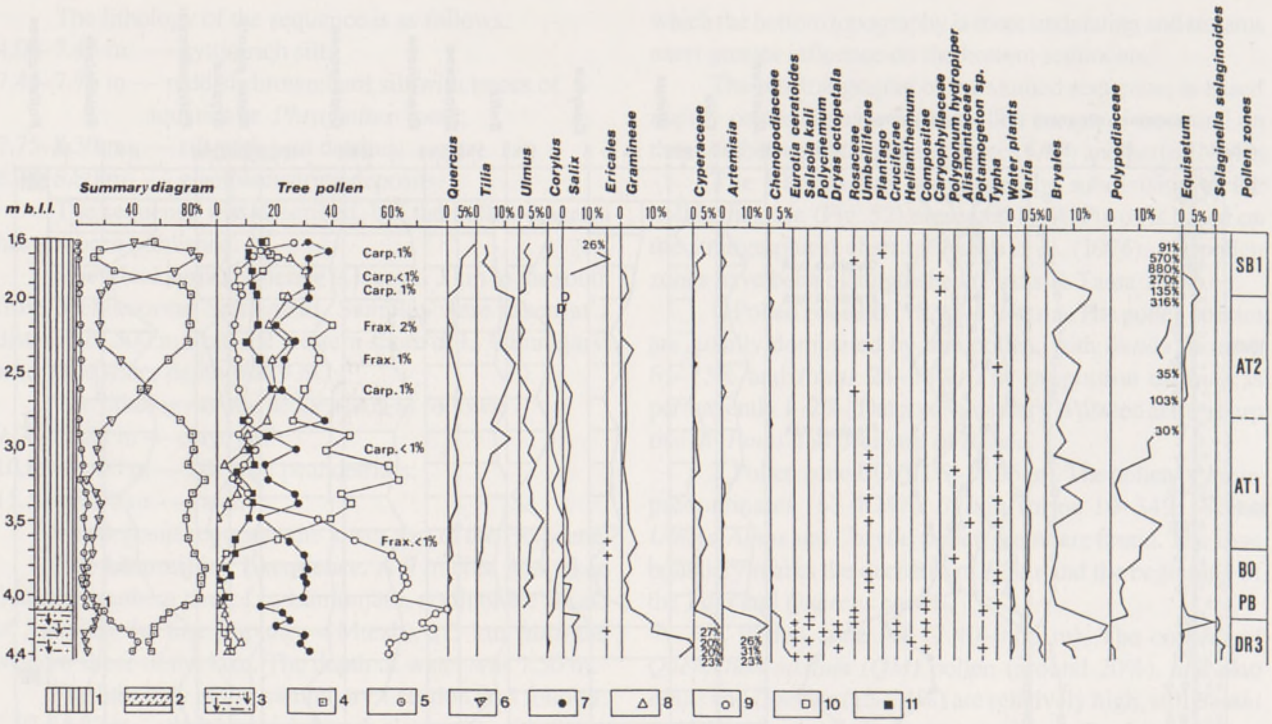

Fig. 54. Pollen diagram of the Laane 1 sequence (after Pirrus *et al.* 1985): 1 - woody peat; 2 - fen peat; 3 - silt with peaty lumps; 4 - trees; 5 - herbs; 6 - spores; 7 - *Pinus*; 8 - *Picea*; 9 - *Betula*; 10 - *Alnus*; 11 - broad-leaved species.

1. Pollen zone DR3 (4.10–4.30 m). On the summary diagram, NAP (mainly Gramineae and *Artemisia*) forms up to 40%. Of trees, *Betula* and *Pinus* dominate, *Picea* is also represented (below 10%), while its curve decreases towards the zone boundary, which is fixed at the decrease of NAP curve on the summary diagram.

2. Pollen zone PB (4.05–4.10 m). *Betula* and *Pinus* predominate. The zone boundary is not fixed on the pollen diagram.

3. Pollen zone BO (3.60–4.05 m). *Betula* is dominating, but its curve is decreasing. *Picea* is represented by single pollen grains.

4. Pollen zone AT1 (2.80–3.60 m). *Alnus* reaches its maximum, and *Betula* its minimum values. The *Picea* curve starts to rise and reaches 10%. *Ulmus* has maximum values, the *Tilia* curve increases.

5. Pollen zone AT2 (2.0–2.80 m). *Tilia* reaches its maximum. *Picea* increases and reaches 20%. The zone boundary marks the *Picea* maximum and the fall of *Tilia*.

6. Pollen zone SB1 (1.60–2.0 m). *Picea* reaches the maximum values, after that its curve starts to decline. The *Quercus* curve increases.

The Lämmijärv 1 sequence. The pollen diagram (Fig. 55) is divided into ten local pollen assemblage zones (Kimmel 1994). Calculations are based on pollen sum, P=100%. The diagram is characterized by the dominance of *Pinus* pollen and slanting pollen curves of *QM*. There seems to be a *hiatus* between zones 2 and 3.

PAZ 1. *Pinus–Betula* (16.00–16.25 m). *Pinus* is dominating (60–70%), *Betula* forms 30–35%. The proportion of *Picea* is permanently 1%. Only single pollen grains of other trees are found. NAP makes up 16–34%, whereas *Gramineae* (up to 20%) and *Cyperaceae* (5%) dominate. The presentation of *Salix* (4–12%), *Betula nana* and *Artemisia* is essential. The pollen of aquatics forms 1–3% and *Equisetum* spores 1–5%. The zone boundary is fixed at the level where the *Betula* curve rises and *Pinus* falls.

PAZ 2. *Betula–Pinus* (15.82–16.00 m). *Betula* dominates (50–70%), *Pinus* makes up 30–50%. Pollen grains of other trees are practically absent. *Salix* accounts for 5%. NAP (mainly *Gramineae)* forms up to 30% on the summary diagram. The content of *Filipendula* is remarkable — 4%. The zone boundary is placed at the fall of *Betula* accompanied by the rise of *Alnus*, *Ulmus* and *Corylus*.

PAZ 3. *Betula–Alnus* (14.20–15.82 m). Compared with the previous zone, the role of *Pinus* and *Betula* has decreased. *Alnus* makes up 10–15%, while *Ulmus* and *Corylus* form 5%, as an average. The pollen of *Tilia*, *Quercus* and *Picea* is continuously represented. The content of *Salix* has fallen to 1–2%. NAP is low. The content of *Polypodiaceae* is remarkable — more than 10%. The zone boundary is fixed at the level, where the curves of *Alnus*, *Ulmus*, *Corylus* and *Tilia* rise and that of *Betula* falls.

PAZ 4. *Alnus–Ulmus–Corylus* (13.10–14.20 m). *Pinus* is still dominating (35–45%) and *Betula* has decreased to 10–20%. *Alnus* (up to 25%) and *Corylus* (over 10%) have their maximum values. *Ulmus* (up to 10%) is dominating among *QM, Tilia* is lower than 5% and *Quercus* lower than 1%. The *Picea* curve rises slowly. The number of aquatics is lower than in the previous zones. The zone boundary has been drawn at the empiric boundary of *Quercus*. *Picea* rises, *Corylus* falls.

PAZ 5. *Ulmus–Tilia–Picea* (12.10–13.10 m). The zone shows maximum values of QM–*Ulmus* — 10%, *Tilia* and *Quercus* 2–4%. As an average, *Pinus* forms 40%, *Betula* is low — 15%. *Picea* curve is rising, reaching 20% in the upper part of the zone. The zone boundary marks the fall of *Ulmus* and *Tilia*.

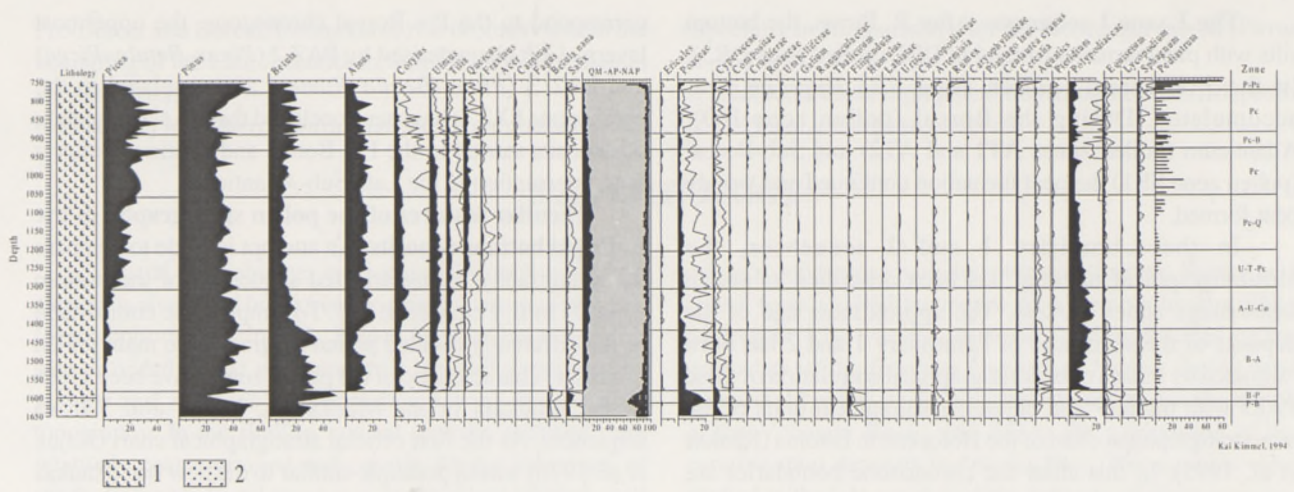

Fig. 55. Pollen diagram of the Lämmijärv 1 sequence: 1 - silt; 2 - sand.

PAZ 6. *Picea–Quercus* (10.60–12.10 m). *Betula* shows its minimum values (10%), *Picea* curve is fluctuating (15–25%). *Ulmus* forms 6–7%, *Tilia* 2–3% and *Quercus* has its maximum values (2–3%). *Fraxinus* is permanent (1–2%). Single pollen grains of *Acer* are found. At the zone boundary, the *Picea* curve rises.

PAZ 7. *Picea* (9.35–10.60 m). *Picea* (20–30%) and *Pinus* (30–35%) are dominating. *QM* is lower than in the previous zone, but stays stable at 10%. The zone boundary fixes the sharp fall of *Picea*.

PAZ 8. *Picea–Betula* (8.92–9.35 m). *Picea* forms 15–20%. The zone is characterized by the fluctuating curves of *Pinus*, *Picea*, *Betula*, and also *Alnus*. *QM* is less than 10%. The boundary coincides with the new rise of the *Picea* curve.

PAZ 9. *Picea* (7.80–8.92 m). *Picea* is the dominant, but the curve is fluctuating remarkably. The content of *Betula* and *Alnus* is unstable, too. *Pinus* rises up to 50%. At the upper zone boundary, *Picea* falls and *Pinus* rises.

PAZ 10. *Pinus–Picea* (7.50–7.80 m). *Pinus* is dominating (50%), *Picea* is falling. NAP forms up to 20% in the summary diagram. *Rumex*, *Artemisia* and *Cerealia* are characteristic of the zone. The stratigraphic border between PAZ 2 and PAZ3 is sharp. A *hiatus* is expected here.

The Lämmijärv 2 sequence. The sharp change in the pollen diagram (Kimmel 1994) at 8.72 m is marking a *hiatus* in the sequence. Three local pollen assemblage zones are distinguished.

PAZ 1. *Pinus–Betula* (8.72–8.90 m). *Pinus* is dominating (50–80%), *Betula* diminishes towards the upper border of the zone from 35 to 10%. Single pollen grains of *Picea*, *Alnus*, *Ulmus* and Corylus are found. *Salix* makes up 10%. NAP is quite low and forms 10% of the summary diagram. The zone boundary marks the level, where the *Picea* curve rises and the *Alnus*, *Corylus*, *Ulmus* curves start. This coincides with the change in the lithology and refers to a probable hiatus.

PAZ 2. *Pinus–Betula–Picea* (7.95–8.72 m). The zone is dominated by *Pinus* and *Betula*, *Picea* curve is falling slowly towards the upper boundary of the zone. *Alnus* forms 15%, rest of the trees 1%. Single *Fagus* and *Carpinus* pollen grains are present. At the upper boundary of the zone *Picea* falls, while *Rumex*, *Filipendula*, *Cerealia* rise.

PAZ 3. *Pinus–Betula–Rumex* (6.50–7.95 m). The bigger amount of NAP on the summary diagram (up to 20%) is essential. *Rosaceae*, *Compositae*, *Filipendula*, *Artemisia* are characteristic. The pollen curves of *Rumex* and *Cerealia* are permanent.

Chronostratigraphy and the age of deposits. An accurate and high-resolution chronology is extremely important in the Holocene stratigraphy. For a reliable description of the development of the basin the sediment sections must be fitted into the chronological scale by direct dating of the sediments. If there are no physical dates, extrapolation of ^{14}C dates of well-dated sites from the region can be done. If this is impossible, the local pollen zones of a studied site may be correlated with the bio- and chronozones of the regional biochronostratigraphical scheme.

At the time being, there are no ^{14}C dates available on the bottom deposits of L. Peipsi. Several well-dated peat and lake sediment sections are known from the coastal area of the lake basin, but due to the big differencies in the basin size the extrapolation of the ^{14}C dates of these sequences can give us unreliable results. Therefore, in the chronostratigraphical interpretation of the studied bottom deposits biostratigraphical correlation with the bio- and chronozones of the stratigraphical chart is used.

As the basis for the subdivision of the sequencies studied by R. Pirrus (Värska Bay, Meeksioja, Laane 1 and Laane 2) served the first official stratigraphical chart of Estonian Holocene deposits (Kajak *et al.* 1976) which is, in its nature, a combination of climato-, bio- and chronostratigraphy. The biostratigraphical units (pollen zones) were directly used to estimate the age of sediments. For example, the pollen zone PB was also interpreted as the Pre-Boreal chronozone.

The Värska Bay sequence. After R. Pirrus, bottom silts (pollen zone PB) were deposited in the Pre-Boreal. *Phragmites* peat following upwards (pollen zone BO) accumulated during the Boreal. In the Atlanticum (pollen zone AT), Sub-Boreal (pollen zones SB1 and SB2) and Sub-Atlanticum (pollen zone SA) gyttja accumulated.

The Meeksioja sequence. According to R. Pirrus, the bottom silts with peat detritus (pollen zone PB and pollen zone BO) formed in the Pre-Boreal and Boreal. The uppermost silt belongs to the Atlanticum (pollen zone AT).

The Laane 1 sequence. After R. Pirrus, the bottom silts with plant remains formed in DR3 (pollen zone DR3). In the Pre-Boreal (pollen zone PB) *Hypnum* peat accumulated. During the Boreal (pollen zone BO), Atlanticum (pollen zones AT1 and AT2) and Sub-Boreal (pollen zone SB1) the peat formation continued and woody peat formed.

In the Lämmijärv 1 and 2 sequences, the biostratigraphical material has been described as pollen assemblage zones (PAZs). The approximate age of the deposits of the sequences of Lämmijärv 1 and 2 has been estimated by means of biostratigraphical correlation of these PAZs with the bio- and chronostratigraphical units of the new stratigraphical chart of the Holocene in Estonia (Raukas *et al.* 1995). In this chart the chronozone boundaries are correlated with those of the Nordic countries and the pollen assemblage zones (PAZ) as time-transgressive units have not been strictly correlated with the chronozones.

The biozones on the chart are the generalization of the information about vegetational development in Estonia, as a whole. The local PAZ of bottom deposits in L. Lämmijärv also represent the vegetation history in an extensive area. Although the pollen assemblage zones of the studied sequences and these of the stratigraphical chart have somewhat different names, the succession of biozones of both of them characterizes the main tendencies in the vegetational history in Estonia, known already from the studies of Thomson (1929). It must be stressed that the result of the correlation is only a rough estimation of the age of the deposits.

The Lämmijärv 1 sequence. According to the biostratigraphical correlation, it is supposed that the bottommost sandy deposits — PAZ 1 (*Pinus–Betula*) and PAZ 2 (*Betula–Pinus*) — correspond to the Pre-Boreal chronozone.

There is a sharp change in the pollen composition which coincides with the change in the sediment type indicating a probable *hiatus* in the sedimentatation. PAZ 3 (*Pinus–Alnus–Corylus*) is correlated to the Late Boreal Chronozone (BO2).

Upwards in the section the homogenous silt has accumulated. PAZ 4 (*Alnus–Ulmus–Corylus*) is correlated to the Early Atlanticum (AT1) and PAZ 5 (*Ulmus–Tilia–Picea*) to the Late Atlanticum (AT2).

As the *QM* curve is gently decreasing and the *Picea* curve is simultaneously gently rising, the zone border between PAZ 5 and PAZ 6 (*Picea–Quercus*) is not distinct. With some reservations, PAZ 6 is correlated to the Early Sub-Boreal (SB1).

PAZ 7 (*Picea*) is correlated to the pollen zone with the maximum of *Picea* in the chart, corresponding to the Late Sub-Boreal chronozone (SB2).

PAZ 8 (*Picea–Betula*) is estimated as corresponding to the chronozone SA1, PAZ 9 (*Picea*) to the chronozone SA2, and the uppermost PAZ 10 (*Pinus–Picea*) to the chronozone SA3.

An abrupt change in the pollen spectra at the level of 15.77-15.28 m is considered to reflect a *hiatus* at the end of the Pre-Boreal and in the beginning of the Boreal.

The Lämmijärv 2 sequence. The bottom sands characterized by PAZ 1 (*Betula–Pinus*) are estimated to correspond to the Pre-Boreal chronozone, the uppermost layers of silt characterised by PAZ 2 (*Pinus–Betula–Picea*) and PAZ 3 (*Pinus–Betula–Rumex*) are correlated to the chronozone SA3. It has been concluded that the accumulation of deposits started in the Pre-Boreal and continued after a long interruption in the Late Sub-Atlanticum.

Peculiar features of the pollen stratigraphy in the L. Peipsi bottom deposits. An attempt is made to compare the stratigraphy of investigated sequences of the bottom deposits in L. Peipsi (Fig. 56). To simplify the comparison of the differently divided pollen diagrams, the main pollen producers characteristic of the pollen zones have been given in the brackets in the Meeksioja, Laane and Värska sequences. As the first official stratigraphical chart (Kajak *et al.* 1976) was in principle similar to the new one (Raukas *et al.* 1995), the correlation of the sequencies seems to be a justified approach. It must be stressed once again, that the biozones are not strictly connected to the chronozones.

On the basis of the studied pollen diagrams it is possible to point out some peculiar features of the pollen stratigraphy of the bottom deposits.

1. The Lämmijärv 1 sequence is a representative sequence for the bottom sediments of L. Peipsi, although there is a short *hiatus* in the sediments.
2. Due to the big size of the pollen deposition basin, the main trends of the succession of pollen zones are similar to those described in the Holocene stratigraphic chart of Estonia.
3. The presence of DR3 spectra in the bottom deposits of the lake is not well proved. These were distiguished by R. Pirrus only in the Laane 1 sequence. L. Saarse (personal comm.) has suggested that the bottommost sands of the Lämmijärv 1 sequence may also belong to the Younger Dryas.
4. On the basis of the studied sequences it is difficult to identify clearly the Pre-Boreal and Boreal pollen spectra because of the complicated geological development of the basin.
5. The *Picea* pollen is represented with low amounts in the

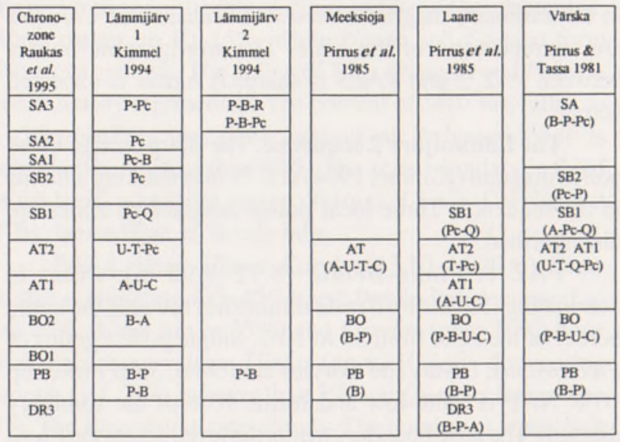

Fig. 56. Pollen zones and pollen assemblage zones of bottom deposits in L. Peipsi plotted against the chronozones of the Estonian Holocene stratigraphical chart (Table 4; Raukas *et al.* 1995) . The Meeksioja, Laane and Värska sequences are divided into pollen zones according to R. Pirrus. For better comparison, the main pollen producers characteristic to the zones are given in brackets. The biozones are not strictly bounded to chronozones. A – *Alnus*; B – *Betula*; C – *Corylus*; P – *Pinus*, Pc – *Picea*; T – *Tilia*; U – *Ulmus*; Q – *Quercus*; R – *Rumex*; A – *Artemisia*.

Pre-Boreal and Boreal. Remarkable rise is observable in the beginning of the Late Atlanticum, somewhat earlier than in western Estonia.

6. Relatively low *Quercetum mixtum* values are characteristic, especially in comparison with the Vooremaa region (Pirrus *et al.* 1987).

7. Additional biostratigraphical studies of the bottom deposits of L. Peipsi are urgently needed.

3.2.3. LITHOLOGY

Coastal and subwater erosion and the suspended load carried by rivers and brooks are principal sources of material for the bottom deposits of L. Peipsi. Some of the material is also provided by the wind and drift ice, and distributed by waves and streams in accordance with the bottom topography. In the shallow coastal regions, sediments are relatively coarse-grained and consist, for the most part, of sands. In the sheltered areas, *e.g.* in the shade of islets, and in the deep central part of the lake, the sediments are prevailingly fine-grained organic-rich sapropels and dominated after the removing of organic matter by clayey silts and silty clays (Raukas & Rähni 1981). Only approximate estimates can be given as to the total supply and the proportion of materials derived from the different sources. The sandy-silty material carried by the Velikaya River, is deposited mostly at the river mouth where, in the long run, a big delta with tens of islets has formed. The contribution of smaller rivers is insignificant. The largest amount of material seems to originate from the erosion of the coastal zone, from both above and below the lake level. However, any quantification of the input from this source has proved impossible.

The basic material used in the present summary was collected in the 1970s and 1980s during the expeditions onboard the research vessels "Leida", owned by the Institute of Geology, and "Limnoloog" belonging to the Institute of Zoology and Botany of the Estonian Academy of Sciences. The sampling density depended on the size of the basin and heterogeneity of the deposits. In L.Peipsi *s.s.,* the distancies between tours were 6–8 km in average, and between the sampling points 4–6 km. In L. Lämmijärv these values were 2–3 km and 1–1.5 km, and in L. Pihkva 3–6 km and 2–3 km, respectively (Fig. 57). Samples were taken mainly with a grip-scoop, 0.025 m^3 in area, and a gravity corer, 1 m in length. For drilling on the lake ice and for sampling, several original pieces of equipment were constructed. Among those was the vibrohammer corer by Endel Rähni which enabled one to receive cores with undisturbed structure. Buried peat layers were studied by means of a Belarus hand drilling equipment. Use was also made of the data obtained by the geologists of the Estonian Geological Survey during sand prospecting in the area of Piirissaar Island and in the mouth of the Võhandu River, and during the exploration of the curative mud deposits in Värska Bay. The samples were studied in the laboratories of the Institute of Geology using traditional methods.

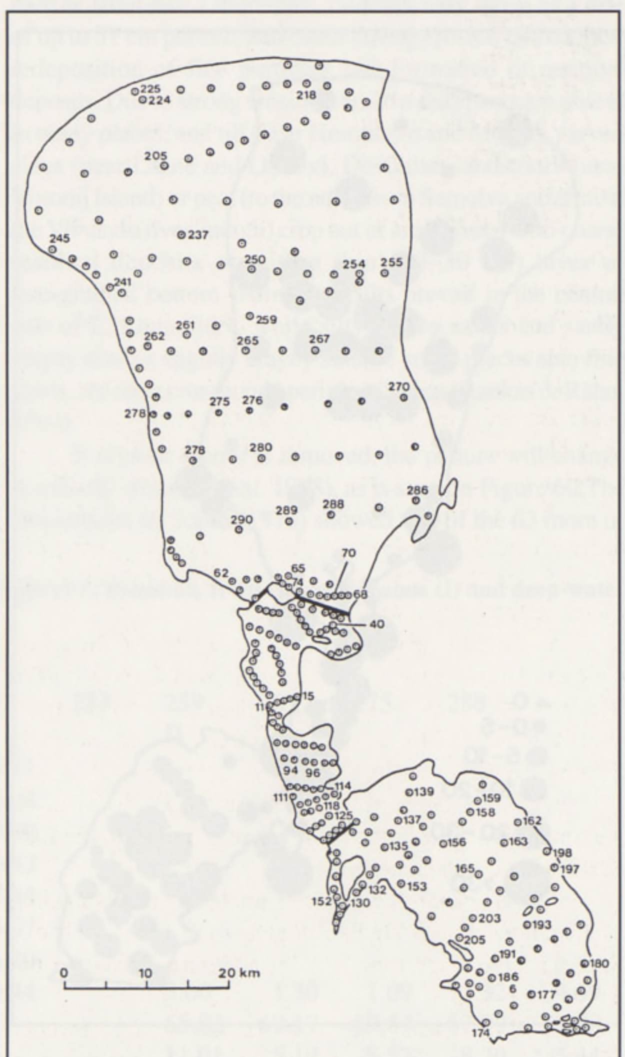

Fig. 57. Sampling sites. Numbers denote the sampling sites presented in tables. Straight lines - boundaries between the different parts of L. Peipsi.

3.2.3.1. GRAIN-SIZE

For the differentiation of mineral particles, the decimal system on the metre-scale was used (Raukas 1964, 1965). Sand (1.0–0.1 mm) was subdivided into coarse (1.0–0.5), medium (0.5–0.25) and fine (0.25–0.1 mm) varieties, silt (0.1–0.001 mm) into coarse (0.1–0.05) and fine (0.05–0.01), and pelite (clay) into coarse (0.01–0.005) and fine (less than 0.005 mm) varieties. Fractions, less than 0.002 mm, were classified as real clay particles. Coarse fractions were subdivided in ascending size into granules (1–10 mm), pebbles (10–100 mm), cobbles (100–1000 mm) and boulders (larger than 1000 mm). Sand, silt and clay mixtures were termed silty sand, (slightly) clayey silt, sandy clay, *etc.*

The bottom deposits of L. Peipsi are rich in organic matter (Fig.58), dispersed mostly in silt and clay fractions. In some areas, including the mouth of the Emajõgi River, some nearshore regions of L. Lämmijärv, it is present as redeposited peat. As to the open parts of the lake, the content of organic matter is at its highest (up to 83.3%) and most variegated in L. Lämmijärv, and rather similar in L. Peipsi *s.s.* (up to 39%) and L. Pihkva (up to 34.5%). In the bays of Värska and Raskopel', there is a specific zone of sedimentation where greenish-grey up to black liquid or semiliquid pelitic highly organic sediments with curative properties occur. The content of fractions coarser than 0.05 mm is, as a rule, less than 10%. The content of organic matter displays a clear regularity: it increases towards fine-grained sediments, being highest in clays, silty clays and clayey silts (more than 10%), lowest (less than 5%) in sands and silty sands and absent in gravel (Figs. 58, 59). It is difficult to judge about the grain-size of bottom deposits, known as sapropels or lake muds, without removing organic particles. For this reason, all grain-size determinations have been done on minerogenous material. To remove organic matter, samples were burnt and pretreated with hydrogen peroxide. Particle-size analysis of samples were carried out using sieving for fractions coarser than 0.05 mm and pipette analyses for finer fractions.

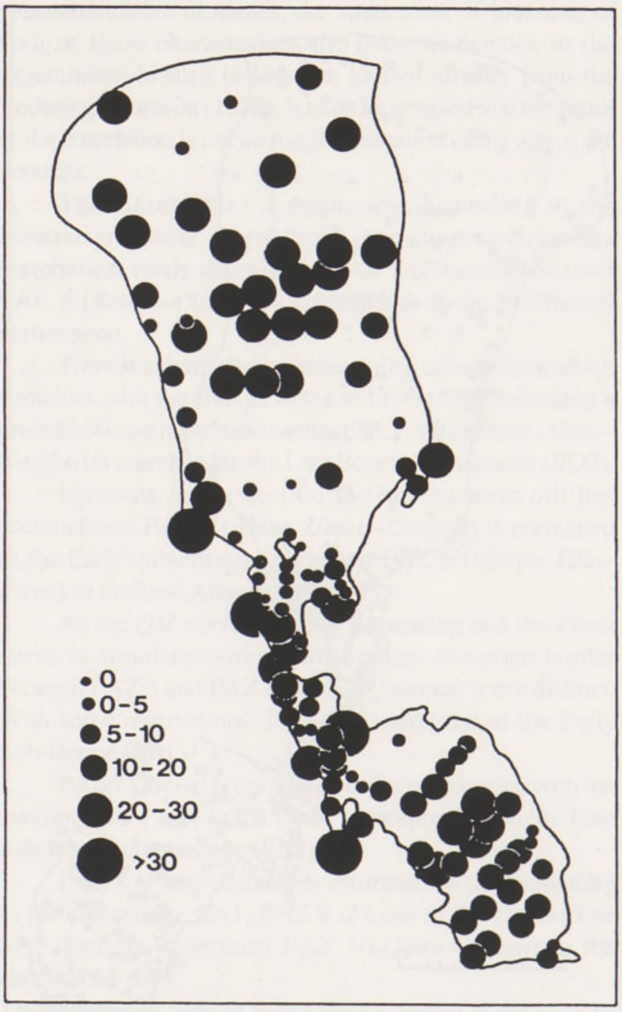

Fig. 58. Organic matter content in the bottom deposits of L. Peipsi, %.

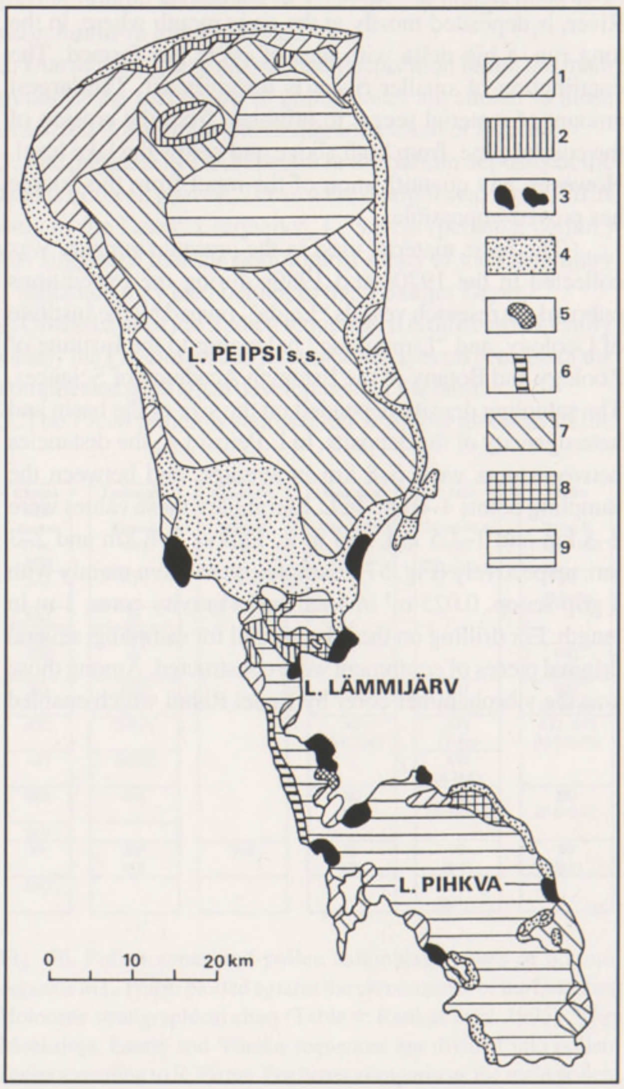

Fig. 59. Bottom deposits of L. Peipsi after removing of organic matter (after Raukas & Rähni 1981): 1 - till; 2 - varved clay; 3 - peat; 4 - sand; 5 - silty sand; 6 - somewhat sandy silt; 7 - somewhat clayey silt; 8 - sandy and clayey silt; 9 - clayey silt.

The process of sedimentation, running under the contemporary physico-geographical conditions, is characterized by a host of specific features, the combined effect of which causes accumulation and erosion areas to change. Wind-drift, wind gradient (compensation), flow, seiche and internal pressure streams (Kallejärv 1973) exist in L. Peipsi.

In L.Peipsi *s.s.*, the surface water moves in the direction of the wind in the northern part of the lake, while anticyclonic and cyclonic circulation is characteristic of the southern part of the lake. Near the bottom in the northern part of L. Peipsi *s.s.*, the compensation streams form; in the shallow southern part, the surface currents reach the bottom.

The facial variety of modern sediments is apparently influenced above all by the availability and amount of suitable sedimentary material. As a rule, erosion and sedimentation proceeded at a stable rate on the borders of the central lake depression, which remained under the water throughout the Late-glacial and Early Holocene, and where initial deposits were also fine-grained (mainly varved clays). The water dynamics was low, which explains the dominance of clayey silts with high organic content in this part of the lake.

The Narva River, the only outflow from the lake, promotes longshore drift to the east along the northern coast of L. Peipsi *s.s.* The sediments gradually block the outflow area, which needs to be dredged periodically. As a result of longshore erosion, modern lake sediments are absent in extensive areas here (Fig. 59), or till and varved clays are covered with a thin layer of residual sediments ranging from several millimetres to some tens of centimetres in thickness. Such sediments result from the erosion of the earlier accumulated deposits, from which finer, mainly silty, pelitic and, in part, also sandy particles have been removed, and the coarser material has remained. On the varved clays, coarser material could have been deposited by heavy waves or strong bottom currents or it may result from the sedimentation of the material incorporated in the ice blocks in the nearshore area. The erosion of the sandy beach in the foreshore between Kauksi and the Narva River, has produced well-sorted (the coefficient of sorting is usually 1.2–1.3) fine- and medium-grained sand (Md 0.15–0.47 mm). This sand, from some tens of centimetres up to several metres thick, overlies the Pleistocene deposits. During low water, the latter are cropping out (Photo 33).

In the southern part of L. Peipsi *s.s.,* in a shallow area with low wave action there is a zone with stable accumulation of sedimentary material in the Piirissaar delta. The delta sediments are covered with poorly-sorted silts, sandy silts and fine sands. The results of particle-size analyses from different parts of L. Peipsi *s.s.* are presented in Table 7.

The bottom topography of L. Lämmijärv is much more rugged than in L. Peipsi *s.s.* or in L. Pihkva and, therefore, modern sedimentation in this part of the lake is much more complicated and the distribution of sediments is more mosaic. Sediments with the same characteristics are spread at different depths, and on the contrary, different granulometric types of sediments can be found at similar depths. L. Lämmijärv has a complicated system of currents (Kallejärv 1973): the water moves from south to north in the lower layers and from north to south in the upper layers under the effect of the wind. Heavy and persistent undirectional winds often cause remarkable water-level changes. As a result, the differences in the water-levels in L. Peipsi *s.s.* and in L. Pihkva can reach 80 cm (Kullus 1972), being in most cases 15–20 cm. In the narrow strait-like Lämmijärv, currents may move at a rate of up to 51 cm per sec, and cause strong erosion on beaches, redeposition of fine particles and formation of residual deposits. Due to strong erosion, modern sediments are absent in many places, and till (near Homutovo and Mtezh), varved clays (near Laane and Ostrov), Devonian sandstones (near Voronij Island) or peat (to the north from Samolva and around the Võhandu river-mouth) crop out or are covered with coarse residual deposits or with a thin (20–30 cm) layer of fine-grained bottom sediments. Silts prevail in the central part of L. Lämmijärv, while silty-clayey sands and sandy clayey silts or slightly clayey silts, in some places also fine sands, are most common in peripheral areas (Raukas & Rähni 1981).

If organic matter is removed, the picture will change drastically (Raukas *et al.* 1988), as is seen on Figure 60.The calculations by Kalm (1976) showed that of the 63 more or

Table 7. Particle-size analyses from the most typical sediment types of L. Peipsi *s.s.* from the coastal zone (I) and deep-water areas (II)

Fractions, mm	Sample sites									
	224	255	262	290	65	237	259	267	275	288
	I						II			
>3.15	3.11	7.61	7.86	0.08	0.11	-	-	-	-	-
2.0–3.15	0.25	0.62	5.57	-	0.04	-	-	-	-	-
1.0–2.0	1.13	2.99	9.81	0.19	0.09	-	-	-	-	-
0.5–1.0	4.98	5.77	14.62	0.30	0.53	-	-	-	-	-
0.25–0.5	35.79	22.43	33.30	11.10	28.54	-	-	-	-	-
0.16–0.25	41.83	26.86	19.05	51.77	64.07	-	-	-	-	-
0.1–0.16	9.07	20.28	6.16	34.13	6.20	-	-	-	-	-
<0.1 or >0.1	3.85	13.43	3.63	2.43	0.44	3.60	1.30	1.09	1.32	4.49
0.1–0.05	-	-	-	-	-	65.82	69.17	69.64	67.99	72.90
0.05–0.01	-	-	-	-	-	11.01	5.19	8.57	8.20	5.44
0.01–0.005	-	-	-	-	-	3.28	3.86	4.64	3.31	3.23
0.005–0.002	-	-	-	-	-	4.06	3.61	2.79	2.63	4.23
0.002–0.001	-	-	-	-	-	2.09	1.55	1.87	2.56	-
<0.001	-	-	-	-	-	10.15	16.22	11.40	13.99	-

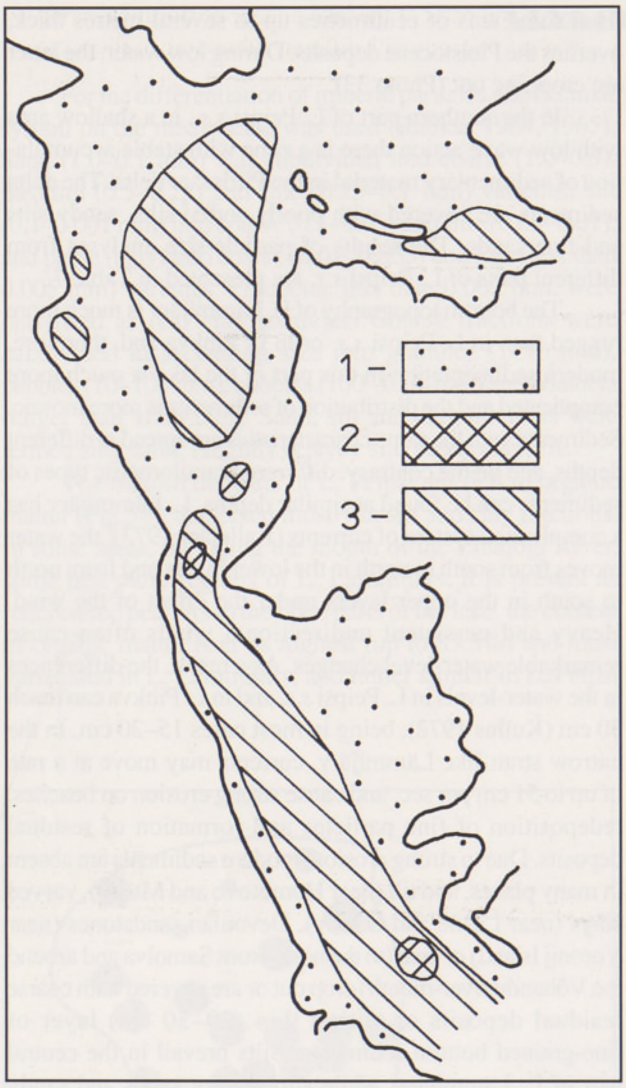

Fig. 60. Sedimentologic types of bottom deposits of L. Lämmijärv defined by particle-size characteristics, as discussed in the text: 1 - sand; 2 - miktites; 3 - silt (after Raukas *et al.* 1988).

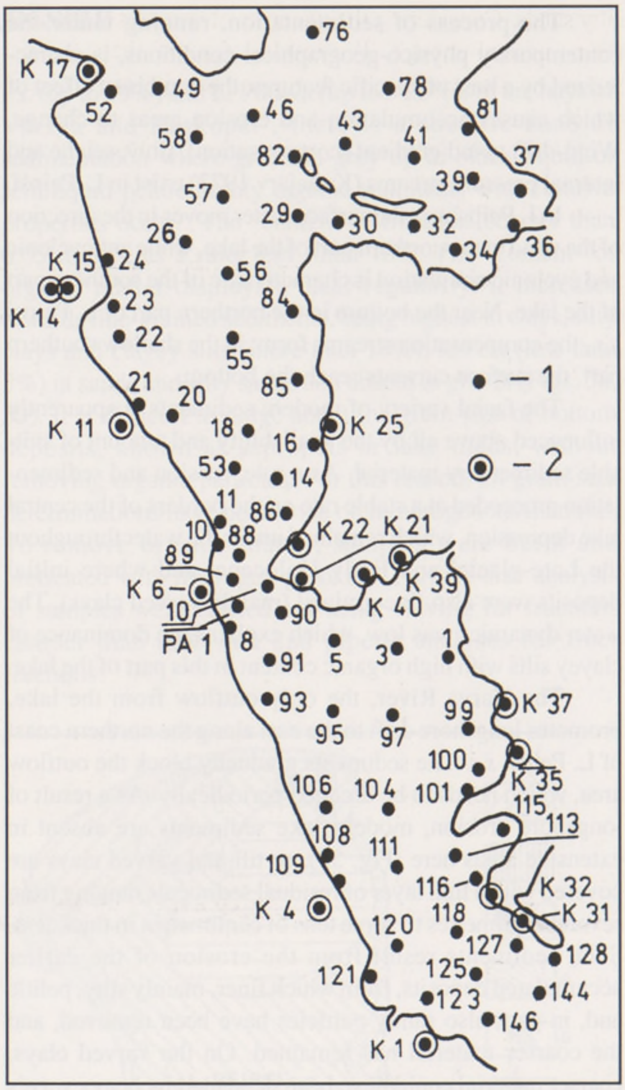

Fig. 61. Sampling sites in L. Lämmijärv: 1 - samples from the bottom deposits; 2 - samples from the coastal and nearshore stations.

Table 8. Particle-size analyses from the deep-water areas of L. Lämmijärv (I) and L. Pihkva (II)

Fractions. mm	Sample sites									
	111	118	125	130	135	163	203	193	186	177
	I			II						
>0.1	1.56	2.51	1.84	3.69	2.57	2.11	1.09	2.79	0.61	0.92
0.1-0.05	64.25	53.54	48.97	64.28	68.42	15.66	62.87	82.21	51.42	38.56
0.05-0.01	8.07	13.99	16.60	9.35	13.27	17.32	5.44	3.72	2.34	15.22
0.01-0.005	4.63	5.32	5.94	2.15	1.89	0.99	2.78	0.33	9.46	9.67
0.005-0.002	2.58	6.36	7.06	3.44	2.60	7.59	5.33	1.67	3.00	8.53
0.002-0.001	1.70	3.68	4.86	2.90	2.00	0.67	3.95	2.37	0.89	5.46
<0.001	17.21	14.61	14.73	14.19	9.25	25.66	18.54	6.91	32.28	21.76

less equally distributed grain-size analyses (Fig. 61), 37 (59%) were sands, 23 (36%) silts, and only 3 (5%) miktites. Among the sands, the following varieties were represented: pure sands (19.1%), sands with silt (14.3%), silty sands (11.1%), slightly clayey silty sands (3.2%) and clayey silty sands (11.1%). The silts were prevailed by slightly sandy and clayey (17.5%) and slightly clayey sandy (12.7%) varieties. Pure varieties were lacking. Silts prevail in the deeper parts of L. Lämmijärv – in the ancient valley of the Velikaya River, and sands in the peripheral parts of the lake (Fig. 60). Silts are the most common sediments also in Samolva Bay, into which they were probably carried by the waters of the Zhelcha River. Sporadically, silts have been found near Laane and Saare on the western coast of the lake. In the vertical section, the deposits are rather homogenous. If there are some changes, then these concern the grain-size which increases in bottom layers, changing from clayey silts to slightly clayey silts, sandy silts to clayey silty sands. Considering the water level rise, this regularity seems normal enough. The results of some characteristic grain-size analyses from L. Lämmijärv are given in Tables 8 and 9.

The granulometric composition of L. Pihkva's sediments is monotonous (Fig. 59). The central part of the lake is covered with clayey and slightly clayey silts. To the south and east of the Talabskij islands sandy and sandy clayey silts occur. The content of sand particles is increasing towards the peripheral parts of the lake and in the north-eastern part of the lake fine sands with different admixture of silt particles prevail. Fine sands occur also near the Island of Kamenka and in the area of the Velikaya River mouth. Like in all deltas, the grain-size here is changing both in horizontal and vertical directions. As elsewhere in L. Peipsi, the bottom deposits in L. Pihkva are poorly sorted. Tables 8 and 10 presents the results of some characteristic granulometric analyses of L. Pihkva.

Table 9. Particle-size analyses of different sediment types of the coastal zone of L. Lämmijärv (after Raukas *et al.* 1988)

Fraction, mm	Sample sites K-1	K-4	K-6	K-11	K-14	K-15	K-17	K-21	K-22	K-25	K-31	K-32	K-35	K-37	K-39	K-40
	Reddish-brown till near Vööpsu	Sandy silt near Ragla	Sand south of Meikoorma	Silty sand near Jõepere	Silty sand near Saare	Sandy silt near Saare	Sandy silt near Uhtina	Sandy silt near Pnevo	Till near Pnevo	Till near Chudskoij Rudnitsy	Silty sand on the Mtezh Peninsula	Sand south of Osotno	Silty sand south of Zazyben'e	Silty sand north of Zazyben'e	Silty sand near Puchkovo	Silty sand east of Puchkovo
>1.0	2.7	0.0	2.1	0.0	0.0	0.0	0.3	1.6	3.5	7.1	0.0	0.2	0.4	0.0	0.0	0.3
0.5–1.0	3.5	0.2	2.4	0.1	0.1	0.0	0.7	3.5	4.3	2.8	0.0	1.2	0.1	0.1	0.6	1.0
0.25–0.5	9.1	7.3	32.4	4.0	1.3	0.6	2.9	2.8	7.0	6.6	1.4	32.3	2.5	1.6	2.9	34.6
0.1–0.25	32.3	25.5	48.4	67.8	50.7	26.0	33.9	61.6	21.4	23.0	71.0	61.5	68.8	61.8	70.0	36.4
0.05–0.1	18.0	50.8	4.3	22.9	45.5	68.2	52.2	20.8	10.2	11.3	22.6	1.2	23.4	34.3	19.7	23.0
0.01–0.05	8.8	14.0	5.2	4.4	2.0	4.0	5.6	6.0	19.2	27.2	4.6	0.4	1.6	1.8	3.6	2.9
0.005–0.01	5.6	2.0	1.2	0.5	0.2	1.1	0.4	0.9	8.4	8.8	0.2	0.4	0.8	0.3	0.4	1.1
0.001–0.005	8.4	0.0	0.4	0.1	0.1	0.1	0.8	0.0	12.0	6.4	0.0	0.0	0.0	0.1	0.1	0.4
<0.001	11.6	0.2	3.6	0.2	0.1	0.0	3.2	2.8	14.0	6.8	0.2	2.8	2.4	0.0	2.7	0.1

For the location of sample sites see Fig. 61.

Table 10. Particle-size analyses from the different sediment types of the shallow-water coastal zone of L. Pihkva (after Raukas *et al.* 1988).

Fractions, mm	Sample sites 139	153	159	162	180	197	198	205
>1.0	1.26	0.86	0.74	-	2.74	0.06	0.26	0.14
1.0–0.8	0.30	0.17	0.01	0.01	0.29	0.06	0.06	0.11
0.8–0.63	0.37	0.26	0.01	0.01	0.30	0.12	0.03	0.24
0.63–0.5	0.71	0.33	0.11	0.07	0.59	0.48	0.14	0.66
0.5–0.4	0.91	0.57	0.25	0.30	1.33	2.45	0.28	3.22
0.4–0.315	0.86	0.86	0.35	2.65	2.61	8.30	0.38	8.39
0.315–0.25	1.07	1.72	1.59	16.07	13.76	17.46	0.60	15.76
0.25–0.2	1.16	7.32	5.58	24.5	25.63	21.46	1.11	18.60
0.2–0.16	2.39	39.75	27.05	31.00	23.68	27.32	5.11	38.14
0.16–0.125	4.51	28.46	18.65	14.69	10.18	9.32	36.99	8.46
0.125–0.1	15.11	8.88	20.13	3.71	6.56	4.70	30.02	1.52
0.1–0.08	60.05	1.87	14.32	1.91	3.55	2.08	15.54	0.53
0.08–0.063	4.07	0.54	2.83	0.34	0.58	0.30	1.61	0.05
0.063–0.05	0.74	0.31	0.63	0.05	0.17	0.04	0.46	0.01
0.05–0.01	0.30	0.33	0.27	0.01	0.14	0.01	0.13	0.03
<0.01	6.22	9.82	7.68	5.62	7.89	5.84	7.67	4.15

For the location of sample sites see Fig. 57.

3.2.3.2. MINERAL AND CHEMICAL COMPOSITION

The petrographic and mineral composition of bottom sediments depends on initial deposits and differs with the parts of the lake (Raukas 1981). Coarse fractions 10–5; 5–2.5, 2.5–1 cm and 10–5; 5–2.5; 2.5–1 mm of modern nearshore sediments are prevailed by the debris of crystalline rocks, which is of the same type as in the near-lying tills. The share of carbonaceous debris is everywhere low.

The mineral composition of fine sand and coarse silt fractions (0.1–0.25 and 0.05–0.1 mm) demonstrates a great qualitative and quantitative instability, caused by the variability of hydrodynamics in different areas, and differences in the initial material. In all likelihood, the essential differences in the mineral composition are also related to wide variations in granulometric composition.

Of more than 50 different minerals and mineral groups identified in the sand and silt fractions, the dominants are quartz, feldspars, carbonates, micas, amphiboles, garnets and ore minerals (mainly ilmenite), which make up about 99.9% of all the minerals recognised. The results obtained by analysing samples from different parts of the lake are presented in Tables 11–14. In comparison with the Pleistocene deposits, lake sediments are richer in micas and accessories. The proportion of the heavy subfraction (density over 2.89 Mg/m^3) is usually small. As an average, in fine-sand fraction it makes up 0.61% in L. Peipsi *s.s.*, 0.54% in L. Lämmijärv and 0.33% in L. Pihkva; in the coarse silt fraction 4.26% in L. Peipsi *s.s.*, 1.86% in L. Lämmijärv and 2.27% in L. Pihkva. The ilmenite-garnet-amphibole association, variable in individual fractions, is present in the deposits all over the lake.

The content of carbonaceous minerals, feldspars, micas and accessories (zircone, fluorite, rutile, sphene, brookite, anatase) in silty fractions increases and the content of quartz, hematite, pyroxenes, amphiboles, epidote, staurolite and disthene decreases. The latter minerals form already in parent rocks bigger crystals. In the northern part of the basin, calcite is the prevailing carbonate mineral in bottom deposits, while in the central and southern parts dolomite predominates as a result of the influence of initial rocks. An important pecularity of the lake sediments is that they do not bear direct relationship to the local bedrock. The minerals have been predominantly carried to the lake through the medium of the Quaternary deposits, hence, just after recurrent redeposition. The decrease in the amount of carbonates and the increase in the quantity of quartz, zircone and tourmaline in a southerly direction should be pointed out as the main quantitative differences. So the content of carbonates in the coarse silt fraction in the bottom deposits of L. Peipsi *s.s.* is 3.9%, in L. Lämmijärv 1.73% and in L. Pihkva 0.29%. In L. Lämmijärv and L. Pihkva we have not found glauconite, which was present in the bottom sediments of L. Peipsi *s.s.* Of the authigenic minerals, ferriferous sulphides (pyrite and marcasite) and ferriferous hydroxides have been found.

Table 11. Examples of the mineral composition of the fine sand (0.1–0.25 mm) of the bottom deposits of L. Peipsi *s.s.*

Minerals	Sample sites													
	62	68	70	74	205	218	224	245	261	270	278	280	286	289
Light fraction (<2.89 Mg/m^3)														
Quartz	84.0	84.2	91.0	85.4	92.7	71.0	88.4	87.0	81.2	94.5	87.5	93.8	87.4	89.0
Feldspars	15.3	15.8	8.7	13.7	6.7	11.0	11.4	12.7	13.7	5.5	12.5	6.2	11.3	11.0
Biotite	-	-	-	0.3	-	7.5	-	-	2.3	-	-	-	-	-
Muscovite	-	-	0.3	-	0.6	-	-	-	1.7	-	-	-	1.0	-
Carbonates	0.7	-	-	0.6	-	-	0.2	0.3	1.1	-	-	-	0.3	-
Other minerals (shell particles)	-	-	-	-	-	10.5	-	-	-	-	-	-	-	-
Heavy fraction (>2.89 Mg/m^3)														
Hematite, limonite	1.0	3.2	3.4	4.3	3.7	1.6	3.3	1.7	1.8	2.6	3.0	9.8	2.2	2.8
Pyrite	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Magnetite, ilmenite	23.0	12.3	19.6	18.0	43.2	18.2	22.0	17.9	10.6	29.6	14.7	24.0	26.4	33.6
Leucoxene	1.0	1.3	3.2	2.4	1.8	1.4	1.0	1.0	1.6	2.0	2.2	0.8	1.6	1.6
Garnet	20.4	26.5	34.8	20.6	19.8	14.7	31.8	21.0	11.6	35.6	21.2	30.0	23.0	25.6
Amphiboles	31.2	23.8	21.6	29.4	8.4	27.0	24.3	41.0	34.8	17.2	33.0	19.6	20.4	17.6
Rombical pyroxenes	0.6	0.6	1.8	1.4	0.8	0.2	1.2	2.8	0.4	1.2	0.6	0.4	0.6	1.0
Monoclinal pyroxenes	1.9	2.9	2.0	1.4	0.6	0.9	2.4	2.5	2.0	0.8	1.6	0.8	1.0	1.0
Biotite	1.0	1.0	0.4	0.2	-	15.2	-	-	20.0	-	1.6	0.8	0.4	-
Zircon	1.7	1.0	2.2	3.9	12.3	8.6	3.9	1.4	1.0	1.2	1.2	2.6	3.8	3.2
Tourmaline	3.3	8.0	4.6	5.1	2.6	-	0.8	3.5	1.4	3.2	3.3	3.8	5.4	3.8
Epidote, zoisite	1.2	2.9	0.8	2.9	1.0	4.6	3.1	2.8	0.8	1.8	3.0	1.6	4.2	2.0
Staurolite	1.7	8.0	1.8	4.9	3.4	-	1.6	0.3	0.4	1.4	1.6	5.0	4.0	4.0
Disthen	0.2	-	-	0.2	0.5	-	0.4	-	0.2	-	-	-	-	-
Andalusite	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Sillimanite	0.3	-	0.2	0.2	0.3	-	0.8	0.4	-	-	0.6	-	-	0.2
Apatite	4.8	3.9	3.4	4.5	0.6	0.7	1.2	1.8	0.8	1.2	4.1	0.8	3.2	2.6
Rutile	0.2	1.3	-	-	0.8	-	-	-	-	0.6	-	-	-	0.4
Sphene	-	-	0.2	-	0.2	0.3	-	-	0.4	-	0.2	-	0.4	-
Brookite, anatase	-	-	-	0.2	-	0.7	-	-	-	-	-	-	-	-
Dolomite	1.5	0.4	-	0.2	-	5.9	1.4	1.1	10.2	0.4	3.8	-	0.6	0.2
Skeletal fragments	5.0	2.9	-	0.2	-	-	0.8	0.7	2.0	1.2	4.3	-	2.8	0.4
Content of heavy fraction	0.34	0.12	0.27	0.21	1.26	0.58	0.93	1.36	0.06	0.85	0.53	0.28	0.49	1.62

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Table 12. Examples of the mineral composition of the coarse silt (0.05–0.1 mm) of the bottom deposits of L. Peipsi *s.s.*

Minerals	Sample sites											
	62	68	70	74	218	224	241	245	254	261	278	286
Light fraction (<2.89 Mg/m³)												
Quartz	88.4	90.7	87.4	85.6	38.8	72.2	62.0	71.0	74.0	74.7	87.3	84.0
Feldspars	9.3	9.3	12.4	13.3	8.5	22.8	21.0	17.8	18.6	18.3	10.9	14.0
Biotite	-	-	0.2	-	27.0	-	2.0	-	2.7	6.0	0.3	0.3
Muscovite	-	-	-	-	5.3	-	6.0	-	-	-	-	0.3
Carbonates	2.3	-	-	1.1	2.6	5.0	8.7	11.2	4.7	1.0	1.5	1.4
Other minerals (shell particles)	-	-	-	-	17.8	-	0.3	-	-	-	-	-
Heavy fraction (>2.89 Mg/m³)												
Hematite, limonite	0.6	0.2	1.8	1.1	1.1	0.4	3.6	0.7	2.0	1.8	0.3	1.4
Pyrite	-	-	-	-	-	-	-	-	-	-	-	-
Magnetite, ilmenite	24.0	44.0	26.2	26.8	29.8	20.2	17.8	28.6	16.1	7.4	31.2	23.6
Leucoxene	3.4	2.0	1.1	2.7	1.7	0.8	1.6	0.3	1.0	2.2	0.8	0.8
Garnet	18.0	21.0	17.4	17.5	32.0	25.0	12.4	24.6	6.3	7.0	28.4	20.4
Amphiboles	26.8	11.0	18.6	22.0	19.8	31.4	41.6	20.4	52.2	37.6	16.6	28.8
Rombical pyroxenes	0.6	0.4	1.1	1.1	0.9	0.8	0.8	0.7	0.4	1.6	0.7	0.8
Monoclinal pyroxenes	1.8	1.2	1.7	2.0	2.1	0.8	1.2	1.2	1.2	1.4	1.3	1.8
Biotite	0.2	-	-	0.4	1.1	-	0.6	-	-	12.4	-	0.2
Zircon	12.4	13.2	21.0	10.7	4.2	11.9	6.4	10.9	10.0	2.2	13.8	7.6
Tourmaline	2.6	1.6	2.0	2.1	1.5	-	1.8	1.6	-	0.2	0.7	1.4
Epidote	2.8	1.6	2.8	3.8	2.1	1.5	4.0	2.6	3.3	2.6	1.5	2.4
Zoisite	-	0.2	0.3	0.1	-	0.6	0.8	-	0.2	0.6	-	0.2
Staurolite	2.4	1.0	0.6	1.4	0.9	0.4	0.2	-	0.2	0.2	1.0	3.2
Disthen	-	-	-	-	-	-	-	-	-	0.2	0.2	-
Andalusite	-	-	-	-	-	-	-	-	-	-	-	-
Sillimanite	-	-	-	0.1	-	-	0.2	-	-	-	0.2	0.4
Apatite	2.2	2.2	3.8	6.4	0.6	1.5	2.2	3.3	3.1	1.6	2.8	3.0
Rutile	1.6	0.2	1.2	0.5	-	0.6	0.4	0.5	0.8	0.4	0.5	0.8
Sphene	-	-	0.3	0.7	0.2	0.2	0.2	0.2	-	0.2	-	0.4
Brookite, anatase	-	-	0.1	-	-	-	0.2	-	0.4	1.0	-	0.2
Dolomite	0.4	0.2	-	0.2	2.0	2.9	2.8	4.4	2.4	14.8	-	0.6
Skeletal fragments	0.2	-	-	0.4	-	-	1.2	-	0.4	4.6	-	2.0
Content of heavy fraction	3.06	7.14	2.94	4.36	3.5	5.5	1.44	10.4	0.6	0.84	9.6	1.69

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Table 13. Examples of the mineral composition of the fine sand (0.1–0.25 mm) of the bottom deposits of L. Pihkva

Minerals	Sample sites								
	137	139	153	159	162	174	180	197	198
Light fraction (<2.89 Mg/m³)									
Quartz	84.4	84.0	80.0	86.0	84.7	58.7	92.3	88.3	87.0
Feldspars	15.6	15.3	20.0	13.7	15.3	2.0	7.7	11.7	13.0
Biotite	-	0.7	-	-	-	-	-	-	-
Muscovite	-	-	-	0.3	-	-	-	-	-
Carbonates	-	-	-	-	-	-	-	-	-
Other minerals (shell particles)		-	-	-	-	-	-	-	-
Heavy fraction (>2.89 Mg/m³)									
Hematite, limonite	8.0	2.1	3.8	5.8	9.0	3.6	7.3	1.4	5.4
Pyrite	-	-	-	-	-	-	-	-	0.8
Magnetite, ilmenite	19.6	27.9	18.8	12.8	18.4	31.6	26.8	28.6	20.4
Leucoxene	1.6	3.0	0.8	3.2	4.2	2.0	2.3	1.2	2.8
Garnet	18.8	21.0	11.8	20.4	13.6	10.0	28.6	38.0	9.4
Amphiboles	24.4	20.6	44.2	27.0	29.2	21.4	14.3	10.8	35.2
Rombical pyroxenes	2.0	0.7	0.8	2.6	0.6	1.2	0.9	0.6	2.6
Monoclinal pyroxenes	1.6	2.0	2.0	1.6	2.0	1.0	0.7	0.6	2.8
Biotite	1.0	0.1	0.2	0.6	0.2	1.4	0.7	-	0.8
Zircon	4.2	8.7	2.2	2.0	2.0	3.2	3.6	4.8	4.2
Tourmaline	6.0	3.9	4.0	8.2	7.4	7.4	5.2	5.0	3.0
Epidote	2.6	2.6	2.6	1.8	2.0	2.0	2.1	2.2	3.2
Zoisite	0.2	0.3	-	0.4	-	-	0.2	0.2	-
Staurolite	3.6	3.0	2.4	7.6	5.2	6.6	3.2	4.2	3.6
Disthen	0.2	-	0.6	0.2	0.2	0.4	0.3	0.2	0.4
Andalusite	-	-	-	-	-	-	-	-	-
Sillimanite	0.4	0.2	-	0.4	0.8	-	0.2	-	0.8
Apatite	4.4	3.0	5.0	3.8	4.0	2.0	2.9	1.8	3.0
Rutile	0.8	0.8	0.6	0.8	0.2	0.2	0.5	-	0.6
Sphene	0.2	-	-	0.2	0.8	0.4	-	0.4	-
Brookite, anatase	-	-	-	0.4	-	-	-	-	-
Dolomite	-	0.1	0.2	-		-	-	-	-
Skeletal fragments	0.4	-	-	0.2	0.2	5.6	0.2	-	1.0
Content of heavy fraction	0.07	0.22	0.11	0.28	0.78	0.69	0.07	0.67	0.09

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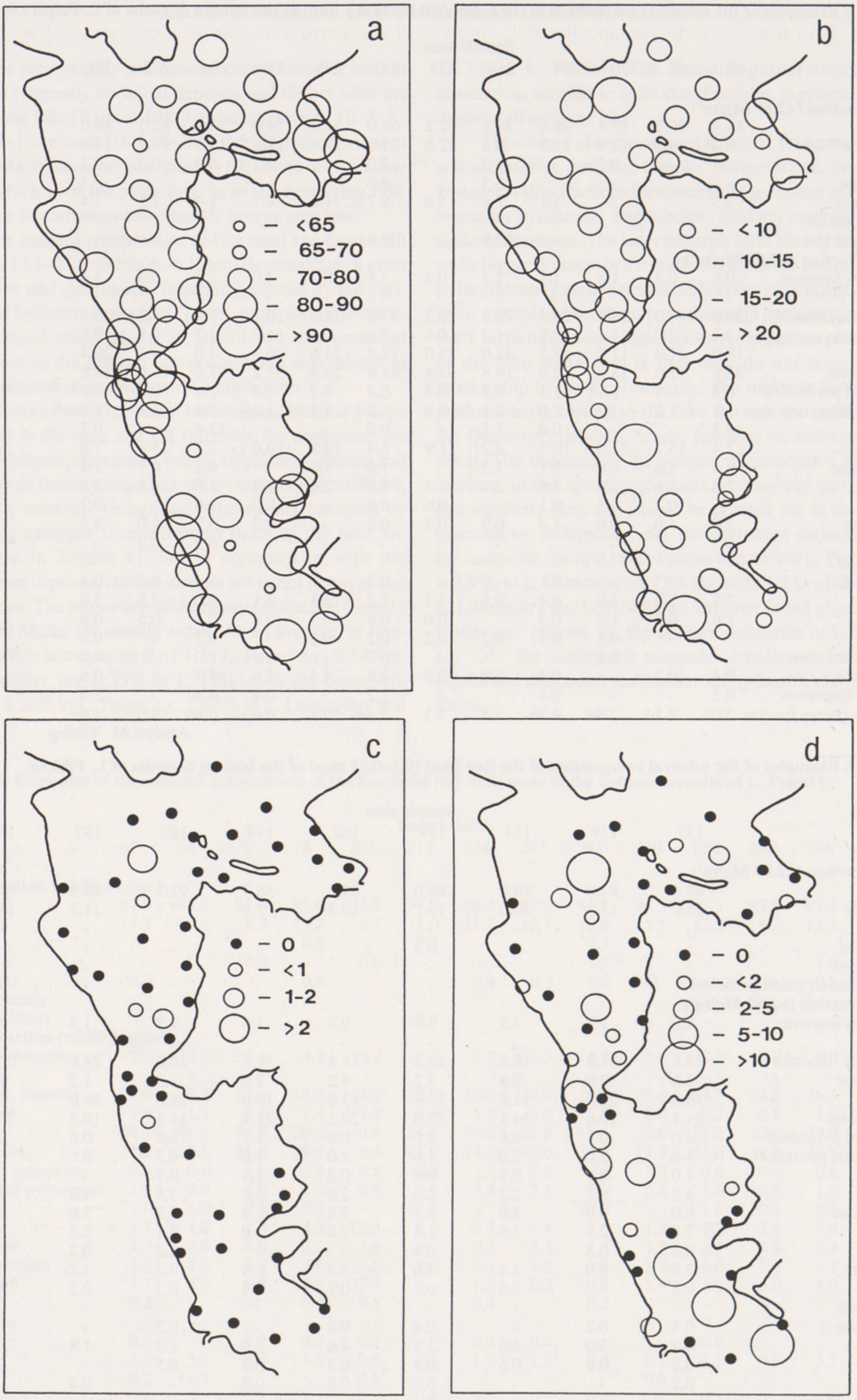


Fig. 62. Distribution of the most important light (<2.89 Mg/m^3) minerals of the fine sand fraction (0.1-0.25 mm) in the bottom deposits of L. Lämmijärv (after Raukas *et al.* 1988): a - quartz; b - feldspars; c - carbonates (calcite and dolomite); d - micas.

Table 14. Examples of the mineral composition of the coarse silt (0.05–0.1 mm) of the bottom deposits of L. Pihkva

Minerals	Sample sites												
	130	132	137	139	145	153	156	159	162	174	177	180	197
Light fraction (<2.89 Mg/m³)													
Quartz	22.0	53.3	79.4	82.7	40.0	75.0	25.1	81.6	83.4	34.0	35.3	89.0	88.6
Feldspars	1.9	15.0	20.6	17.3	4.9	22.8	5.8	18.0	15.3	5.3	-	10.7	11.4
Biotite	-	2.9	-	-	1.8	0.9	0.6	-	-	-	-	-	-
Muscovite	-	-	-	-	-	0.4	-	-	1.3	1.3	1.5	-	-
Carbonates	-	-	-	-	-	0.9	-	0.2	-	-	2.4	0.3	-
Other minerals (shell particles)	76.1	28.8	-	-	53.3	-	68.5	0.2	-	59.4	60.8	-	-
Heavy fraction (>2.89 Mg/m³)													
Hematite, limonite	2.4	2.2	1.6	1.8	3.2	1.4	2.4	3.4	4.7	3.2	2.2	3.1	2.1
Pyrite	-	-	-	-	-	-	-	-	-	-	-	-	-
Magnetite, ilmenite	26.2	21.6	23.9	26.8	32.4	26.4	28.8	19.0	25.0	37.4	11.6	39.0	33.6
Leucoxene	-	1.2	3.0	3.8	-	5.6	1.0	3.0	3.7	2.5	-	0.4	1.1
Garnet	19.2	13.2	20.2	12.1	15.4	22.6	11.6	15.2	17.3	8.9	18.4	16.0	22.2
Amphiboles	28.4	36.4	23.2	32.2	26.6	19.6	27.4	37.6	24.6	24.0	34.8	15.8	13.1
Rombical pyroxenes	0.4	0.8	0.6	0.4	0.6	0.2	0.2	1.0	0.8	0.2	1.2	1.2	0.7
Monoclinal pyroxenes	1.0	1.2	1.0	2.1	1.8	1.0	0.8	0.8	0.8	0.9	1.2	0.6	0.6
Biotite	-	-	-	-	0.4	-	-	0.2	-	1.0	-	0.2	-
Zircon	11.6	12.6	12.1	9.8	10.0	8.6	17.6	8.6	7.7	11.9	16.4	15.7	18.6
Tourmaline	0.6	1.2	2.7	1.8	-	4.2	0.2	5.0	3.5	0.3	0.4	1.7	1.1
Epidote	4.2	4.0	1.9	1.6	2.6	1.4	4.4	1.8	3.2	3.2	6.8	1.9	2.4
Zoisite	0.6	0.6	0.6	0.2	0.6	-	1.0	-	0.7	0.7	2.2	-	-
Staurolite	0.4	0.4	1.0	1.8	0.4	2.6	0.6	0.8	1.8	0.4	0.8	0.8	1.4
Disthen	-	0.2	-	0.2	0.2	-	-	-	-	-	-	-	0.1
Andalusite	-	-	-	-	-	-	-	-	-	-	-	-	-
Sillimanite	-	-	-	0.4	-	0.6	-	-	0.2	-	-	-	-
Apatite	0.4	2.0	5.1	3.2	1.0	3.6	0.6	2.8	3.8	1.9	1.6	2.1	1.0
Rutile	1.6	1.4	1.6	1.4	1.2	1.0	1.2	0.6	1.0	1.2	0.8	0.8	1.6
Sphene	1.0	0.6	-	0.2	1.0	0.4	1.0	0.2	0.7	1.0	1.2	0.2	0.4
Brookite, anatase	0.6	0.4	0.5	0.2	1.2	0.4	0.4	-	-	0.7	0.4	0.3	-
Dolomite	1.4	-	-	-	0.4	0.2	0.8	-	0.2	0.6	-	-	-
Skeletal fragments	-	-	-	-	1.0	0.2	-	-	0.3	-	-	0.2	-
Content of heavy fraction	0.07	0.42	4.02	0.22	0.06	16.7	0.05	0.12	2.98	0.31	0.08	4.41	0.05

Analyst M. Viiding

In particular detail we have studied the mineral composition of sediments in the northern part of L. Peipsi *s.s.* (see Ch. 4) and in L. Lämmijärv (Raukas *et al.*1988). The content of quartz in fine sand fraction of L. Lämmijärv is lowest (64–78%) in the central part of the lake where, in an ancient valley, fine-grained sediments occur (Fig. 62a). At the same time, the highest concentrations of feldspars (up to 25.2%) and micas (muscovite up to 21.7 and biotite up to 40.5%) have been recorded in the central part of the lake (Fig. 62b and d). The proportion of feldspars is rather high near the river mouths, *e.g.* 20.7% in the Võhandu mouth, the average being between 10–15%. Among the minerals of heavy subfraction, magnetite-ilmenite concentration is highest (up to 50%, mainly 20–30%) in the peripheral part of the lake. In the central part, it is much lower, sometimes less than 5% (Fig. 63a). Leucoxene (mainly 1.5–3.5%, occasionally up to 8%), apatite (up to 8.4%) and garnets (15–30%, max. up to 48%) are more common in the northern, and zircone (3–10, max. 15.4%) in the southern part of the lake, due to the influence of the Devonian bedrock (Viiding 1968) and close-lying tills (Raukas 1978). Particularly high concentrations of zircone, tourmaline, garnets, ilmenite-magnetite and other heavy minerals have been recorded near Mehikoorma and Pnevo (Figs. 63, 64). This is evidently due to the hydrological conditions; in this part the lake is at its narrowest and provides the best conditions for intense separation of heavy minerals. High concentrations of tourmaline are also found to the south from the Island of Piirissaar (up to 7%) and in the southern part of the lake near Raigla and Osotno (Fig. 64b).

The results of X-ray analysis have shown that fine-pelitic (less than 0.001 mm) fraction is everywhere dominated by clay minerals (90–95%), such as illite, chlorites, kaolinite, mixed-layered smectite-illite and smectite-chlorite (Table 15). Some terrigenous minerals, including quartz (in all samples), feldspars (in most of samples), calcite (40% of samples) and dolomite (25% of samples), with less siderite, hematite, pyrite and goethite were identified in fine-pelitic fraction. Siderite and goethite were found only in the bottom deposits of L. Pihkva. Vermiculite was not identified, however, in glaciolacustrine clays of southern Estonia it reaches 20% (Saarse & Utsal 1974). In most cases, illite makes up 55–75%, while the proportion of kaolinite (in most cases 5–15%), chlorites (up to 22%) and smectites (5–20%, in some places in L. Pihkva and L. Lämmijärv up to 40%) is, as a rule, modest. Among the chlorites, ferrugineous varieties

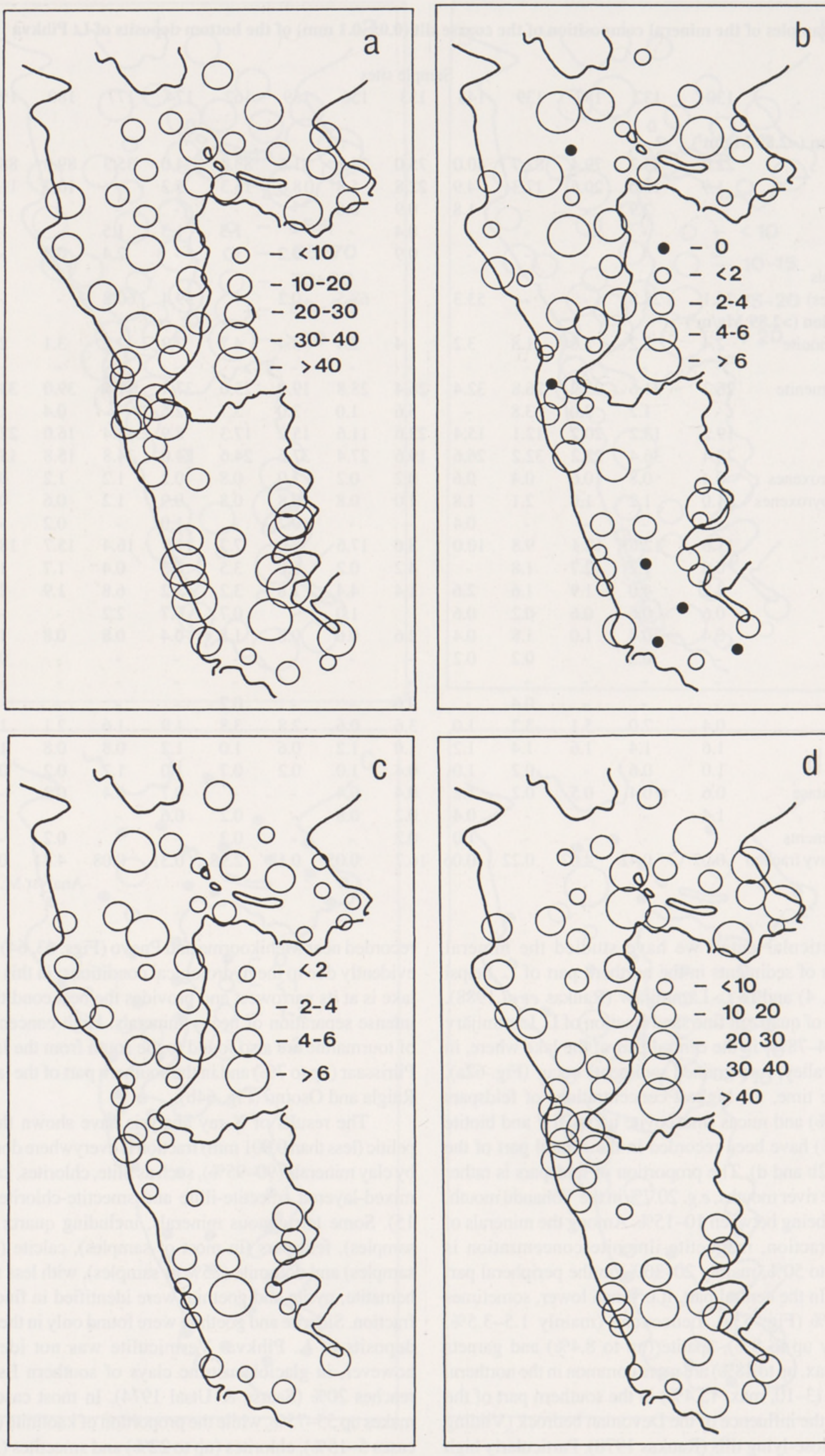


Fig. 63. Distribution of heavy (>2.89 Mg/m^3) minerals in fine sand fraction (0.1–0.25 mm) in the bottom deposits of L. Lämmijärv (after Raukas *et al.* 1988): a - magnetite and ilmenite; b - leucoxene; c - apatite; d - garnet.

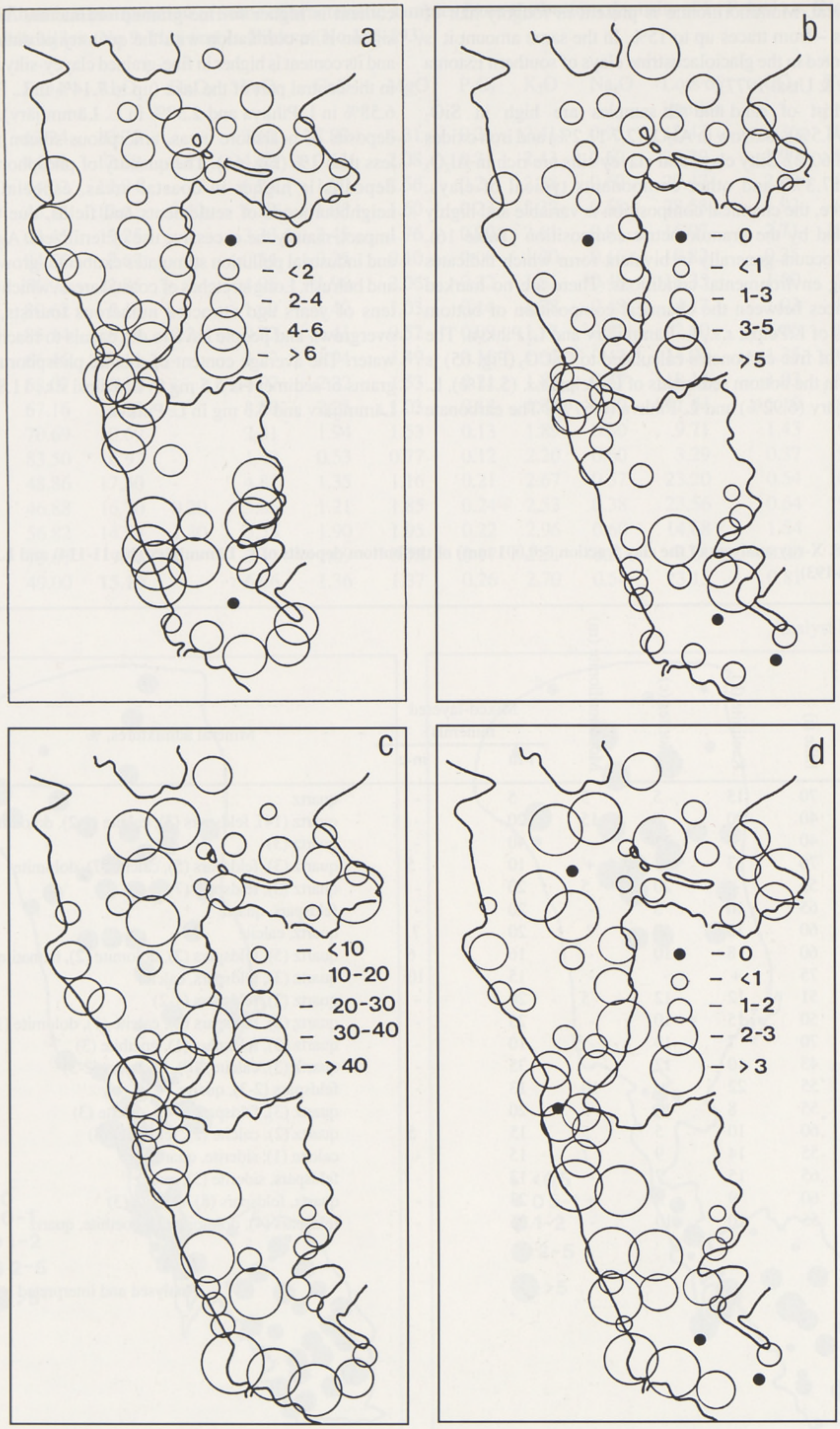


Fig. 64. Distribution of heavy (>2.89 Mg/m^3) minerals in fine sand fraction (0.1–0.25 mm) in the bottom deposits of L. Lämmijärv (after Raukas *et al.* 1988): a - zircon; b - tourmaline; c - amphiboles; d - pyroxenes.

are typical. Montmorillonite is present in roughly half of samples – from traces up to 15%. In the same amount it is represented in the glaciolacustrine clays of southern Estonia (Saarse & Utsal 1977).

Most of sand and silt samples are high in SiO_2 (66.5–94.5%), and low in Al_2O_3 (2.7 11.2%) and iron oxides (0.34–3.60%). Silty clays and clayey silts are rich in Al_2O_3 (9.25–17.3%) and other components typical of clays. Therefore, the chemical composition is variable and highly controlled by the granulometric composition (Table 16). Ferrum occurs generally in bivalent form which indicates reducing environmental conditions. There are no marked differences between the chemical composition of bottom deposits of L.Peipsi *s.s.*, L. Lämmijärv and L. Pihkva. The content of free carbonates calculated to $CaCO_3$ (Fig. 65) is similar in the bottom sediments of L. Peipsi *s.s.* (5.11%), L. Lämmijärv (6.92%) and L. Pihkva (6.67%). The carbonate content is higher in fine-grained sediments. Amorphous silicon is in correlation with the quantity of diatom algae, and its content is higher in fine-grained clayey-silty sediments in the central part of the lake (up to 7.14% in L. Peipsi *s.s.*, 6.58% in L.Pihkva and 2.58% in L. Lämmijärv). In sandy deposits of nearshore areas, amorphous silicon makes up less than 1% (Fig. 66). The quantity of phosphorus in lake deposits is higher in coastal areas, especially in the neighbourhoods of settlements and fields, due to human impact, mainly the excessive use of fertilizers. Agricultural and industrial pollution stimulates exuberant growth of reed and bulrush. Long stretches of coastal areas, which still some tens of years ago attracted numerous tourists, are now overgrown and people have to dig canals to reach the open water. The average content of mobile phosphorus per 100 grams of sediment is 5.5 mg in L. Peipsi *s.s.*, 11.8 mg in L. Lämmijärv and 6.3 mg in L. Pihkva.

Table 15. X-ray analyses of the clay fraction (<0,001 mm) of the bottom deposits of L. Lämmijärv (No 11-114) and L. Pihkva (No 130-193)

Sample sites	Illite (i)	Kaolinite (k)	Chlorites (c)	Montmorillonite (m)	Mixed-layered minerals		Mineral admixtures, %
					i-m	m-c	
11	70	15	5	5	5	-	quartz
15	40	20	5	15	20	-	quartz (14), feldspars (3), calcite (1-2), dolomite (1)
19	40	5	15	-	40	-	quartz (3)
25	70	3	12	+	10	5	quartz (3), feldspars (5), calcite (7), dolomite
40	55	-	20	5	20	-	quartz (3), feldspars (7)
49	65	10	5	-	20	-	feldspars, quartz
77	60	10	10	?	20	?	quartz, calcite
96	60	8	10	-	16	6	quartz (5), feldspars (2), dolomite (2), hematite (3)
110	75	+	-	-	15	10	quartz (3), feldspars, calcite
112	51	12	12	5	20	-	quartz (2), feldspars (1-2)
114	50	15	10	-	25	-	quartz (5), feldspars (2), calcite (1), dolomite (1)
130	70	7	13	-	10	-	quartz (3), feldspars (1), goethite (3)
132	43	10	12	-	35	-	quartz (3), calcite (10), feldspars (2-3)
145	55	22	+	10	13	-	feldspars (2-3), quartz
156	55	8	12	5	20	-	quartz (3), feldspars (1-2), calcite (3)
165	60	10	5	5	15	5	quartz (2), calcite (2), feldspars (3)
174	55	14	9	7	15	-	calcite (1), siderite, quartz
177	65	15	7	-	13	-	feldspars, siderite (3), quartz
191	60	9	8	+	23	-	quartz, feldspars (8), siderite (3)
193	55	10	10	-	25	-	feldspars (4), dolomite (2), goethite, quartz

Analysed and interpreted by E. Pirrus

Table 16. Chemical composition of some characteristic samples from the bottom deposits of L. Peipsi *s.s.* (No 218-290), L. Lämmijärv (No 9-111) and L. Pihkva (No 146-193)

Sample sites	SiO_2	Al_2O_3	Fe_2O_3	FeO	CaO	MgO	P_2O_5	K_2O	Na_2O	Loss of ignition	CO_2	Total
218	67.24	10.75	-	3.60	2.09	1.31	0.21	2.41	0.93	12.43	2.36	100.97
250	44.96	12.69	0.27	4.82	2.11	2.08	0.21	2.43	0.60	29.29	1.99	99.46
265	44.02	12.94	-	6.24	1.90	1.56	0.24	2.28	0.60	29.47	2.00	99.25
274	46.14	12.62	0.40	5.22	1.95	1.60	0.22	2.32	0.58	28.58	1.92	99.63
288	63.45	9.25	-	3.33	2.45	1.76	0.20	2.61	0.91	16.07	2.71	100.03
290	94.55	2.67	-	0.34	0.25	0.10	0.08	1.00	0.16	0.82	0.35	99.97
9	56.91	12.85	-	4.95	1.96	2.06	0.22	1.81	0.48	18.35	1.50	99.59
14	80.62	8.17	-	1.69	1.40	1.03	0.14	1.78	0.48	4.17	1.03	99.48
43	85.14	6.78	0.21	0.72	1.41	0.52	0.09	1.33	0.52	2.60	1.06	99.32
93	66.46	11.11	-	3.57	1.94	1.49	0.16	1.65	0.47	14.05	1.35	100.90
111	58.07	13.38	-	4.07	2.22	1.53	0.21	1.88	0.47	18.14	1.03	99.97
146	67.16	11.18	-	3.50	2.25	1.03	0.18	1.87	0.50	11.64	1.10	99.31
152	70.69	10.69	-	2.31	1.94	1.53	0.13	1.89	0.50	9.71	1.43	99.39
158	83.50	6.97	-	1.42	0.53	0.77	0.12	2.20	0.60	3.29	0.37	99.40
163	48.86	17.30	-	4.81	1.35	1.16	0.21	2.67	0.57	23.20	0.54	100.13
165	46.88	16.70	0.20	5.85	1.21	1.85	0.24	2.53	0.38	23.56	0.64	99.40
177	56.82	14.94	1.80	3.55	1.90	1.95	0.22	2.96	0.60	14.68	1.54	99.42
182	73.67	9.36	-	2.42	1.09	0.78	0.17	2.21	0.50	8.81	0.23	99.01
193	49.00	15.10	-	6.06	1.36	1.37	0.26	2.70	0.54	23.15	0.81	99.55

Analyst L. Säga

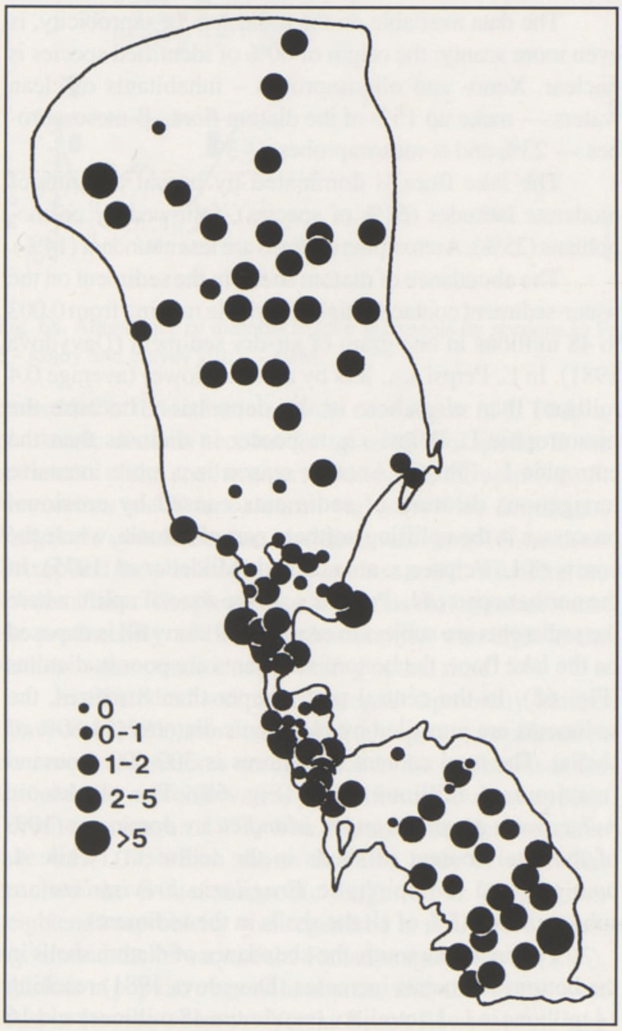


Fig. 65. Content of free carbonates in the bottom deposits of L. Peipsi, % (after Raukas 1981, with complements).

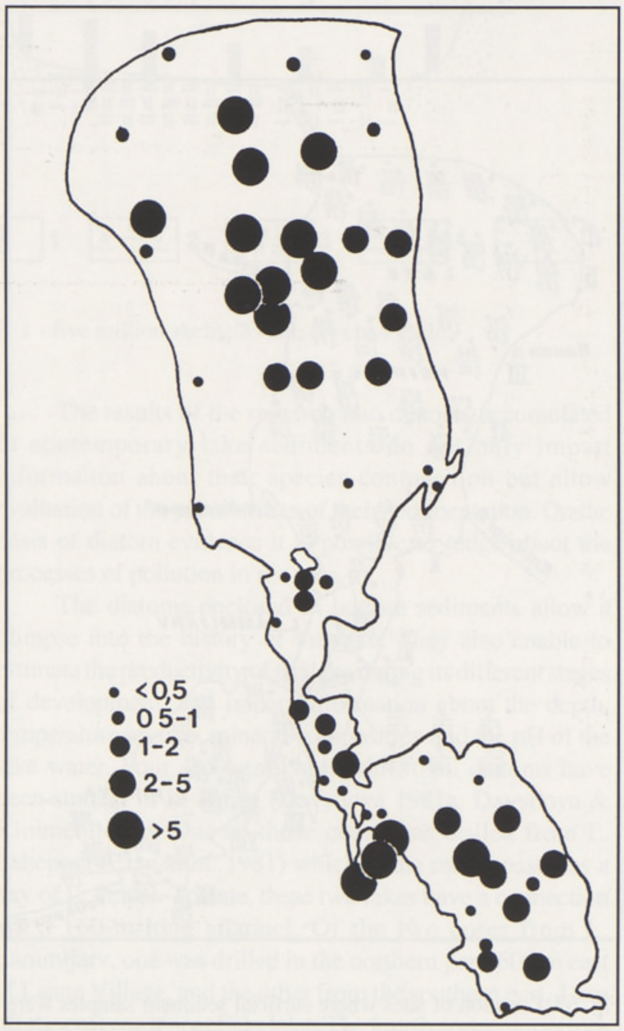


Fig. 66. Content of amorphous silicon in the bottom deposits of L. Peipsi, % (after Raukas 1981, with complements).

3.2.4. FLORA IN THE SURFICIAL LAYER

3.2.4.1. DIATOMS

Diatoms are a leading group of algae in both the planktonic and benthic communities of the lake, where their species composition is diverse (Laugaste 1966, Yastremskij 1977). Diatoms were studied in both the surficial layer and bottom deposits (Davydova 1981, 1981a). Of the algal flora in the lake, diatoms have the highest number of species and taxa (357). During the studies, pelogene samples with a thickness of up to 1 cm were taken from standard profiles at 65 stations (Fig. 67) by means of an Ekman grip-scoop. The samples were treated using quantitative methods (Davydova 1985). The aim was to study the species composition and quantitative accumulation of diatoms in bottom sediments, but also the peculiarities of their sedimentation in L. Peipsi *s.s.* and L. Pihkva. Indications of the zones of anthropogenic pollution in this limnologically complex body of water were studied on the basis of monitoring data.

Of the 220 taxa of diatoms identified in the surface layer of bottom deposits, 17% species were planktonic, dominated by *Aulacoseira ambigua* (Ehr.) Simonsen, *A. granulata* (Grun.) Simonsen, *A. islandica* (O. Mull.) Simonsen, *Stephanodiscus rotula* (Kütz.) Hendey, and *S. minutulus* (Kütz.) Cleve & Möller (Davydova 1981). Benthic diatoms displayed a diverse species composition, the most widespread species being the epiphytes *Fragila leptostauron* var. *martyi* (Herib.) Lange-Bertolet, *F. brevistriata* Grunow and *F. construens* (Ehr.) Grunow with varieties, and the bottom species *Amphora pediculus* (Kütz.) Grunow and *Gyrosigma attenuatum* (Kütz.) Cleve. Bottom diatoms make up 51% of the total number of species, while epiphytic diatoms account for 32%. This can be explained with the weak development of macrophytes in the lake. The diatom flora is prevailed by typical fresh-water diatoms; indifferent taxa account for 81%, halophobes — inhabitants of acid waters with a low content of dissolved mineral salts — are rare enough (4%) in the lake where the total dissolved salt content is 200 mg l^{-1}. Halophytes account for only 15% of the total species diversity. Of the diatoms identified in the bottom sediments, 64% are adapted to grow in water basins with different active reaction of the medium. Alkaliphilic and alkalibiontic diatoms, capable of inhabiting alkaline waters in which the pH value is 8 and even higher during the vegetation period, make up two thirds of the taxa. This group includes *A. ambigua, A. granulata, Fragilaria brevistriata, F. construens* and *F. leptostauron* var. *martyi* (Herib.). 16% of species are inhabitants of neutral waters, and 3% prefer acid swamp water.

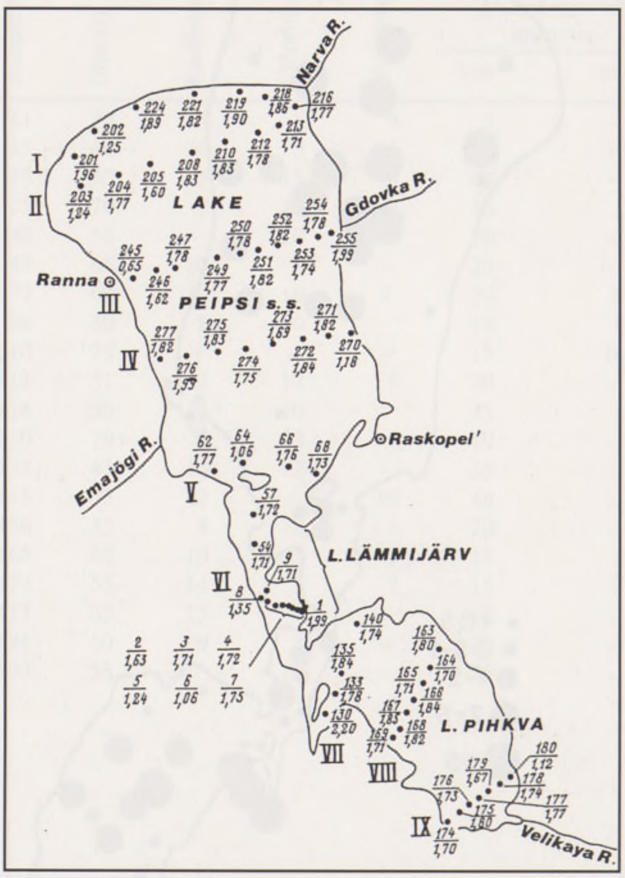


Fig. 67. Location of sites where surficial sediment samples were taken for diatom analysis. Roman numerals - sections, Arabic numerals - sampling sites (upper) and integral saprobicity indices (lower).

The data available on the indicators of saprobicity, is even more scanty: the origin of 60% of identified species is unclear. Xeno- and oligosaprobes – inhabitants of clean waters — make up 15% of the diatom flora, β-mesosaprobes — 23% and α-mesosaprobes — 3%.

The lake flora is dominated by boreal diatoms of moderate latitudes (61% of species), followed by cosmopolitans (25%). Arctoalpine diatoms are less abundant (14%).

The abundance of diatom shells in the sediment on the water-sediment contact is highly variable ranging from 0.003 to 48 millions in one gram of air-dry sediment (Davydova 1981). In L. Peipsi *s.s.*, it is by an order lower (average 0.4 million) than elsewhere in the depression, because the mesotrophic L. Peipsi *s.s.* is poorer in diatoms than the eutrophic L. Pihkva. Another reason is a more intensive terrigenous dilution of sediments caused by erosional processes in the uplifting northern part of Estonia, where the coasts of L. Peipsi *s.s.* are situated (Miidel *et al.* 1975). In the northern part of L. Peipsi *s.s.*, in the zone of uplift, where the sediments are subject to erosion and stony till is exposed on the lake floor, the bottom sediments are poor in diatoms (Fig. 68). In the central part, deeper than 8 metres, the sediments are prevailed by planktonic diatoms (54-80% of shells). The total content of diatoms is 300-600 thousand (maximum 1 million) shells (Fig. 68). The planktonic *Aulacoseira granulata* and *A. islandica* are dominants (10% of the total content of shells in the sediment), while *A. ambigua* and the epiphytic *Fragilaria brevistriata* are subdominants (5% of all the shells in the sediment).

Farther to the south, the abundance of diatom shells in the bottom sediments increases (Davydova 1981) reaching 14 millions in L. Lämmijärv (maximum 48 millions) and 16 millions in L. Pihkva (maximum 30 millions)(Fig.68). Epiphytes are dominants and form 76% of the shells

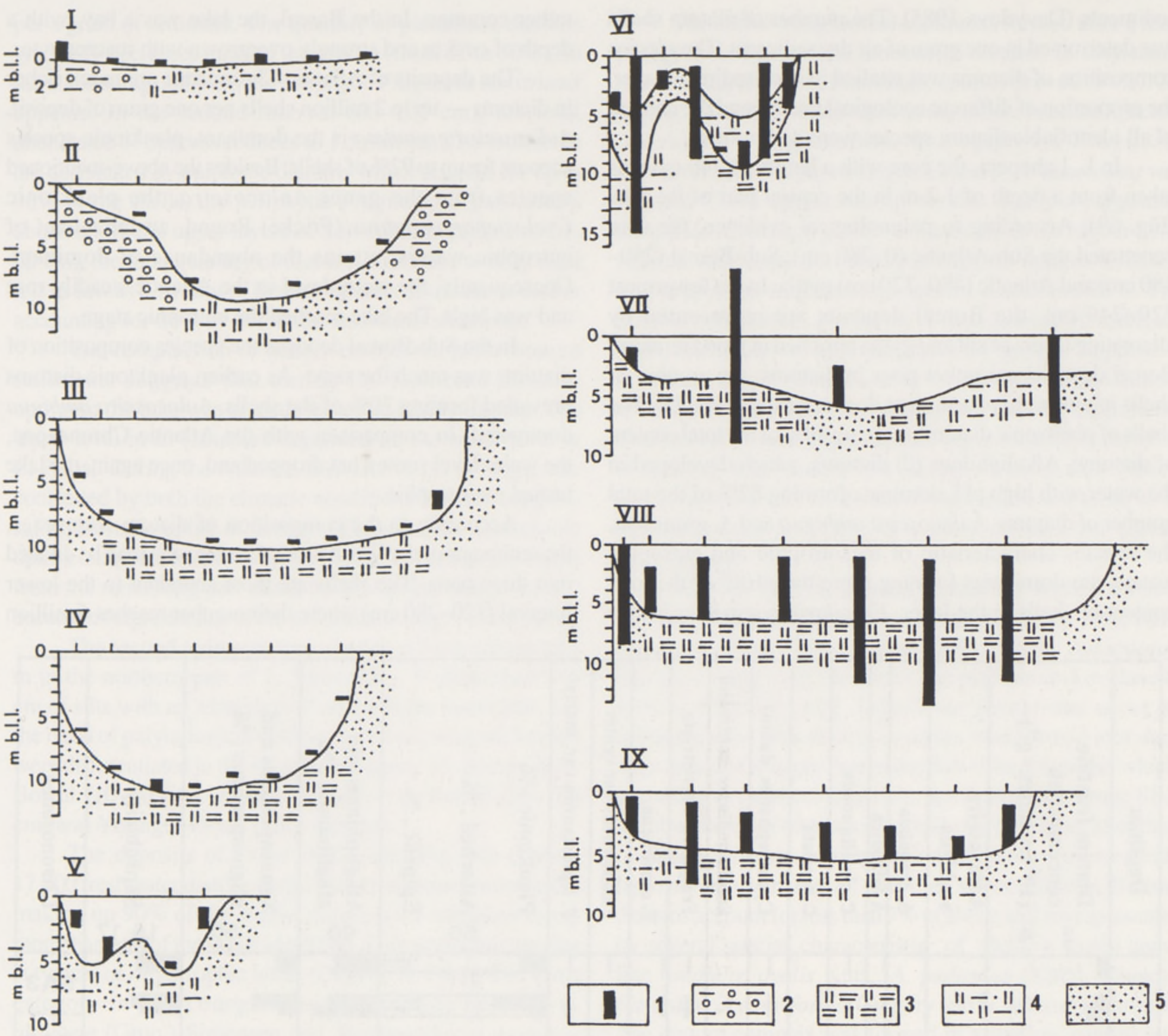


Fig. 68. Abundance of diatoms in lake sediments by sections in Fig 67: 1 - five million shells; 2 - till; 3 - clayey silt; 6 - sandy and clayey silt; 7 - sand.

embedded in the sediments. In the central part of L. Pihkva, planktonic diatoms account for up to 49% of shells. In the sediments of the southern part of this area, the epiphytes *F. brevistriata* and *F. lepostauron* var. *martyi* are dominants. *Fragilaria construens* var. *binodis* et var. *venter*, *F. heidenii* Ostrup and the planktonic *A. granulata* and *A. islandica* are subdominants. The prevalence of benthic diatoms shows that the productivity of benthic communities was high in the shallow water conditions prevailing at that time.

In all studied sediments, integral saprobicity indices (Davydova 1985), were calculated taking into consideration the amount of diatoms accumulated in the sediments at the site and the saprobic valency of different diatom species (Unifitsirovannye... 1975). The average index for the whole lake was 1.77, which shows that this body of water has reached the β-mesosaprobic stage. In L. Peipsi *s.s.*, heightened saprobicity was registered in near-shore zones, *e.g.* around the estuary of the Rannapungerja River (1.96), which is a popular recreational area, and also at the west coast in the area where the Värska Brook flows into the lake (2.20), and where the Värska Settlement and a balnaeological resort are situated.

The results of the research into diatoms accumulated in contemporary lake sediments do not only impart information about their species composition but allow evaluation of the peculiarities of their sedimentation. On the basis of diatom evidence it is possible to judge about the processes of pollution in the lake.

The diatoms enclosed in bottom sediments allow a glimpse into the history of the lake. They also enable to estimate the productivity of the lake during its different stages of development, and impart information about the depth, temperature regime, mineral composition and the pH of the lake water. Four sediment cores with fossil diatoms have been studied in L. Peipsi (Davydova 1981a, Davydova & Kimmel 1991). One of those cores was drilled from L. Lahepera (Paap *et al.* 1981) which in the recent past was a bay of L. Peipsi. To date, these two lakes have a connection via a 160-m-long channel. Of the two cores from L. Lämmijärv, one was drilled in the northern part, 600 m east of Laane Village, and the other from the southern part, 1 km off the east coast near Salu Islet. The fourth core comes from L. Pihkva, in the area east of Kamenka Island. Diatoms were studied by means of the quantitative analysis of bottom sediments (Davydova 1985). The number of diatom shells was determined in one gram of air dry sediment. The species composition of diatoms was studied in each sediment layer, the proportion of different ecological groups and the content of all identifiable diatom species were determined.

In L. Lahepera, the core with a length of 740 cm was taken from a depth of 1.2 m in the central part of the lake (Fig. 69). According to palynological evidence, the core penetrated the Sub-Atlantic (0–280 cm), Sub-Boreal (280–480 cm) and Atlantic (480–720 cm) gyttja. In the lowermost 720–740 cm, the Boreal deposits are represented by alternating layers of silt and gyttja enriched in wood remains. Boreal deposits are rather poor in diatoms; the number of shells in one gram of sediment does not exceed 20,000. The shells of planktonic diatoms make up 60% of the total content of diatoms. Alkaliphilous (?) diatoms, which developed in the water with high pH, dominate forming 82% of the total number of diatoms. *Aulacoseira ambigua* and *A. granulata*, the species characteristic of mesotrophic and eutrophic waters, are dominants forming more than 10% of the total content of shells in the layer. *Fragilaria construens* is also rather common. In the Boreal, the lake was a bay, with a depth of *ca* 5 m and strongly overgrown with macrophytes.

The deposits of Atlantic Chronozone are much richer in diatoms — up to 2 million shells per one gram of deposit. *Aulacoseira granulata* is the dominant, planktonic species account for up to 92% of shells. Besides the above-mentioned species from the genus *Aulacoseira*, the planktonic *Cyclostepanos dubius* (Fricke) Round, an inhabitant of eutrophic waters, attains the abundance of dominant. Consequently, the water level in the Atlantic steadily rose and was high. The lake reached the eutrophic stage.

In the Sub-Boreal deposits the species composition of diatoms was much the same. As earlier, planktonic diatoms prevailed forming 70% of the shells. *Aulacoseira ambigua* dominated. In comparison with the Atlantic Chronozone, the water level somewhat dropped and, once again, the lake turned mesotrophic.

According to the composition of diatom complexes, the sediments of the Sub-Atlantic Chronozone may be divided into three parts. The shells are most abundant in the lower interval (120–280 cm) where their number reaches 6 million

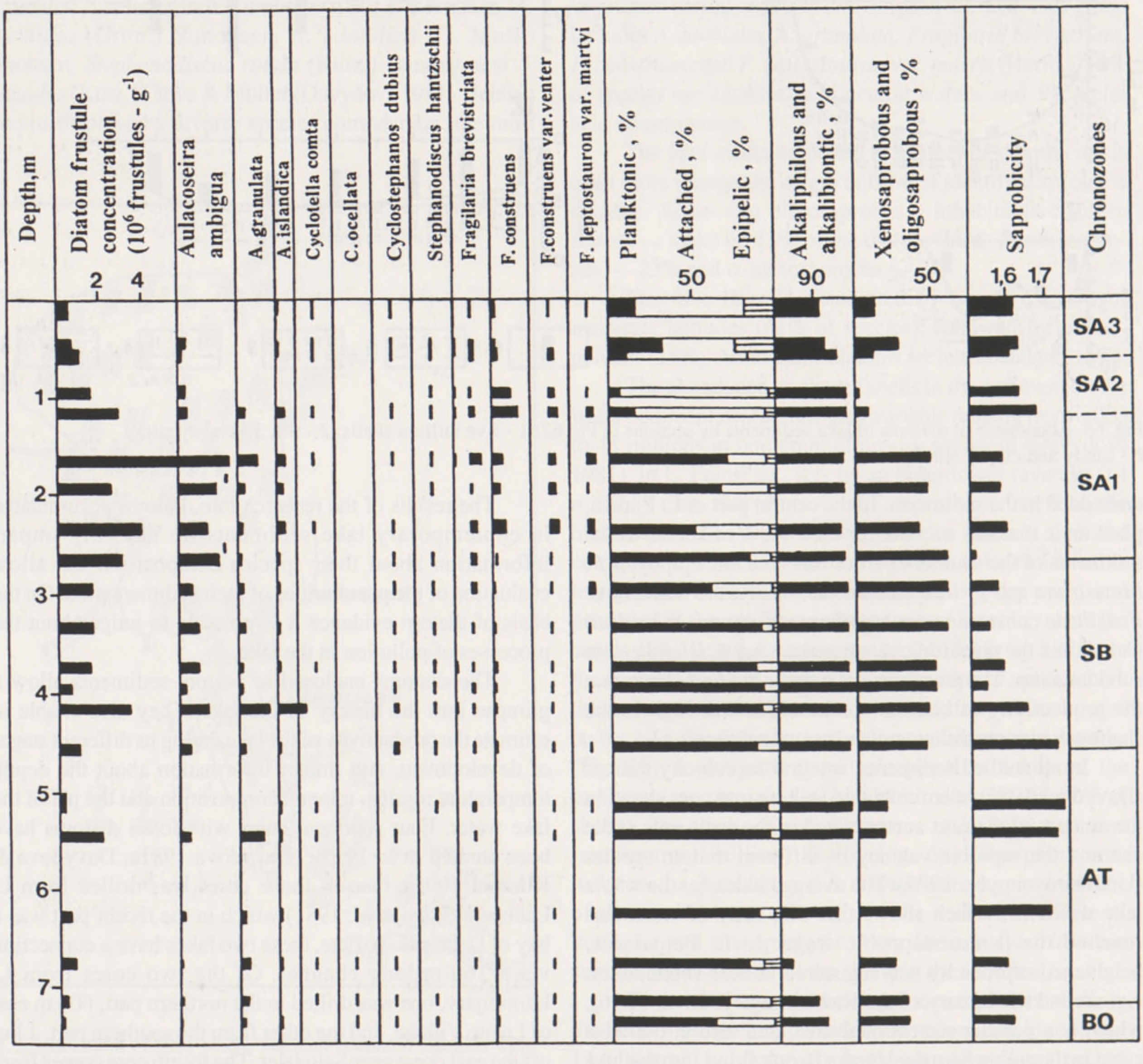


Fig. 69. Diatoms in L. Lahepera sediment core: 1 - planktonic; 2 - attached; 3 - epipelic.

per a gram of sediment. The quantity of planktonic diatoms gradually decreases upward the profile (from 81 to 50%). In the dominating complex, the periphytic *Fragilaria construens* appears. In the second interval (60–120 cm), the total abundance of diatoms reduces to 1.5 million. The sediments are dominated by epiphytic diatoms which account for up to 95% of the shells, the periphytic *Fragilaria* is particularly abundant. In the upper layers of Sub-Atlantic sediments (0–60 cm), the total quantity of diatoms continues to drop until it is as low as 0.5 million. Benthic species dominate as earlier accounting for 88% of shells, the lake turns shallower.

The composition of diatom complexes in the bottom sediments suggests that during the Holocene climatic optimum the lake was at its deepest and highly trophic. The water-level in L. Lahepera depended on the water-level in L. Peipsi. During the Atlantic, the water-level in Peipsi was controlled by both the climatic conditions and the compensating isostatic uplift of northern Estonia. The neotectonic uplift continued, and in the Sub-Atlantic L. Lahepera isolated from the remaining part of the basin, turned shallower, and became overgrown due to the sediments filling the lake basin.

The second sediment core was taken from a depth of 4 m in the northern part of L. Lämmijärv. It penetrated 135 cm of silts with an admixture of wood in the lower part. On the basis of palynological evidence, the following sediments were differentiated in the core: Sub-Atlantic (0–10 cm), Sub-Boreal (10–30 cm), Atlantic (30–55 cm), Boreal (55–100 cm) and Younger Dryas (100–135 cm).

The deposits of Older Dryas age (Fig. 70) contain 17,200 freshwater diatom shells. Planktonic species dominate making up 90% of the shells. *Aulacoseira islandica* forms more than half of the shells identified. This planktonic species is typical of oligotrophic lakes (Davydova 1985). The same category in the Younger Dryas sediments includes *A. alpigena* (Grun.) Simonsen and *Stephanodiscus niagarae* Ehrenberg. The planktonic *Aulacoseira ambigua* and *Tabellaria fenestrata* (Lyngl.) Kütz. occur as subdominants. The complex of diatoms involves phytoplankton with a variable species composition and poorly developed benthic diatom communities. It was formed in a deep and slightly meso-trophic cold fresh-water basin.

The above-lying Boreal sediments formed after a long stratigraphic hiatus. The amount of diatoms in silty muds reaches 32,000 shells. Planktonic species prevail (94%). As before, *Aulacoseira islandica* is the most abundant species, followed by *A. granulata*. The epiphytes *Fragilaria brevistriata*, *Achnanthes exigua* Grunow, *A. lanceolata* var. *elliptica* Cleve appear among benthic diatoms suggesting the presence of macrophytes in the basin. In the higher lying Atlantic sediments the number of diatoms reaches 150,000. The proportion of planktonic species reduces to 76%. The dominants include *A. granulata, A. islandica* and *A. ambigua.* The contribution of epiphytes from the genus *Fragilaria – F. brevistriata, F. construens et* var. *binodis* (Ehr.) Grunow *et* var. *venter* (Ehr.) Grunow increases indicating a significant distribution of submerged aquatic vegetation. This complex of diatoms was formed in a shallower basin than earlier. Environmental conditions there were much the same as in the present-day L. Peipsi, at some distance from coastal macrophytes.

The Sub-Boreal deposits yielded the highest amount of shells for the whole profile – up to 0.6 million. Planktonic species dominate forming up to 95% of shells. *Aulacoseira islandica* is the main dominant, the planktonic *Asterionella formosa* Hassall and *Tabellaria fenestrata* serve as subdominants. The diatom complex was formed in a deep open lake. In the upper part of the Sub-Boreal deposits which, according to palynological data, is referred to the zone SB2, the abundance of diatoms reduces to 0.25 million; as earlier, planktonic diatoms dominate. *Fragilaria brevistriata* and *F. leptostauron* var. *martyi* serve as subdominants. Bottom diatoms account for less than 7% of shells and are represented by several species characteristic of shallow sandy areas, like *Amphora ovalis* Kütz., *A. pediculus* (Kütz). *Grunow, Navicula cocconeiformis* Gregory and *N. jentzschii* Grunow. The diatom complex was formed in a shallow coastal area of an open lake.

The Sub-Atlantic sediments have been studied only in the topmost layer of deposits where the content of diatoms reaches 29 million shells. Epiphytes (54% of shells) dominate, followed by planktonic (30%) and bottom species (16%). Among the dominants, the epiphytic *Fragilaria*

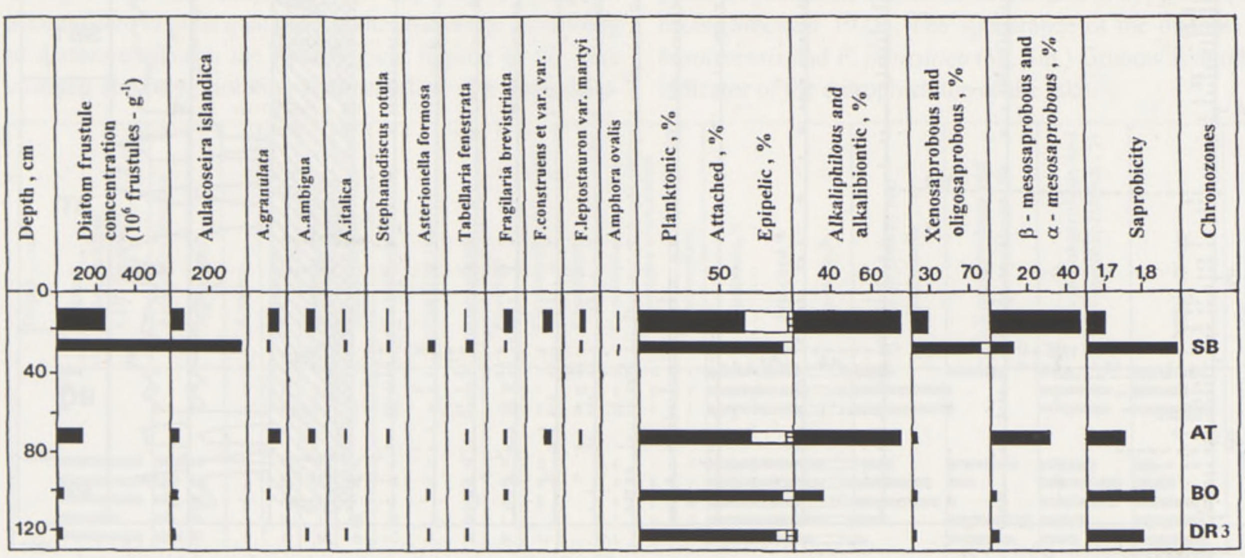


Fig. 70. Diatoms in the sediment core from the northern L. Lämmijärv (east of Laane Village). For legend see Fig. 69.

leptostauron var. *martyi* is most abundant, followed by the planktonic *Aulacoseira islandica, A. granulata* and the species of genus *Fragilaria*.

The results of diatom analysis suggest that during the Younger Dryas the basin of Lake Lämmijärv was part of a deep fresh-water basin. The studied bottom deposits impart information about its pelagial which was rather far away from the coast. The next sedimentation period started after a long stratigraphic hiatus when the present-day lake basin was dry land. In the Boreal, the basin was once again filled by the waters of a large and relatively deep lake, which is still there. During the climatic optimum, the lake level stayed high. The lake turned eutrophic with a well-developed plankton and benthos. During the first half of the Sub-Boreal, the lake level continuously rose and the productivity increased. At the end of the Sub-Boreal and at the beginning of the Sub-Atlantic, the lake became shallower. As never before, the benthic communities gained in importance among the near-shore vegetation. The percentage of alkaliphilous diatoms and alkalibionts increased which, as is known (Granberg 1972), is an indicator of the increasing trophicity in the lake.

The third sediment core was obtained from a depth of 7.5 m in an ancient valley in the southern part of L. Lämmijärv *ca* 1 km from the east coast close to Salu Islet. According to palynological evidence, the core penetrates the whole Holocene complex. In the Pre-Boreal deposits, the content of diatoms does not exceed 1000 shells. The shells of planktonic and benthic diatoms were present almost in the same amounts.

The dominants include the planktonic species *Aulacoseira islandica* and the epiphyte *Fragilaria* (Fig. 71). In the lower part, the Boreal deposits are dominated by planktonic diatoms. Further, the dominants are represented by the epiphyte *Fragilaria*, the bottom species and the planktonic species *Aulacoseira ambigua*. In the upper part of the Boreal, the abundance of diatoms increases, but benthic species dominate as before. *Navicula scutelloides* W. Smith appears among sub-dominants. The role of alkaliphilous diatoms and alkalibionts increases showing that the body of water has reached the eutrophic stage in its development.

Benthic diatoms continue to dominate in the lower part of the Atlantic sedimentary complex. In this part of the core,

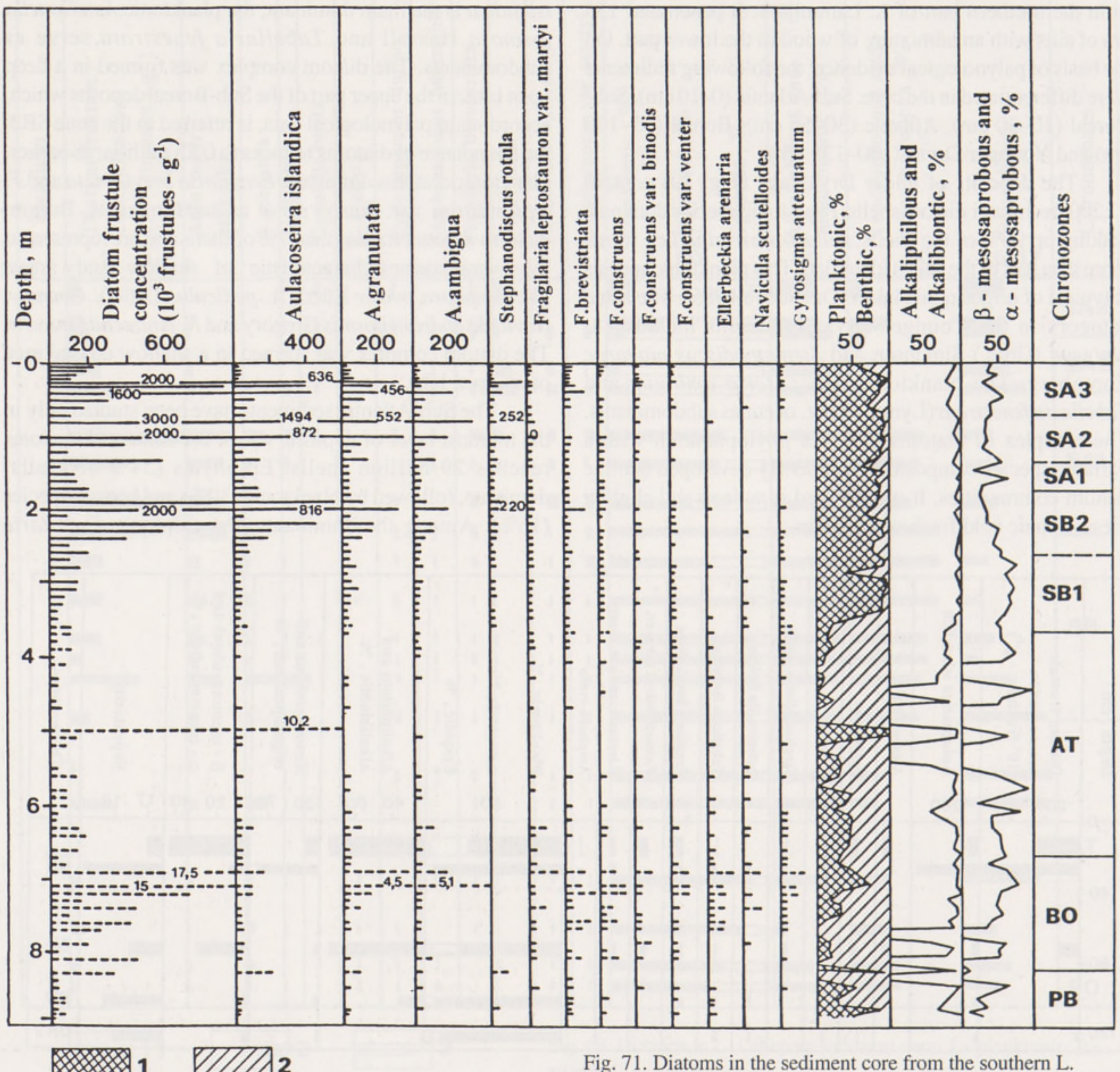


Fig. 71. Diatoms in the sediment core from the southern L. Lämmijärv (near Salu Islet): 1 - planktonic; 2 - benthic

the number of species is at its highest throughout the sedimentation period. In a short section of Atlantic sediments the planktonic *A. islandica* and *Cyclotella comta* (Ehr.) Kützing, higher in the profile they are replaced by benthic diatoms. The proportion of acidophilic diatoms increases up to 23%.

In the Sub-Boreal deposits, the composition of the diatom complex changes significantly: planktonic species become most abundant, forming about 90% of the total number of shells. The earlier dominating species of *Aulacoseira* are supplemented by *A. italica* (Ehr.) Simonsen. Representatives of the planktonic species *Cyclotella bodanica* Grunow, *C. comta, Cyclostephanos dubius, Stephanodiscus rotula* have been identified. Benthic diatoms are low in number, but diverse in the species composition.

In the Sub-Atlantic deposits, the abundance of diatoms continues to increase reaching 3 million shells which is the highest value for the whole profile. Planktonic diatoms account for 96% of the total content of shells. The species of *Aulacoseira* serve as dominants. The planktonic *Fragilaria capucina* Desmazieres *et* var. *mesolepta* Rabenhorst and *Stephanodiscus hantzschii* Grunow, characteristic of eutrophic bodies of water, occur in great numbers.

In the beginning of the Sub-Atlantic, both the abundance of diatoms and the proportion of benthic species in diatom complexes increased. The earlier identified dominants were supplemented by *Stephanodiscus rotula.* An inhabitant of eutrophic waters – *Fragilaria berolinensis* (Lemm.) Lange-Bertalot – makes its first appearance. In the middle of the Sub-Atlantic, the abundance of diatoms in the sediments was at its highest and reached 3 million shells. Planktonic diatoms prevailed. The appearance of *Asterionella formosa* and *Fragilaria crotonensis* Kitton suggests a further growth in eutrophication. In the higher-lying sediments, the abundance of diatoms decreases abruptly, the epiphyte *Fragilaria brevistriata* is once again among the dominants. In the uppermost 5 cm, epiphytes dominate.

The results of diatom analysis indicate that in the Pre-Boreal and Boreal, *i.e.* during the first stage of the lake development, sediments accumulated in a deep lake. In the Boreal and Atlantic, the water basin was shallow and overgrown with macrophytes. Epiphytic diatoms accumulated in great quantities in the sediments. According to diatom evidence, the hydrological regime of the lake changed abruptly, initiating sedimentation in a large deep-water basin with a concomitant increase in the productivity of diatom communities. During the first half of the Sub-Atlantic, *i.e.* from approximately 500 BC up to the 11th century AD, the productivity reached its maximum. The next stage in the evolution of the lake associates with the Little Ice Age (12th–17th centuries). During that stage the abundance of diatoms decreased by an order. The steadily improving climatic conditions promoted a further growth in the total abundance of diatoms and the proportion of the plankton increased. In the 20th century, the rate of sedimentation increased. This was mainly due to the ploughing of the territory and formation of cultural landscapes in the drainage area.

The fourth sediment core was taken from a depth of 5.5 m in L. Pihkva. According to palynological evidence, this 90-cm-long core comprises deposits of the most recent stage of the Sub-Atlantic. Based on diatom complexes, two intervals are distinguished (Fig. 72). In the lower interval (40–90 cm), the number of shells reaches 3–5 millions in one gram of sediment. The planktonic diatoms, forming up to 85% of shells, are prevailed by *Aulacoseira islandica* (up to 40% of shells). The abundance of two other dominants – *A. granulata* and *A. ambigua* – increases significantly upwards. Abundant benthic diatoms are represented by the species of epiphytic *Fragilaria.* In the upper interval (0–40cm), the abundance of diatoms reduces and that of benthic diatoms increases. The planktonic *A. islandica* and *A. ambigua* dominate as earlier. Among the benthic diatoms, *Fragilaria brevistriata, F. construens et* var. *binodis, F. heidenii* appear in great quantities and are often subdominants.

Input of terrigenous material into the lake brought about the shallowing of the water basin. As a result, the total number of diatoms decreases upward the core. At the same time, the species composition of the benthic diatoms diversifies and their content in the sediments increases. The changes in the diatom species composition during the formation of sediments refer to ever growing eutrophication of the lake. Evidence is also derived from an abrupt increase in the index A/C (from 12 to 15% in the lower interval up to 30–37% in the upper interval). The index shows the ratio of centric and pennales planktonic diatoms, characteristic of hypereutrophic lakes (Stockner 1971). The appearance of the diatoms *F. berolinensis* and *F. parasitica* (W. Sm.) Grunow is another indicator of the eutrophication of the lake.

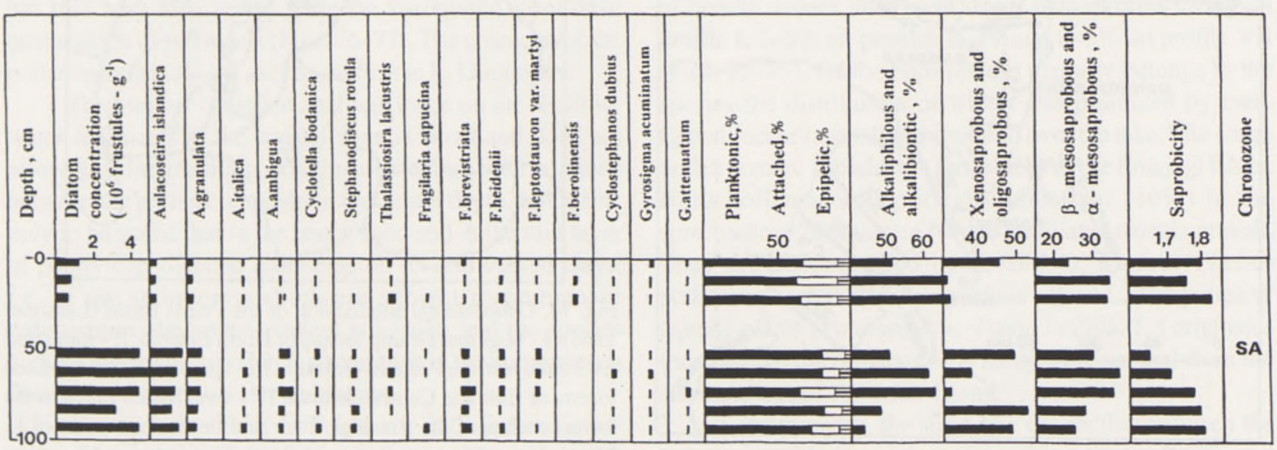


Fig. 72. Diatoms in L. Pihkva sediment core (east of Kamenka Island). For the legend see Fig. 69

Based on the results arising from the study of the species composition of diatoms in L. Peipsi, it is possible to reconstruct elements of the hydrological regime in proglacial lakes which existed in its basin during the Late-glacial and in the Holocene. The diatom evidence also support basic palaeogeographical stages established by means of other geological methods. The results of diatom analysis agree well with the geological assumption (Raukas & Rähni 1969), that during the Younger Dryas the lake depression was part of a huge glacial basin. In the Holocene, the deepest part of the depression was occupied by a lake called Small Peipsi (Raukas & Rähni 1969, Kvasov 1975). Pre-Boreal sediments were identified only in the core drilled in the southern part of L. Lämmijärv. These sediments were formed in a river valley, along which the water flowed to the north. In the Boreal and Atlantic, northern Estonia experienced an isostatic uplift, which obstructed the movement of water along the Narva River into the Gulf of Finland, and little by little the lake retreated southwards. As a result, Lahepera inlet became shallow and isolated from the main aquatory. The depression in the area of the present-day L. Lämmijärv turned deeper. In the Sub-Atlantic it was occupied by L. Pihkva. At the present time, the uplift of the northern coast is ongoing and the lake moves continuously in a southerly direction. The lake is subject to ever increasing eutrophication caused by natural and anthropogenic processes.

3.2.4.2. POLLEN AND SPORES

The composition and distribution of terrestrial and aquatic pollen in the uppermost layer of L. Peipsi's bottom deposits were studied at 120 points on 14 profiles (Fig. 73).

The purpose of the studies was to find out to what extent the recent surrounding plant cover is reflected in pollen spectra and how the currents, coastal erosion and granulometric composition of sediments affect the formation of pollen spectra.

After Laasimer (1958), the lake basin belongs to the East-Baltic Geobotanical Subprovince which is situated in the northern part of broad-leaved deciduous (hardwood) and coniferous-mixed forest zone. According to the Geobotanical Classification of Estonia (Laasimer 1958), the basin of L. Peipsi is subdivided into five geobotanical subzones. Together with the eastern part of the lake (Sochava *et al.* 1960), there are six geobotanical subzones (Fig. 74).

L. Peipsi *sensus stricto* is the deepest and largest part of the lake with a poorly dissected shoreline and flat bottom topography. Organic matter bearing pelitic silts are the prevailing surficial deep-water sediments. In the extreme northern and southern parts of the basin, till or varved clay is partially overlain by a 11–10-cm-thick layer of slightly consolidated silt and pelite, occasionally it is lacking. For

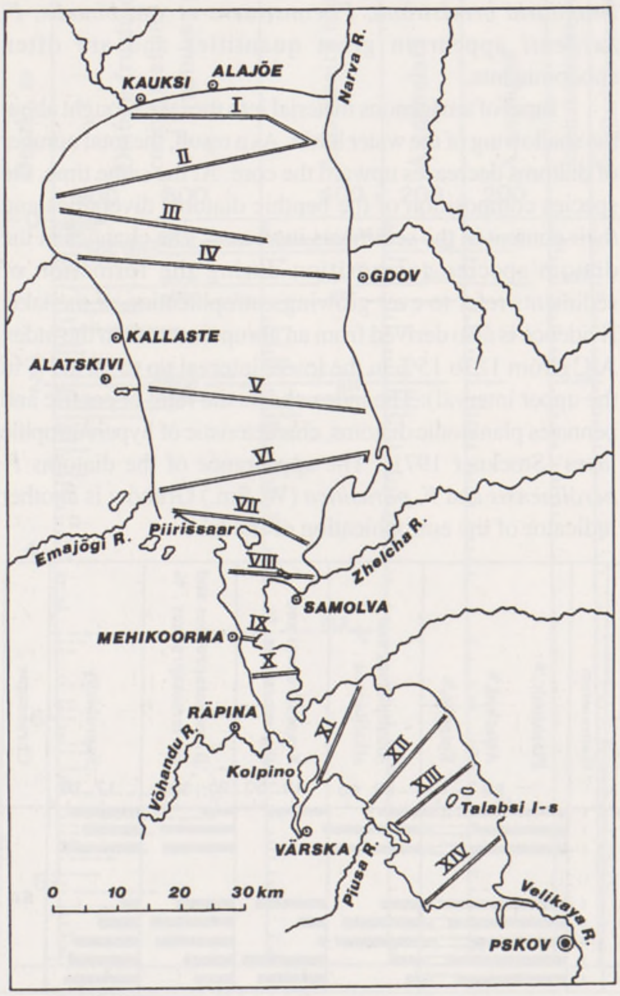


Fig. 73. Location of the studied profiles.

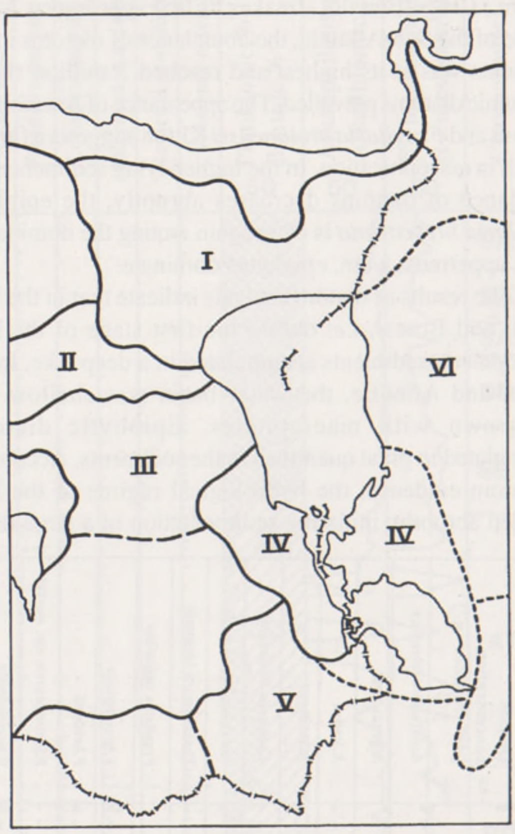


Fig. 74. Geobotanical subdivision of the Peipsi Basin (Laasimer 1964): I - bogs and swamp forests of North Estonia; II - mires and river valleys in the Pedja River basin; III - spruce and spruce mixed forests of East and Central Estonia; IV - river valley mires in the lower reaches of the Emajõgi River and on the east coast of L. Peipsi; V - sandy pine forests of South-East Estonia; VI - moss spruce forests and aspen-birch forests of the Gdov – Osmino region.

this reason, the sampling frequency was lower than initially planned. Sand and silt occur in the coastal zone of the lake.

Swamps, marshy pine forests, mixed forests and, to a lesser extent, black alder marshes are spread in the northern part of the Peipsi Basin. Those plant communities make up 60% of the vegetation. Transitional bogs and fens, pine and spruce forests, spruce-pine mixed forests account for 20–25%. The proportion of arable land does not exceed 5% (Laasimer 1964). Spruce and pine forests are the dominating forest types.

The spruce and mixed forest zone of eastern and central Estonia is spread between the Emajõgi and Omedu rivers on the west coast of the lake (Fig. 74). Fields dominate. Due to man's interference, spruce forests have been frequently substituted by birch and aspen woods. There is an obvious similarity between the vegetation on the west and east coast of the lake (Sochava *et al.* 1960). Mires in the river valley and marshy pine forests are the characteristic landscape components in the area adjacent to the Emajõgi River. Cultural landscapes are practically lacking. The opposite shore is covered with pine forests, bogs and fens.

In the northern part of the lake, the direction of water movement coincides with the direction of prevailing winds. Compensating currents reach the bottom of the lake (Jaani 1973, Kallejärv 1973).

Samples from 54 points on 7 profiles in L. Peipsi *s.s.* were subject to palynological studies (Fig. 73). Tree pollen is dominated by pine (average content 46%), which is more abundant in the northern and southern parts of the basin (Figs. 75, 76; profiles I, II, VI). Birch is the next abundant species (average 26%). On the profiles, from north to south, its proportion is 13–16, 19–28, 27–36, 30–38, 20–26, 8–30 and 8–50 (66)%, respectively (Figs. 75–77). The content of birch is clearly higher in the central part of the lake. The spruce pollen accounts for 14% and is more abundant in the north-eastern part of the lake on the Omedu profile (Fig. 76, profile IV) and very evenly spread on the Lahepera profile (Fig.76, V). Alder (average 13%) is abundant in the middle of the lake (Fig. 76, VI) where it makes up 14–18% of arboreal pollen. The proportion of broad-leaved trees is 0–3%. Willow is insignificant or entirely lacking.

On observation points within single profiles, the proportions of tree pollen are rather similar (Figs. 75–77). Pronounced differences in adjacent spectra occur only in profile VII. This may be due to redeposition of pollen by the bottom-reaching circulating currents and weathering of subwater peat, but this may also relate with the variegated grain-size composition of sediments (Figs. 76–77). The character of the pollen spectra is almost as changable as in L. Lämmijärv.

The number of pollen analyses made on the shallow-water sediments in the coastal zone is small and does not allow any far-reaching conclusions about interrelations between the pollen composition and water depth. It remains only to be stated that in the beach sand and in the thin layer of poorly consolidated sediments on till and varved clays, *i.e.* in the sediments which are subject to permanent redeposition, the pollen content is smaller and the species composition of herb pollen is poorer than in the homogenous fine-grained deeper-water sediments.

The surface area of L. Peipsi *s.s.* is large but, nevertheless, the generalised pollen spectra reflect essential changes in the vegetation in the surroundings of the lake as well. Thus, the high pine pollen content suggests a wide spread of pine forests and mires in the northern and southern parts of the lake. Ericaceae, including *Calluna vulgaris L.*, are often present. Birch is high in the central part of the lake where, on the both coasts, secondary birch woods have developed instead of pine groves. Alder is also more abundant in the middle of the lake where cultural landscapes are widespread on both coasts. Pollen of cultivated cereals and herbs (*Artemisia*, Chenopodiaceae, *Rumex*, *etc.)*, indicative of human impact, is present. Herb pollen forms 2–3% on profile I, 3–5% on profiles II–IV and 0–3% on profile VII (Figs. 75–77). Of the herb pollen, majority belongs to the species the distribution of which was promoted by man. Cereals occur in small quantities all over the lake. The sedge pollen is more abundant in the estuary of the Emajõgi River. Herb pollen is exceptionally abundant (10%) in the surroundings of Mustvee, where cultivated cereals prevail. They are accompanied by *Artemisia, Rumex*, Chenopodiaceae, including *Chenopodium album* L. The pollen of aquatic plants (*Potamogeton, Typha latifolia* L.) originates mostly from the estuary of the Emajõgi River and from the surroundings of Piirissaar Island.

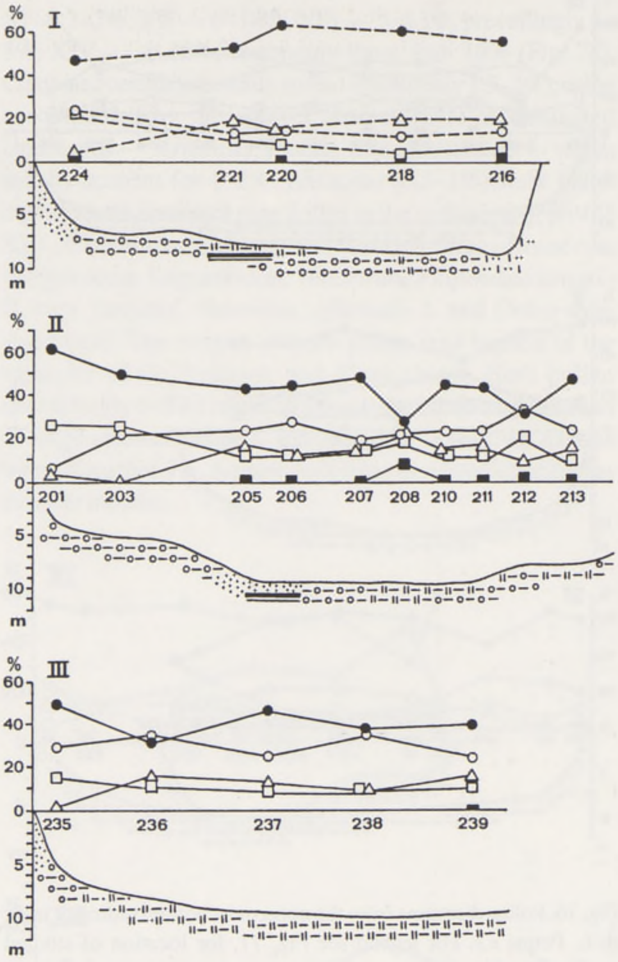


Fig. 75. Pollen diagrams from the uppermost bottom sediment layer in L. Peipsi *s.s.* For legend see Fig. 77, for location of studied profiles see Fig. 73.

L. Lämmijärv, the strait-like connection between the larger parts of L. Peipsi, has a jagged coastline and dissected

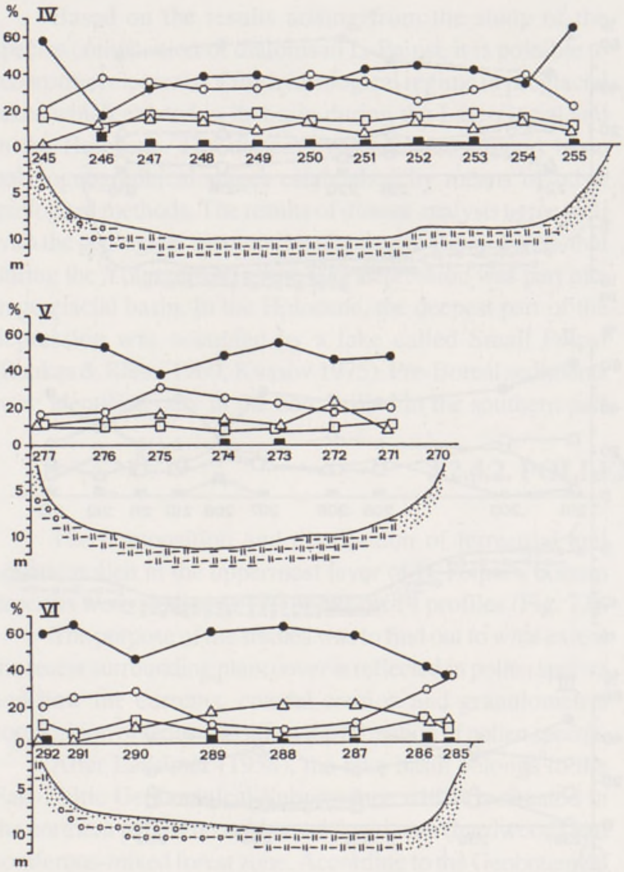


Fig. 76. Pollen diagrams from the uppermost bottom sediment layer in L. Peipsi *s.s.* For legend see Fig. 77, for location of studied profiles see Fig. 73.

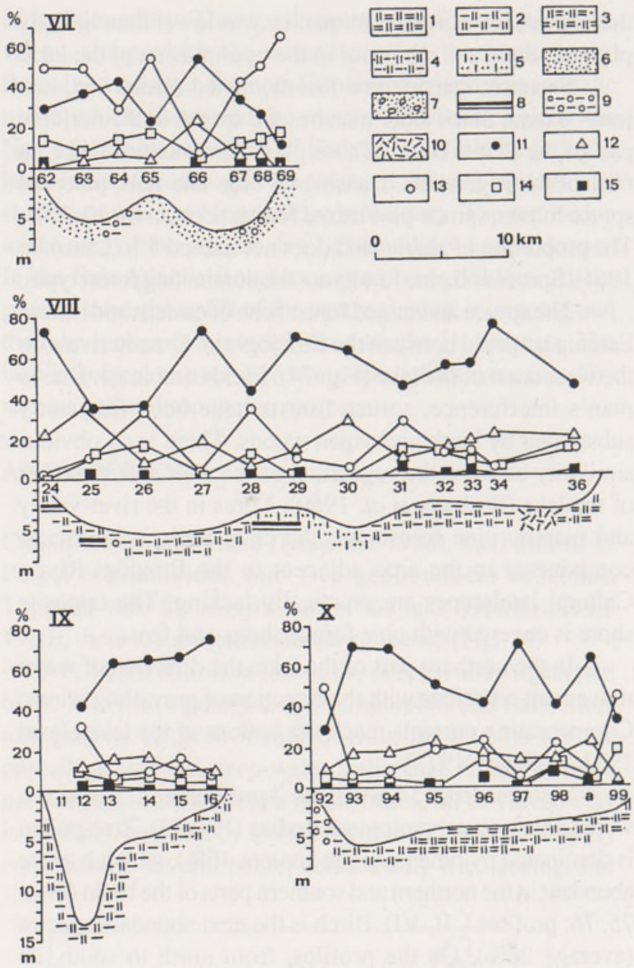


Fig. 77. Pollen diagrams from the uppermost bottom sediment layer in L. Lämmijärv: 1 - silt; 2 - somewhat clayey silt; 3 - somewhat sandy clayey silt; 4 - somewhat clayey and sandy silt; 5 - silty sand; 6 - sand; 7 - somewhat silty sand; 8 - varved clay; 9 - till; 10 - fen peat; 11 - pine; 12 - spruce; 13 - birch; 14 - alder; 15 - broad-leaved trees. For location of the studied profiles see Fig. 73.

bottom relief. In L. Lämmijärv, the water moves both to the south and regardless of the direction of wind. In autumn, the flow is prevailingly to the north (Kallejärv 1973).

Bottom sediments are prevailed by clayey and sandy clayey silts (Fig. 59). In the northwestern part, but also between Meeksioja and Raigla, varved clays crop out on the lake floor. Subwater wood and reed peat is encountered. In the bottom sediments peat occurs in lumps and also in a scattered form (Fig. 58).

Marshy pine groves, bogs and fens are spread in the environs of the lake, mainly on the east coast. Mixed forests with birch and spruce cover higher spots, particularly on the west coast. Bilberry-rich pine and spruce forests make up 30% of the forests (Valk & Eilart 1974).

In L. Lämmijärv, the pollen composition was studied in the topmost layer of bottom sediments. Samples were taken from 26 points on three profiles (Figs. 73, 77). Although the spruce and pine pollen content differs from profile to profile (Fig. 77), pine clearly dominates. From north to south (profiles VIII-X, Fig. 77), the average content of pine pollen increases from 39 to 54%. The amount of birch ranges from 15 to 20%. Compared to L. Peipsi *s.s.*, spruce is relatively high: 9–30% (average 18%), 6–37% (average 20%) and 9–22% (average 19%), respectively. The proportion of alder averages 7-9%. The pollen of broad-leaved trees ranges from 0 to 4%, and only in one sample it reached 7%, evidently due to redeposition from older peat. In half of samples willow was absent, in another half it made up 0.5–2%. Only in the shallow marshy coastal area near the east shore (profiles VIII,X, Fig. 77), willow increases to 3–5%. Since willow is an insect-pollunated species, its pollen does not spread far.

The content of herb pollen varies regularly within profiles: VIII — 1–6%, IX — 1–6%, X — 1–4% (Fig. 77). As an exception serves the west coast where herb pollen amounted to 9%. The pollen of cultivated cereals has been constantly found on the Mehikoorma profile (IX) and in the Samolva area in the estuary of the Zhelcha River. The variability of herbs is modest. The most common representatives are *Artemisia*, Chenopodiaceae and *Rumex*. *Ericales* is also frequent. Cereals and sedges occur in small amounts everywhere, but sedges are somewhat more abundant in the estuary of the Zhelcha and Võhandu rivers.

The changes in the ratio of forest-forming species (pine, birch, spruce, alder) pollen in the bottom sediments of L. Lämmijärv are more abrupt than in L. Peipsi *s.s.* or L. Pihkva. This is evidently caused by the erosion of the peat exposed on the bottom of L. Lämmijärv or on the shore. Pollen may have been deposited by bottom currents causing the erosion of bottom sediments. Nevertheless, the presence of spruce and pine forests around the basin is reflected in the pollen spectra of lake sediments. The content of spruce and pine pollen in bottom sediments is higher in L. Lämmijärv than in L. Peipsi *s.s.* or L. Pihkva. But it is difficult to explain why the sediments in L. Lämmijärv are low and impoverished in herb pollen, although the proportion of arable land is at least 30–40% on its west coast (Laasimer 1964). One possible reason may be the location of fields farther away from the lake.

L. Pihkva is the southernmost part of L. Peipsi. It has flat bottom and only slightly dissected shoreline. The shoreline is somewhat more dissected in the northern part of the lake. Clayey silts are spread in the deeper parts and fine-grained clayey silty and clayey sands in the coastal zone of the lake (Fig. 59). In the northern part, till is in places overlain by a thin (2–3 cm) layer of lake sediments. The direction of the movement of prevailing currents coincides with that of prevailing winds (Kallejärv 1973).

The surroundings of the lake are swampy. Fens, raised bogs and pine forests dominate. Moss-rich spruce forests are spread on the east coast. Several geobotanical districts, with a modest spread of mires, are distinguished south of the lake. Dry and swampy meadows abound. Valley flat meadows and pine forests with an abundance of mosses are spread in the surroundings of the Velikaya River. Spruce forests or spruce forests with an admixture of broad-leaved trees, and pine forests occur on higher spots.

In the topmost layer of lake sediments, pollen composition has been determined at 50 points on 4 profiles and, additionally, also in the coastal zone of the lake (Figs. 73, 78).

In L. Pihkva, pine and birch dominate in the bottom sediments. The quantity of pine pollen is lower than in L. Lämmijärv, but it is still higher than birch pollen (Fig. 78, profiles XI, XIII). Birch and pine are represented in almost equal amounts, or birch is slightly prevailing, in the sediments of the profile starting from the mouth of the Piusa River (Fig. 78, profile XII), and also in the sediments of the southernmost profile which is situated near the mouth of the Velikaya River (Fig. 78, XIV). The mean content of pine pollen in lake sediments is 45%, on the profiles (from north to south) it is 50, 39,47 and 40%, respectively (Fig. 78). The mean content of birch pollen is 29%, and on the profiles it is 27, 34, 27 and 30%, respectively (Fig. 6). Alder accounts for 14%, being at its highest in the vicinity of Kamenka Island, in the middle of the lake and in the surroundings of the Velikaya River mouth.

Bottom deposits of L. Pihkva are much poorer in spruce pollen than the other parts of L. Peipsi. The average content of spruce pollen is 10%, on the profiles (from north to south) — 11, 10, 12 and 12%, respectively. The highest values were recorded in the environs of the Velikaya River mouth, north of Talabskij islands and north-east of Kolpino Island.

The pollen of broad-leaved trees was present at all observation points. It usually formed 2-3%, in the southern part of the lake - 4-5%. Willow was present in most of samples; it usually reached 2-3% in both the deeper-water part of the lake and in the coastal zone.

In the bottom sediments of L. Pihkva, the proportion of herb pollen is relatively high and variable in species. In the studied profiles, the proportion of herb pollen was as follows: XI — 3–7%, XII — 6–10%, XIII — 5–13%, prevailingly 6–8%, XIV — 6–12%, commonly more than 10% (Fig. 78). Cultivated cereals are evenly spread (commonly 1.5–2%), being accompanied by the species characteristic of cultivated landscapes — *Artemisia, Rumex,* Chenopodiaceae — which usually account for 1–2%. Ericaceae (0.5–1%) were found together with abundant pine pollen in the sediments of profile XIII. Cruciferae, Compositae, Umbelliferae, Rosaceae, Polygonaceae, Leguminosae, *Thalictrum, Filipendula ulmaria L.* were frequent. *Valeriana officinalis L.* and *Urtica* were also found. The content of herb pollen was highest in the estuaries of the Velikaya and Piusa rivers. Herb pollen (prevailingly 4–5%) is spread all over the lake, the proportion of sedge is lower. Sedge is spread mainly in the river mouths where it reaches 5%. Aquatic pollen has also been identified in the river mouths.

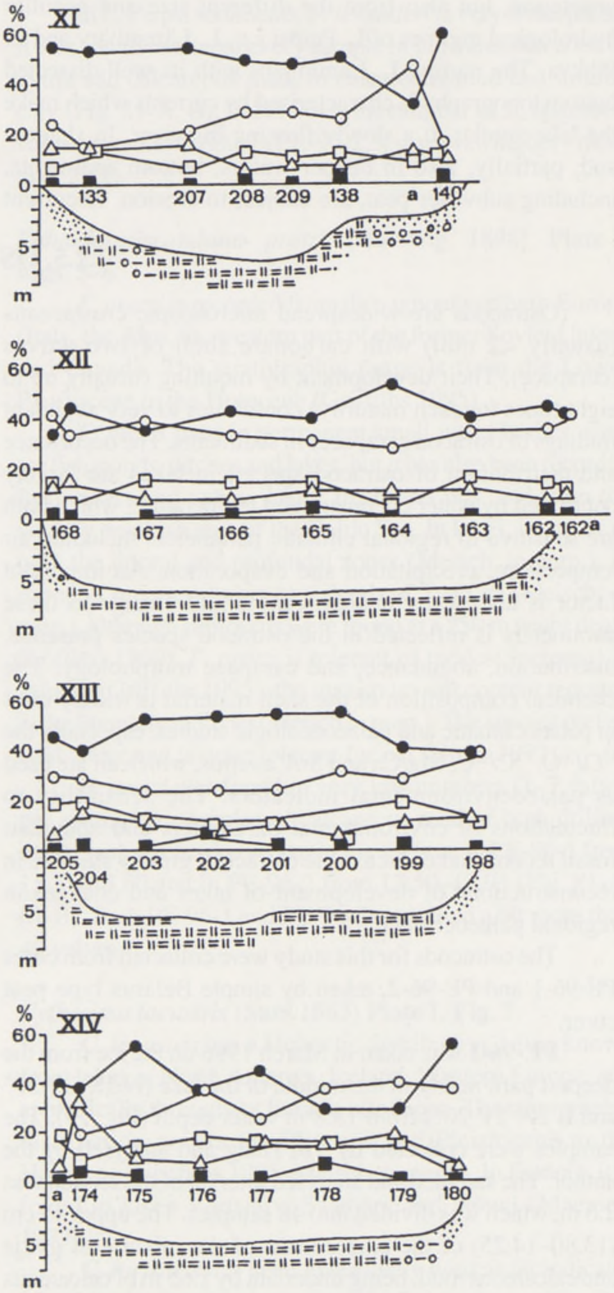


Fig. 78. Pollen diagrams from the uppermost bottom sediment layer in L. Pihkva. For legend see Fig. 77.

Pollen spectra reflect rather well the differences in the composition of vegetation around the southern tip of L. Pihkva. These are revealed, first of all, in the lower concentration of pine pollen and higher concentration of alder, birch, spruce, broad-leaved species and herbs on the southernmost profile.

To sum the above, it may be said that the features characteristic of both the vegetation zones and subzones are well reflected in the pollen spectra of bottom sediments in L. Peipsi. Differences between geobotanical regions are also evident as are the consequences of human influence on the vegetation.

Pollen input from valley flat meadows and cultural landscapes has been registered in the vicinity of the estuaries of bigger rivers, *e.g.* the Velikaya. Variations in the distribution of pollen established in the different parts of the lake result not only from the changes in the surrounding vegetation, but also from the different size and peculiar hydrological regimes of L. Peipsi *s.s.*, L. Lämmijärv and L. Pihkva. The narrow L. Lämmijärv with its well-dissected bottom topography is characterised by currents which make the lake similar to a slowly flowing big river. In shallow and, partially, also in deeper waters, bottom sediments, including subwater peat, are subject to erosion. Recurrent resedimentation of pollen has caused abrupt and unexpected changes in the distribution of pollen of different trees. In L.Peipsi *s.s.* and L. Pihkva, pollen is spread evenly enough. Fluctuations in pollen quantity occur mostly close to the shore, in the zone of intensive wave action. In deeper areas, the concentration of pollen from the plants on coast, equalizes. It is noteworty that in L. Peipsi *s.s.*the quantity of herb pollen is only half as high as in L. Pihkva. The reason may be the large area of L. Peipsi *s.s.* In both lakes, the quantity of herb pollen is higher in the deeper parts which indicates a great role of wave action in the distribution of pollen in large lakes.

Palynological research into the surface layer of bottom sediments in L. Peipsi confirms once again that the profiles from deeper-water and, hence, calmer sedimentation environments provide a more reliable basis for classification of sediments even in large lakes (Pirrus 1981). While interpreting the pollen analysis data on shallow-water and transitional zones, one has to consider the possibility that winds may have caused irregular changes in the concentration of pollen of main forest-forming species and herbs. In the event of L. Lämmijärv and L. Pihkva, some pollen embedded in Holocene sediments may have reached there by way of redeposition from subwater or inundated Holocene peat.

3.2.5. OSTRACODS

Ostracods are widespread microscopic crustaceans (usually <2 mm) with carbonate shell of two valves (carapace). Their development by moulting (usually up to eight times to reach maturity) contributes to very abundant findings of ostracod carapaces in sediments. The occurrence and distribution of ostracod species in lakes are largely controlled by water chemistry and temperature, which both are sensitive to regional climatic parameters including air temperature, precipitation and evaporation. An important factor is also water energy. Any slight change in these parameters is reflected in the ostracod species presence, distribution, abundance, and carapace morphology. The chemical composition of the shell material is widely used in palaeoclimatic and palaeoecologic studies, especially the $^{18}O/^{16}O$, $^{13}C/^{12}C$, Mg/Ca and Sr/Ca ratios, which all are used as palaeoenvironmental indicators. The sensibility to fluctuations of environmental parameters and abundant fossil record makes lacustrine ostracods greatly suitable in reconstructions of development of lakes and changes in regional palaeoclimate.

The ostracods for this study were collected from cores PE-96-1 and PE-96-2, taken by simple Belarus type peat corer.

PE-96-1 was taken in March 1996 on the ice from the deepest part, nearly in the middle of the lake (N 58° 39' 59" and E 27° 21' 20") from 13.8 m water depth (Fig. 79). The samples were collected by Tiit Hang and analysed by the author. The sampled and analysed interval of the section was 2.6 m, which was divided into 48 samples. The upper 45 cm (13.80–14.25) of the core consist of the mixture of gyttja and calcareous mud, being underlain by 1.65 m of calcareous mud (14.25–15.90) and 0.5 m (15.90–16.40) of reddish clay (Fig. 80-A).

PE-96-2 was taken in March 1996 on the ice in the western part of L. Peipsi (N 58° 44' 49.7" and E 27° 12' 05.5") from 9.5 m water depth (Fig. 79). Samples were taken and analysed by the author. The length of the core was 5.8 m, which was divided into 75 samples. The upper third of the section (9.70–11.60) consists of gyttja, grading deeper into calcareous mud (11.60–13.64). A thin layer of grey sand (13.64–13.70) was not sampled for ostracods. The lower third (13.70–15.50) of the sequence consists of homogenous (grey and red) and varved clays (Fig. 80-B).

The clays in the deepest part of the sections are supposedly of the Late Weishelian age and are overlain by Holocene calcareous mud and gyttja.

The ostracod collection (Coll. No. 1043 – Quaternary), samples of calcareous mud and drilling cores are deposited in the Institute of Geology, University of Tartu.

For choosing the size of serial samples, a set of test samples of different sediment types was processed first. 10-cm intervals of sediment cores were taken from gyttja, calcareous mud, grey and red clay (varved clay). The samples were wet sieved through 125μ sieve and the number of ostracods in samples was estimated visually. Ostracod density appeared to be high in calcareous mud and grey clay, and low in varved and red clay and in gyttja.

According to the preliminary results, the sample size for throughout sampling was chosen as follows: 5 cm interval (*ca* 30 cm^3) for calcareous mud, grey and red clay, and 10 cm (60-70 cm^3) for gyttja and varved clay. Before wet sieving with a 63μ sieve (used for keeping also early juvenile specimens), the volume of samples was measured in scale class for further data unification and determination of the sand fraction in samples. After sieving the samples were dried at 80° C, then the weight of sediment fraction above 63μ were measured. The sand fraction content in sediment in grams per litre was calculated (see Figs. 80-A and 80-B).

Fig. 79. Location of drilling cores in Lake Peipsi.

Ostracod shells were picked out under a binocular microscope with a fine wet brush from the fraction above 63μ. The identification was performed using literature and consulting with Dr U. v. Grafenstein, Dr H. Griffiths and Dr C. Meisch. Specimens were counted for analysis of abundance and assemblage structure in samples and different sediment types.

The ostracods were photographed with the scanning electron microscope JEOL in the Laboratory of Material Study at Tallinn Technical University.

Taxonomic composition. In the two drill cores studied, more than 24,000 ostracod valves are attributed to 8 species of the order Podocopida Sars, 1866 (suborder: Podocopina). Some notes on the distribution and ecological preferences of the species are presented below.

Candona candida **(O.F. Müller 1776) Plate I, Figs. 1–2**

C. candida is widespread throughout North America, Europe and Eurasia (Holarctic), rare in the south (Meisch, in prep.). The species also occurs within the Arctic (Griffiths, 1995). The stratigraphic range of *C. candida* is from the Upper Pliocene to the Holocene (Meisch, in prep.).

C. candida occurs in an exceptionally wide range of aquatic habitats; it is found in lakes, (fish)ponds, ditches, swamps, acid peaty waters, brooks, rivers, springs, wells and in different types of subterranean habitats, also in temporary pools. In lakes, it occurs mainly in the littoral zone, but is found in small numbers down to a depth of 300 m (Griffiths & Martin 1993). The maximum salt content reported in the literature is 5.77‰ (Meisch, in prep.). *C. candida* is considered a permanent species in those habitats where the water temperature does not rise above 18°C (Hartmann & Hiller 1977). Zubowicz (1978, 1983) has used *Candona sp.* in palaeoecological reconstructions as an indicator species of cold and at least 1 m deep water, who prefers muddy bottom sediment with rich vegetation.

In the sediments of L. Peipsi, *C. candida* is one of the most abundant species. It is present in most samples taken from the transition zone between gyttja and calcareous mud, in calcareous mud and in grey clay. The number of valves is more than 1000 in PE-96-1 and more than 1450 in PE-96-2. In both drilling cores, the richest-in-ostracods samples are from the middle part of the calcareous mud (Fig. 81-A, B).

Fabaeformiscandona levanderi **(Hirschmann 1912) Plate I, Figs. 3–4**

F. levanderi is recorded from several places throughout Europe (Meisch, in prep.). The stratigraphic range is from the Lower Pleistocene to the Holocene (Meisch, in prep.).

F. levanderi occurs in (fish)ponds and lakes, in both littoral and profundal zones (the greatest depth reported is 62 m; Löffler 1969). The reported salt range of the species is 1.0–6.0‰ (Meisch, in prep.). *F. levanderi* prefers cold waters and is tolerant of different water velocity (Meisch, in prep).

In L. Peipsi sediments, *F. levanderi* is very widespread. It was found in almost every sample in the transition zone of gyttja and calcareous mud, in calcareous mud and in grey clay (Fig. 81-A, B). In PE-96-1, the number of *F. levanderi* valves exceeded 1600. In PE-96-2, it was even higher - more than 2300.

***Fabaeformiscandona protz*i** **(Hartwig 1898) Plate I, Figs. 5–6**

F. protzi is recorded from throughout northern Europe (Italy, the Alps, the western part of the former Soviet Union) and Canada. The stratigraphic range is from the Lower Pleistocene to the Holocene (Griffiths 1995).

F. protzi lives in permanent small water bodies, such as (fish)ponds, ditches and lakes, but it has also been recorded from rivers and oxbow lakes, from peat pits, and from the slightly brackish area of the Baltic Sea. In lakes, it occurs in both the littoral and profundal zones (Meisch, in prep.). *F. protzi* is reported living at a depth of 7–45 m (Meisch, in prep.), although subfossils were found at a 250 m water depth (Löffler, 1969). *F. protzi* is tolerant of modest increases in salinity (Griffiths 1995), the maximum salt content reported in the literature is 5.8‰ (Meisch in prep.). The species prefers cold water and is quite tolerant for pH (Hiller 1972).

F. protzi was found in very low numbers (1–7 valves per sample) in samples of calcareous mud of both drilling cores, with a remarkable gap in the section, in PE-96-1 from 15.35–15.60 and in PE-96-2 from 12.50–12.95 (Fig. 81-A, 81-B). Both PE-96-1 and PE-96-2 contained a bit more than 40 valves.

Cytherissa lacustris **(Sars 1863) Plate I, Fig. 7**

C. lacustris has a Holarctic distribution, being known from lakes in North America, Iceland, Western Europe, and sporadically throughout Eurasia into Japan. The stratigraphic distribution covers the Pliocene and Pleistocene to the Holocene (Griffiths 1995; Meisch, in prep.). In Estonia, it is found in lakes Võrtsjärv, Saadjärv and Nõuni (Mäemets 1977).

C. lacustris has been extensively used as an indicator of cold, oligotrophic waters, and of changing patterns of mixes and trophic status in the development of lakes. *C.*

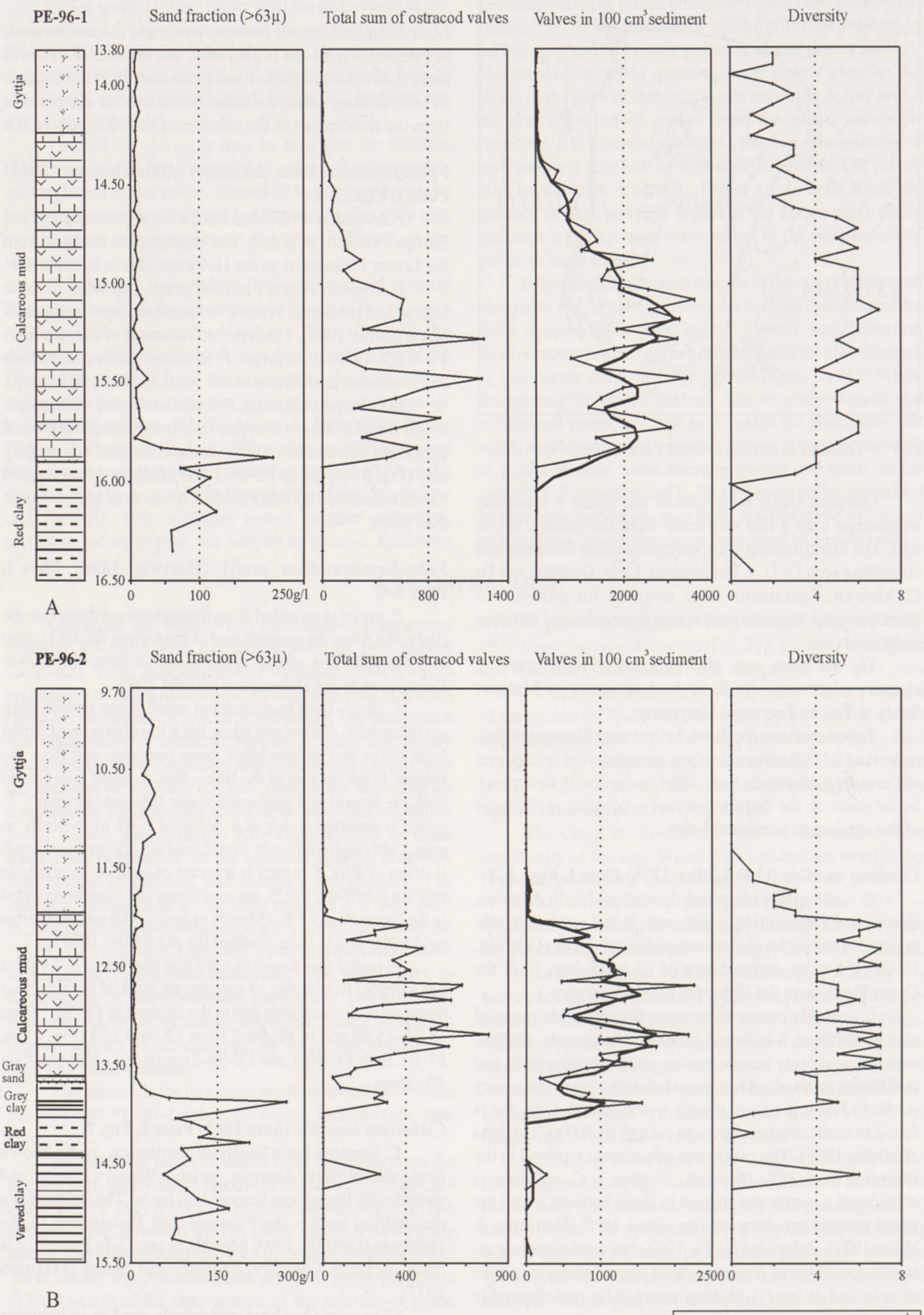


Fig. 80. **A.** Content of the sand fraction, total and relative amount of ostracod valves and diversity of the core PE-96-1. **B.** Content of the sand fraction, total and relative amount of ostracod valves and diversity of the core PE-96-2.

lacustris exclusively inhabits the benthos of cold, well-oxygenated lakes of low productivity, where it lives in the sublittoral and profundal zones (Griffiths 1995) from a depth of about 3 m down to the very great depths of the profundal zone (up to 200 m, with one record even at a depth of 220 m; Löffler 1969). The highest densities seem to occur in oligo-mesotrophic lakes at depths between 12 and 40 m (Meisch, in prep.). It lives on fine, silty sediments, in which it may burrow to depths of several centimetres (Griffiths 1995: Meisch, in prep.). However, the species is also reported from shallow ponds, shallow littoral zone of lowland lakes and even from swamps in Sweden. At Hemmelmarkersee, the species lives at a depth of 1.5 m in slightly brackish water (salinity 0.5–1.5‰) subject to wide annual thermal fluctuations (20°C between summer and winter; Delling 1981). *C. lacustris* is a temperature tolerant form with preference for cold water (Meisch, in prep.)

C. lacustris is the dominant species in two sampled core sections. It is present almost in every ostracod-containing sample, except for some rare samples in the transitional zone between gyttja and calcareous mud (Fig. 81-A, 81-B). The number of *C. lacustris* valves is more than 9100 in PE-96-1, and more than 7700 in PE-96-2 .

Ilyocypris bradyi **(Sars 1890) Plate I, Fig. 8**

I. bradyi has been recorded almost in every part of Europe, except the Crimea and in the area around the northern Baltic. *I. bradyi* is also known from Canada and the United States, North Africa, the Middle East, Central Asia (Meisch, in prep.). Its stratigraphic range is from the Lower Pleistocene till the Holocene (Griffiths 1995, Meisch, in prep.).

I. bradyi is found in the bottom sediments of ponds, eutrophic gravel pits, swamps, lakes, and streams (Griffiths 1995, Sokolov 1989). It is also reported from interstitial water (Marmonier & Creuzé des Châtelliers 1992) and is occasionally found in temporary pools (Meisch, in prep.). The species lives on muddy and sandy substrates. Bronshtein (1947) gave also records from springs, lakes, oxbows and river tributaries up to depths of about 4 m, and upon most types of substratum. However, he believed that *I. bradyi* was strongly associated with springs so that its occurrence in other habitats was primarily through drift. Zubowicz (1978, 1983) regards *I. bradyi* as an indicator of very shallow water (5–50 cm), although this range is probably too restricted. *I. bradyi* has been recorded from slightly brackish inland waters in Europe (salt content 0.3–0.4‰; Vesper 1975). The species prefers cold water, although in one locality it was found at 24°C (Vesper 1975).

The distribution of *I. bradyi* is related to calcareous mud and grey clay (Fig. 81-A, B). In PE-96-1, the number of valves reaches nearly 140 and in PE-96-2 nearly 100.

Limnocythere sanctipatricii **(Brady & Robertson 1869) Plate I, Figs. 9–10**

L. sanctipatricii is found in the lakes of North and Central Europe, and in the North. The stratigraphic distribution is from the Lower Pleistocene to the Holocene (Griffiths 1995; Meisch, in prep.). In Estonia it is found in lakes Nõuni and Saadjärv (Mäemets 1977)

L. sanctipatricii is a cold-stenothermal form, almost invariably occurring in lakes and other stable still bodies of water, such as ponds and ditches. There are no records from flowing waters. Occasionally it is found also in peaty waters. One record from interstitial water is probably related to passive drift (Marmonier 1985). *L. sanctipatricii* may be sensitive to changes in trophic degree (Griffiths 1995). In lakes, it occurs from the shallow littoral zone down to the great depths of the profundal zone. In Lake Constance, living specimens were captured at a depth of 250 m, which is the greatest depth of the lake (Löffler 1969). The species seems to be restricted to oligotrophic habitats (Scharf 1981). The species has been reported from waters with salinity up to 3‰ (oligohaline range; Meisch, in prep.).

L. sanctipatricii spreads mostly in calcareous mud, although some valves occur in grey clay as well. The number of valves of *L. sanctipatricii* is more than 50 in PE-96-1 and more than 140 in PE-96-2.

Darwinula stevensoni **(Brady & Robertson 1870) Plate I, Fig. 11**

D. stevensoni has a cosmopolitan distribution, and is known from the whole world, except Australasia and Antarctica N (Griffiths 1995). Within Europe, *D. stevensoni* seems to be present in all countries south of the 60° latitude (Griffiths & Butlin 1994). Its stratigraphical distribution is from the Mid-Oligocene to the Holocene (Meisch, in prep.).

D. stevensoni occurs in (fish)ponds, lakes and slow streams, being also reported from the interstitial groundwater. It lives on both muddy and sandy substrates. The reported water depths of their distribution are from 0 to 8.5 m (Griffiths 1995), but Zubowicz (1978, 1983) regards this species as an indicator of deep water (at least 2.0–2.2 m), what may be a little too deep for being limiting. Anyway, according to literature, the species seems to reach the peak abundance between 1.5-6 m depth. It tolerates increase in salinity up to a maximum of 15‰ (*ca* 7.5 g Cl^{-}/l) (Meisch, in prep.).

In the bottom sediments of L. Peipsi, *D. stevensoni* is not very abundant. In PE-96-1, it is represented in the middle part of calcareous mud with 1–5 valves per sample. In PE-96-2, it is also found in the transitional layer between calcareous mud and gyttja, 1–16 valves per sample. In both PE-96-1 and PE-96-2 there is a gap where specimens *of D. stevensoni* absent, which regarding the presence of shallow water species may refer to a shallowing of the lake (Fig. 81-A, 81-B). The number of valves in PE-96-1 reached almost 40 and in PE-96-2 it was a bit more than 60.

Herpetocypris reptans **(Baird 1835) Plate I, Fig. 12**

H. reptans is cosmopolitan, except for Antarctica. It can be found from virtually every part of Europe and even in North Africa and Zaire (Griffiths 1995). The stratigraphic distribution is from the Middle Pleistocene to the Holocene (Griffiths 1995; Meisch, in prep.).

H. reptans prefers permanent bodies of water, usually with rich vegetation and muddy bottom, such as (fish)ponds, ditches, slow streams and rivers. Other records come from springs, swampy waters and temporary bodies of water. In lakes, it is generally restricted to the littoral zone (Scharf

1981; Scharf *et al.* 1995), although it was recorded down to a depth of 15 m (Löffler 1969). There is one record from a hyporheic and another from an interstitial habitat (Nüchterlein 1969, Tabacchi & Marmonier 1994). The salinity range is 0.5–6.0‰ (Usskilat 1975).

In the studied sections, *H. reptans* occurs in well defined quite narrow levels in calcareous mud: in PE-96-1 it is present in level 15.40–15.75 cm (1–9 valves per sample) and in PE-96-2 in level 12.50–12.95 (1–24 valves per sample). Some random samples containing odd valves of *H. reptans* (1 valve per sample) were also found in PE-96-2. The occurrence of *H. reptans* coincides with the absence of *D. stevensoni* and may indicate a shallowing of the lake (Fig. 81-A, 81-B).

Stratigraphic distribution of ostracod fauna and its palaeogeographical importance

PE-96-1. The total number of ostracod valves per sample in the section PE-96-1 fluctuates from 0 to more than 1100 per sample (Fig. 80-A). In red clay ostracods are very rare and their distribution is related mostly to calcareous mud.

The dynamics of amount of ostracod valves in 100 cm^3 of sediment (Fig. 80-A) repeats almost the dynamics of absolute numbers, but has well expressed three peaks of the maximum abundance. The first increase of the sum of valves topped at 15.70–15.75, then decreased until 15.60–15.65. The next peak is at the interval of 15.45–15.50. The third appears at the 15.05–15.10 with density of more than 3500 valves per 100 cm^3 of sediment, which is followed by slow decrease of the density of ostracod valves in sediment.

The ostracod diversity (Fig. 80-A) is high at the interval 15.85–14.20 m. From this point, the decrease in faunal diversity begins.

Three species (*C. lacustris*, *F. levanderi* and *C. candida*) are most abundant in assemblage and are distributed over the whole fossiliferous interval. The densities of *C. lacustris* (max 2546 valves in 100 cm^3 of sediment) and *F. levanderi* (max 776 valves in 100 cm^3 of sediment) are quite proportional with the total density of valves (Figs. 80-A and 81-A).The density of *C. candida* (max 672 valves in 100 cm^3 of sediment) is following also the main fluctuations of the total density of ostracods, but its maximum is a bit earlier (15.70–15.75; Fig. 81-A). The distribution span of *I. bradyi* (max 103 valves in 100 cm^3 of sediment, Fig. 81-A) is shorter than that of previous three species – it spreads from 14.70–15.70, and has its maximum at the point of 15.45–15.50. *F. protzi* (max 42 valves in 100 cm^3 of sediment; Fig. 81-A) occurs in two intervals: 15.60–15.80 and 14.60–15.30. *D. stevensoni* (max 42 valves in 100 cm^3 of sediment; Fig. 81-A) appears first in the sediments only from 15.45 m and disappears at 14.35 m in the section. Most of *L. sanctipatricii* (max 67 valves in 100 cm^3 of sediment per sample; Fig. 81-A) is found in the interval of 14.65–15.10, although it occurs also in the lower part of the section. *H. reptans* (max 50 valves in 100 cm^3 of sediment; Fig. 81-A) occurs only in a thin layer from 15.40 to 15.75 m of the section.

PE-96-2. The total number of ostracod valves per sample in the section PE-96-2 fluctuates (Fig. 80-B) from 0 to more than 800 per sample. After the considerable amount of ostracods at the exceptional level of 14.50-14.70 m in varved clay there is following quite remarkable level where no ostracods are found. Ostracods appear again at the level of grey clay and are quite abundant in the latter and in calcareous mud.

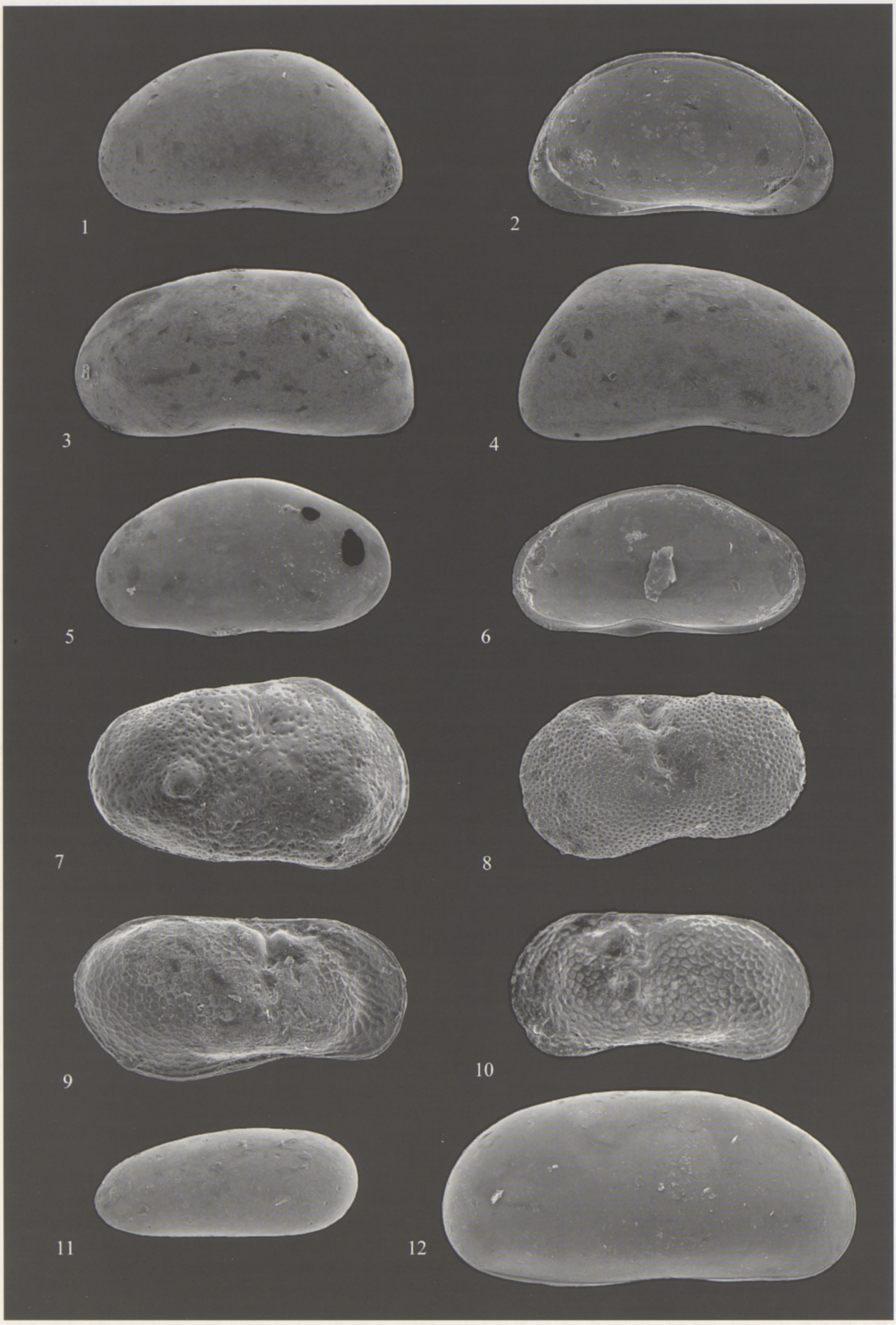
The amount of ostracod valves in 100 cm^3 of sediment (Fig. 80-B) varies from 0 to more than 2250 valves. Like in PE-96-1, three levels of great abundance of ostracod valves are noticeable and 2 low-density levels between them (better expressed by moving 3-sample average). The first peak is at the level of 13.85–13.90 (1555 valves per 100 cm^3 of sediment) followed by the decrease of ostracod density until 13.60–13.64 (184 valves per 100 cm^3 of sediment). The highest peaks and sinks fall into the interval of 13.05–13.10 (*ca* 1500 valves per 100 cm^3) and 12.65–12.70 (*ca* 2250 valves per 100 cm^3). The minimum between them is at 12.97–13.00 (500 valves per 100 cm^3).

The faunal diversity in Fig. 80-B is quite high in the clay sample at 14.50–14.70 and in calcareous mud (with a quick rise from 14.00–14.05 and a bit slower steady fall from 12.05–12.10).

C. lacustris, *F. levanderi* and *C. candida* are also here the most abundant species in assemblage and are distributed along the whole fossiliferous interval. The densities of *C. lacustris* (max 1383 valves per 100 cm^3 of sediment) and *F. levanderi* (max 765 valves per 100 cm^3 of sediment) are proportional with the total density (Figs. 80-B and 81-B). The density of *C. candida* (max 378 valves per 100 cm^3 of sediment) follows also the main fluctuations of the total density of ostracods, but its maximum (378 valves per 100 cm^3 of sediment) is a bit higher (13.05–13.10; Fig. 81-B). *L. sanctipatricii* (max 105 valves per 100 cm^3 of sediment; Fig. 81-B) is distributed almost along the whole sequence, has its maximum at the level of 12.45–12.50 m, but its range

Plate I ▷

1. *Candona candida* Müller (1776); adult female left valve; PE-96-2 – 13.10-13.15; length 1.00 mm: 58×.
2. *Candona candida* Müller (1776); adult female left valve; PE-96-2 – 13.10-13.15; length 1.00 mm: 58×.
3. *Fabaeformiscandona levanderi* Hirschmann (1912); adult female left valve; PE-96-2 – 12.50-12.55; length 1.13 mm; 57×.
4. *Fabaeformiscandona levanderi* Hirschmann (1912); A-1 female right valve; PE-96-2 – 13.25-13.35; length 1.05 mm; 61×.
5. *Fabaeformiscandona protzi* Hartwig (1898); adult male left valve; PE-96-2 – 12.15-12.20; length 1.05 mm; 53×.
6. *Fabaeformiscandona protzi* Hartwig (1898); A-1 male right valve; PE-96-2 – 13.40-13.45; length 0.98 mm; 57×.
7. *Cytherissa lacustris* Sars (1863); adult female right valve; PE-96-2 – 12.20-12.30; length 0.95 mm; 64×.
8. *Ilyocypris bradyi* Sars (1890); adult female left valve; PE-96-2 – 13.00-13.05; length 0.90 mm; 60×.
9. *Limnocythere sanctipatricii* Brady & Robertson (1869); adult male right valve; PE-96-2 – 12.40-12.45; length 0.88 mm; 72×.
10. *Limnocythere sanctipatricii* Brady & Robertson (1869); adult female left valve; PE-96-2 – 12.15-12.20; length 0.86 mm; 66×.
11. *Darwinula stevensoni* Brady & Robertson (1870); adult female left valve; PE-96-2 – 12.05-12.10; length 0.75 mm; 67×.
12. *Herpetocypris reptans* Baird (1835); adult female right valve; PE-96-2 – 12.70-12.75; length 2.03 mm; 39×.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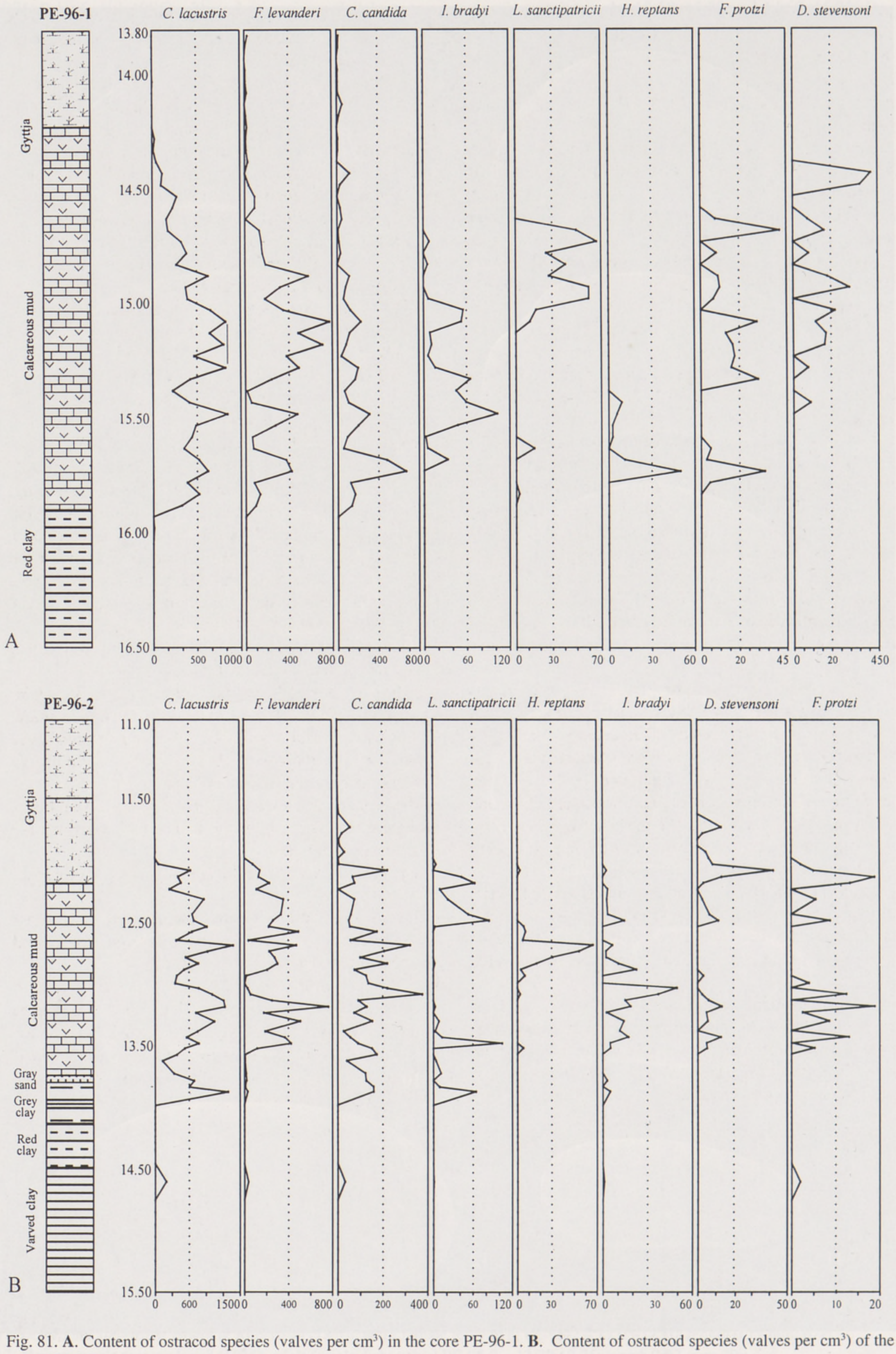


Fig. 81. **A**. Content of ostracod species (valves per cm^3) in the core PE-96-1. **B**. Content of ostracod species (valves per cm^3) of the core PE-96-2.

is interrupted between 12.55-13.15. *H. reptans* (max 67 valves per 100 cm^3 of sediment; Fig. 81-B) occurs mainly in the layer between 12.50 and 13.10 m, though it is present also in some random samples above and below this level. *I. bradyi* (max 50 valves per 100 cm^3 of sediment; Fig. 81-B) seems to be present in most of the ostracod containing samples. *D. stevensoni* (max 43 valves in 100 cm^3 of sediment; Fig. 81-B) appears first in the sediments at 13.54 m and disappears at 11.70 m in the section, having its maximum density at the level of 12.05–12.10 m. Noticeable is the interval 12.50–12.90, where *D. stevensoni* is absent. As in PE-96-1 *F. protzi* (max 19 valves per 100 cm^3 of sediment; Fig. 81-B) is distributed into two separated intervals: 12.97–13.54 and 12.00–12.50.

The distribution of ostracodes in L. Peipsi is very closely related to the sediment type and its sand fraction content (Figs. 80-A, B). Thus, the varved clay and gyttja contain very few ostracods, while the calcareous mud and grey clay may contain more than 1000 valves per sample and the density of ostracod valves may reach 3500 valves per 100 cm^3 of sediment. Sandy samples usually do not contain ostracods, although their appearance level in both sections is distinctly sandy. The faunal diversity is high (up to 7 species per sample) in calcareous mud and grey clay, and low (1–2 species) in varved clay and gyttja.

The anomalous level in varved clay of PE-96-2 (14.50–14.70), featuring with high faunal diversity and ostracod content, could be associated with slides on the bottom of lake and should be treated with caution (Hang, pers. comm.).

The considerably abundant distribution of ostracods in both sections begins in a sandy interval (Figs. 80-A, B), although generally large sand fraction content, caused by fast sedimentation rate, seems to dilute the actual microfaunal content. These sandy levels in both cores (in PE-96-2 just beside the sand layer) may refer to the period of Lake Small Peipsi (Hang, pers. comm.), when the lake was supposed to have been much shallower than today (Miidel & Raukas 1997). The abundance of *C. lacustris*, *C. candida*, *F. levanderi* and *I. bradyi*, the species of cold water preference, in shallow water may refer to partial groundwater feed in the lake this time. The decreasing sand fraction content refers to the respective decrease in sedimentation rate and increase of water level.

Three levels of high ostracod abundance in both sections (Figs.80-A, B) tempt to correlate cores to each other on these peaks, but looking closer at the assemblage structure and the placement and pattern of these peaks, these seem to be rather occasional coincidences. Therefore, the only correlation of sections was made on sediment type, i.e. sedimentation of calcareous mud in both cores was presumed to begin simultaneously.

Three most abundant species (*C. lacustris*, *F. levanderi*, and *C. candida*) are very tolerant to different ecological conditions and are really widespread along the cores (Fig. 81-A, 81-B). In terms of palaeoecological conditions, more informative are the rarer occurring species. The presence of *I. bradyi*, which is believed to be a very shallow water species, in both sections almost throughout the entire interval of calcareous mud, refers to a water depth not much deeper than 4 m during the time of carbonate precipitation.

The restricted presence of another shallow water species, *H. reptans*, coincides with the absence of the deeper water species *D. stevensoni* in both cores, suggesting even shallower water than 4 metres (less than 2 m according to Zubowicz 1978, 1983) and possible warming accompanied by increased productivity. The shallowing referred by the presence of *H. reptans* has probably occurred at least twice. Once, during about the presumable Lake Small Peipsi period, marked by the presence of *H. reptans* in PE-96-1 and possible sedimentation break in PE-96-2, closer to the coastline. Later, when *H. reptans* was present in PE-96-2, the water level in PE-96-1 could have already been deep enough, that the minor shallowing did not affect the ostracod assemblage in the middle part of the lake. Several changes in water level of L. Peipsi during the Holocene have also been reported in the previous studies (Miidel & Raukas 1997).

Summing up the results of the study, it is possible to draw some general **conclusions:**

1. In the two cores studied from L. Peipsi, 8 ostracod species were found: *Darwinula stevensoni*, *Candona candida*, *Fabaeformiscandona levanderi*, *Fabae-formiscandona protzi*, *Cytherissa lacustris*, *Limnocythere sanctipatricii*, *Ilyocypris bradyi and Herpetocypris reptans*.
2. The ostracod density depends on the type of sediment - calcareous mud is fossiliferous, while clays and gyttja are almost unfossiliferous.
3. The only correlation between cores was performed on sediment type.
4. The spread of ostracods starts in shallow water with cold water preferring species. This may refer to possible partial groundwater feed at this time in the lake.
5. The continuous presence of *I. bradyi* in both sections refers to water depth not very much more than 4 metres during the deposition of calcareous mud in L. Peipsi. Two levels featuring with associations, where *H. reptans* is present and *D. stevensoni* is absent may refer to even smaller water depths, warmer water mass and increased productivity.

3.2.6. MINERAL RESOURCES

Lake Peipsi with a variety of shore types offers good recreational possibilities and favourable bathing conditions. It also holds remarkable reserves of sanative mineral wealth, including different kinds of mineral water and curative mud (Fig. 82).

Mineral water with the content of total dissolved solids more than two grams per litre evidently occurs throughout the lake basin (Verte 1961, 1963). During the last decades, it has been found at Värska and Mehikoorma and in many other places in the vicinity of the lake basin, including Kuningaküla, Alatskivi, Põlva, Võru, Pechory, *etc.* (Vingisaar 1978). Owing to significant variations in the chemical composition, different kinds of mineral water can be used for treating different diseases: Põlva, Pechory and Värska III — chronic arthritis and diseases of the peripheral nervous system; Värska II and IV, Võru and Alatskivi — chronic diseases of the digestive tract; Värska I — diseases of the digestive tract, biliary duct, gall-bladder and liver. Mineral water is also used to cure gynaeocological and several other diseases (Vingisaar 1978).

In Estonia, the commercial exploitation of mineral water was started in 1968 at Värska. The mineral water derived from four aquifers there (259–314 D_{2-1}, 451–500 O – Є, 520–535 Є –V Voronka and 575–595 m Є –V Gdov), can be bottled for drinking and used in bath for treating different disorders. Currently, the Cl-Na-Ca water with the content of total dissolved solids 2.0–2.2 grams per litre comprised in the Ordovician-Cambrian aquifer system at Värska, is bottled (Vallner 1997). During the bottling, it is often enriched with carbon dioxide.

Large sections of the floor of L. Peipsi *s.s.*, L. Lämmijärv and L. Pihkva are covered with fine-grained and organic rich sediment, termed curative mud (Fig. 57). Its reserves in L. Peipsi are practically unlimited. Detailed studies which covered only one third of Värska Bay (Tsetshladze 1970, Tassa 1976, Pirrus & Tassa 1981) revealed the existence of a 4.12-metre-thick layer of mud with the water content 60% (Fig. 83). The reserves of mud in this part of the bay are estimated at 3.52 million m^3 and in the whole bay at 44.5 million m^3. According to geological mapping, muds with a high content of organic matter extend from Värska to the Island of Kolpino. For medicinal purposes, some 600–800 m^3 of curative mud is produced from Värska Bay annually.

Large reserves (at least 6 million m^3) of curative mud occur in L. Lahepera (Paap *et al.* 1981) and L. Umbjärv, and, according to geological mapping (Kajak *et al.* 1974), in several semiclosed bays (*ca* 30 million m^3 in Raskopel' Bay, *ca* 7 million m^3 at Gorodets in the mouth of the Zhelcha River, *etc.*). Our studies showed that the area at Salu Islet and some kilometres to the north in L. Lämmijärv is also rich in curative mud (Fig. 47).

According to the classification used in Estonia (Ramst 1992), lake mud is divided into biogenic, clastic and mixed varieties. In a deposit, acknowledged as lake mud, organic matter must form at least 15% of dry mass. The deposits of lake mud with a high content of carbonaceous material (dry solid matter makes up at least 40% of CaO) are registered as a separate mineral resource - lacustrine lime, which is not found in L. Peipsi.

The main constituents of lake mud are organic matter and sand, silt and clay particles, which have been carried into the lake. Silt fractions are most common, accounting for 40–70%. The colour of lake mud varies from light-beige to black (mainly greenish-grey) and consistence between jelly-like and plastic.The average water content is 90%, after drying the colour would change. The chemical composition of mud is variable, depending on the grain-size composition and formation peculiarities (Table 17).

The organic part consists of carbohydrates, humic acids, cellulose and hemicellulose, bitumens and nonhydrolized residue. Several bioactive components (vitamins, hormones, antibiotics) and different microelements are also present (Pirrus & Tassa 1981). Lake mud can be used for agricultural purposes and for producing fertilizers. In all kinds of sapropel, which we studied jointly with the Latvian agrochemist Vimba (Vimba *et al.* 1978), the dry matter content was medium or high (30–80%). The content of carbonaceous matter was remarkable; the content of CaO ranged from 10–15% in L. Lahepera to 22% in L. Umbjärv,

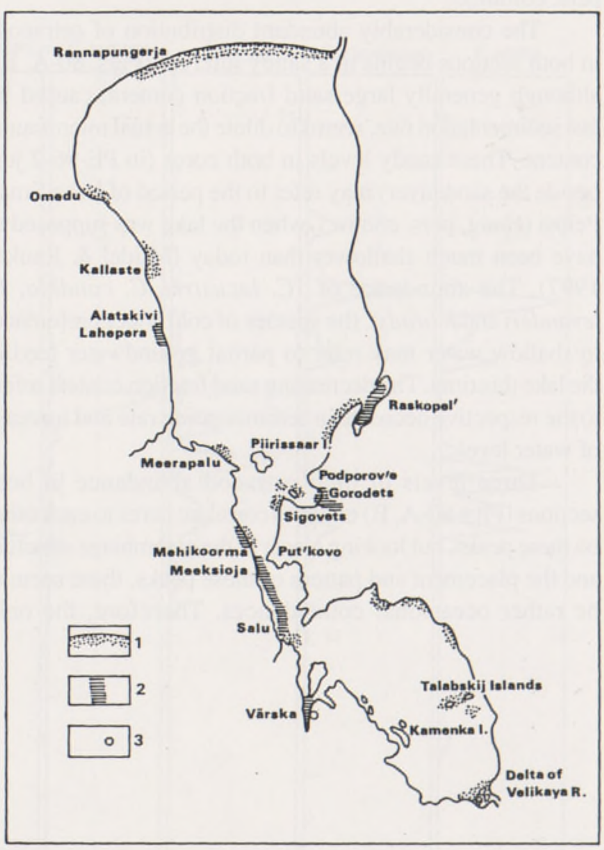


Fig. 82. Mineral resources of L. Peipsi: 1 – sand deposits; 2 – curative mud deposits; 3 – mineral water deposit.

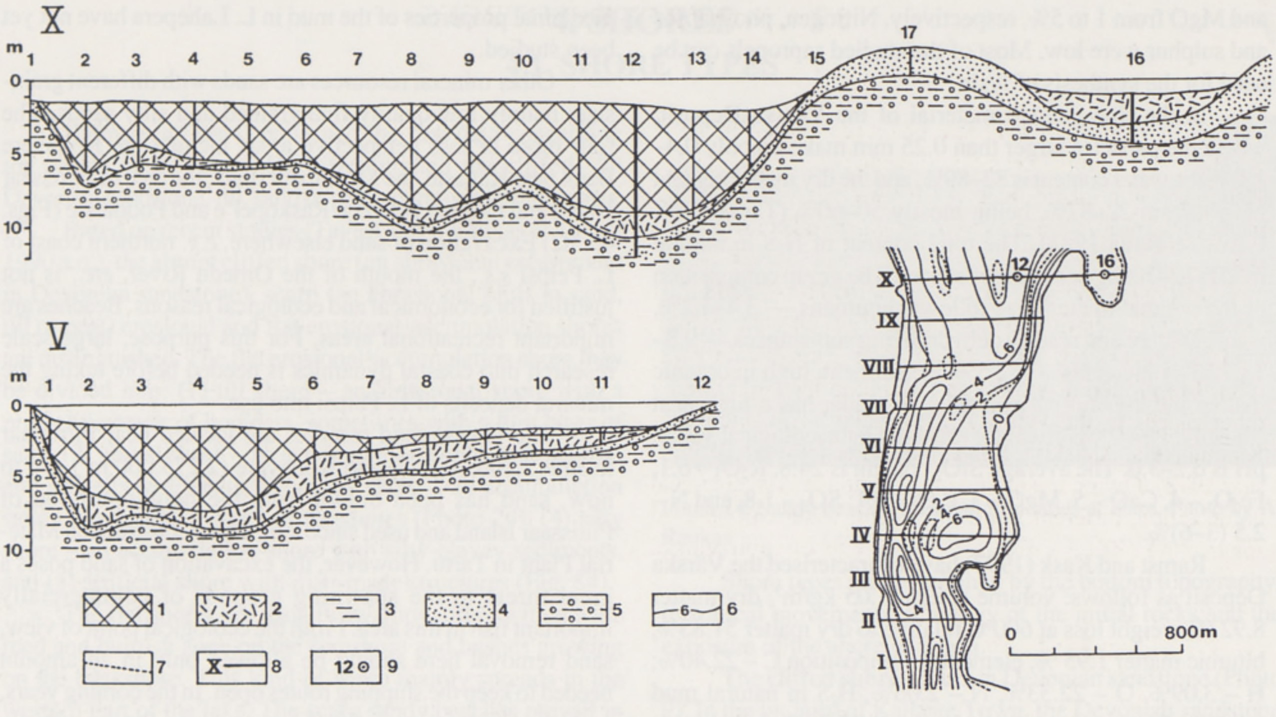


Fig. 83. Sections showing the geology and thickness of curative mud in the Värska Deposit (after Pirrus & Tassa 1981): 1 - gyttja; 2 - peat; 3 - silt; 4 - sand; 5 - till; 6 - isopachyte (m) of curative mud (gyttja); 7 - boundary of the distribution of peat under gyttja; 8 - geological sections; 9 - palynologically studied boreholes.

Table 17. Chemical composition of curative mud in Värska Deposit (after Pirrus & Tassa 1981)

Number of profile and sample site	Organic matter, %	Dry matter, %	% of dry mass							
			SiO_2	R_2O_3	Fe_2O_3	CaO	MgO	P_2O_5	SO_3	N
II 8	54.82	45.18	27.16	7.14	3.23	4.14	1.23	0.17	1.50	2.43
II 8	65.88	34.12	15.24	5.26	4.08	6.11	1.03	0.10	2.14	3.16
II 3	51.34	48.66	23.92	3.17	-	-	-	-	-	-
III 7	58.68	41.32	24.01	8.61	5.11	3.83	0.89	0.10	1.34	2.08
III 7	65.14	34.86	17.92	6.15	4.13	5.24	1.16	0.07	2.13	3.05
IV 1	15.59	84.41	-	-	-	-	-	-	-	-
IV 6	53.46	46.54	28.78	4.02	-	-	-	-	-	-
IV 9	51.77	48.23	27.69	8.15	4.87	4.13	1.19	0.18	2.75	1.98
IV 9	73.43	26.57	15.07	4.56	2.32	3.24	0.48	0.06	1.49	3.52
IV 10	36.41	63.59	45.06	5.02	-	-	-	-	-	-
V 3	53.19	46.81	23.81	9.34	5.19	5.03	1.17	0.13	1.97	2.18
V 4	52.28	47.72	23.17	4.08	-	-	-	-	-	-
VI 7	54.31	45.69	23.17	10.32	4.39	4.16	1.11	0.15	2.06	2.53
VI 6	70.46	29.14	18.09	3.16	-	-	-	-	-	-
VII 6	54.35	45.65	23.13	3.94	-	-	-	-	-	-
VII 5	35.75	51.17	28.54	4.32	1.79	15.24	1.04	0.17	1.25	1.79
X 4	45.47	54.53	32.86	11.27	5.69	3.39	0.97	0.21	1.46	2.31

and MgO from 1 to 5%, respectively. Nitrogen, phosphorus and sulphur were low. Most of the studied sapropels can be used for the synthesis of lysin.

In the fine-grained material of the Värska Deposit, where the particles larger than 0.25 mm make up only 0.4–1.1%, the water content is 82–89%, and the dry matter content ranges from 22–87%, being mostly 50–60% (Tassa 1976, Pirrus & Tassa 1981). The total content of H_2S in natural mud is 12–70 mg per 100 g of extract. The group composition of the organic matter is as follows: bitumens — 3.4–4.2%, water-soluble and readily dehydralizing sub-stances — 9.8–13.2%, humic acids — 32.5%. A sediment, high in organic matter and containing hydrogen sulphide, has a high heat capacity which significantly raises its balnaeological value. pH is 6.2–3.0. The average SiO_2 content is 24.8, R_2O_3 – 6.1, Fe_2O_3 – 4, CaO – 5, MgO – 1, P_2O_5 – 0.1; SO_3 – 1.8, and N – 2.5 (3–6)%.

Ramst and Kask (1997) have characterised the Värska Deposit as follows: volume weight 1.05 kg/m^3, dry matter 8.92 %, weight loss at 600°C relative to dry matter 51.83%, bitumic matter 1.95 %, elemental composition C – 22.40%; H – 3.09%, O – 22.33%, N – 2.06%, H_2S in natural mud 0.096%; free humic acids in dry matter 1.74%, fulvic acids 1.87% and hymatomelanic acids 1.38%.

From the agricultural point of view, the sapropels of L. Lahepera are much more promising than those of the largest, Värska Deposit. In the deposits of L. Lahepera and Värska, the content of C is 54.5 and 24%, H – 7.7 and 3.2%, and N – 5.1 and 2.2%, respectively. Unfortunately, the medicinal properties of the mud in L. Lahepera have not yet been studied.

Other mineral resources are sands with different grain-size; mainly low-quality fine-grained and silty sands. The four more or less perspective areas include the Kodavere Bank near Kallaste, the Meerapalu Bank, the area of Piirissaar Island and the area between Raskopel'e and Podoles'e (Figs. 45, 82) Excavation of sand elsewhere, *e.g.* northern coast of L. Peipsi *s.s.*, the mouth of the Omedu River, *etc.*, is not justified for economical and ecological reasons. Beaches are important recreational areas. For this purpose, large-scale research into coastal dynamics is needed before taking the mineral deposits of L. Peipsi into use.

The largest sand reserves are located near Piirissaar Island (2,100,000 m^3) and Kodavere (2,275,000 m^3). Up to now, sand has been drawn from the neighbourhood of Piirissaar Island and used since long at the Construction Material Plant in Tartu. However, the excavation of sand poses a great threat to the spawning grounds of commercially important fish in this area. From the ecological point of view, sand removal here should be allowed only in an amount needed to keep the shipping routes open. In the coming years, with the renovation of the shipping route Praaga – Piirissaar – Pskov, big quantities of sand have to be removed. This sand should be deposited on the land and used in future for meeting the low demands of local users.

The potentialities of L. Peipsi as a transport way are far from being exhausted. The share of road-metal haulage can be increased to cover the needs of Tartu and Pskov.

4. SHORES

4.1. SHORE TYPES

Helmersen (1864) presented the first data concerning the coasts of L. Peipsi. He differentiated the cliffed- (in Devonian sandstone, till, sand or peat), dune- and peaty coasts.

Based on recent studies (Tavast 1984, Raukas & Tavast 1990 a.o.), the abrupt cliffed shore (an abrasional escarpment in Devonian sandstone), scarp (an abrasional bluff in sand, till or peat) erosional and flat erosional-accumulation shores are distinguished. The flat erosional-accumulation shore may be divided into: (1) till shore – an abrasional shore with a protective cover of boulders, sometimes with a thin layer of sand, (2) gravel shore – an accumulation shore with pebbles and gravelly sediments, (3) sandy shore – an accumulation sandy beach with a ridge of foredunes (Photo 24), (4) silty shore – an accumulation shore with silty-clayey sediments, and (5) artificial shore with man-made structures (Fig. 84).

The flat shores are usually swampy, often with a wide reed and bulrush zone on the foreshore and bushes growing on the backshore. This kind of shore mainly spreads in the western part of the lake. The scarp sandy beaches prevail in the northern part. Like the majority of lakes in the Northern Hemisphere (Klinge 1889), Peipsi has a more open eastern and a more swampy and overgrown western shore. Due to the prevailing southwesterly and westerly winds, the active erosion-accumulative or erosional shores are spread in the northern and eastern parts of the lake. The swampy shores overgrown with bushes, bulrush and reed are characteristic of the western and southern parts of the lake (L. Lämmijärv and L. Pihkva).

Photo 24. Sandy beach with small foredunes at Kuru. Photo by A. Raukas

Shore types are determined by the bottom topography, tectonical movements, lithology of the initial rocks and the exposure of the shore to winds.

The **cliffed shore** occurs in Devonian sandstone (Photo 9). In the vicinity of Kallaste Town, the Devonian sandstone cliff has a height of 8 m and is overlain by a 1.5-m-thick layer of till and a thin layer of sand. The sandstone outcrops, covered with till and sand, are also encountered near Skamya and Polichno villages on the eastern coast of L. Peipsi *s.s.* Numerous erratic boulders and wave-cut notches occur at the foot of the cliffs.

The scarp shore in loose Quaternary deposits is widespread. Bluffs may be eroded into till or into different types of sand (kame, dune), seldom into peat.

The **scarp till** shore (Fig. 85 - I) is spread on the eastern coast (Podlip'e, Klenno, Vetvenik) of the lake. It consists of scarps, from some hundred metres up to several kilometres in length. The layer of sand with a thickness of 5–100 cm covers the till (Klenno). A stony floor with numerous boulders and a thin layer of sand and pebble occurs at the foot of the scarp in the foreshore (Photo 25). The area in front of dead till scarp is covered with sod and frequently bushes are spread. Occasionally, there is a narrow strip of sandy beach on the shoreline, *e.g.* on the northern coast of the Talabskij Islets.

The **erosional sandy scarp beach** (Photo 26, Fig. 85 - II) is widespread on both shores of the lake. The bluff has been eroded into kames (Podporov'e), ancient dunes or beach ridges (northern coast). The active sandy beach begins from the Kauksi Brook and runs as far as the outflow of the Narva River. One or several bluffs have developed on the ancient dune sands (Photo 26). Shrubs seldom grow in front of the bluff; there is neither reed nor bulrush on the foreshore. The dunes are usually 5–6 m high. At the river mouths, the dunes are higher, *e.g.* up to 13 m at Smolnitsa. In some places they form several parallel ridges.

The **peaty scarp shore**, 70 cm high and 1.5 km wide, occurs, for instance, near Podoleshche Village (eastern coast).

Flat shores may be located on till, sand, silt or peat. The flat till shore (Fig. 85 - III) is not widespread. Erratic boulders,

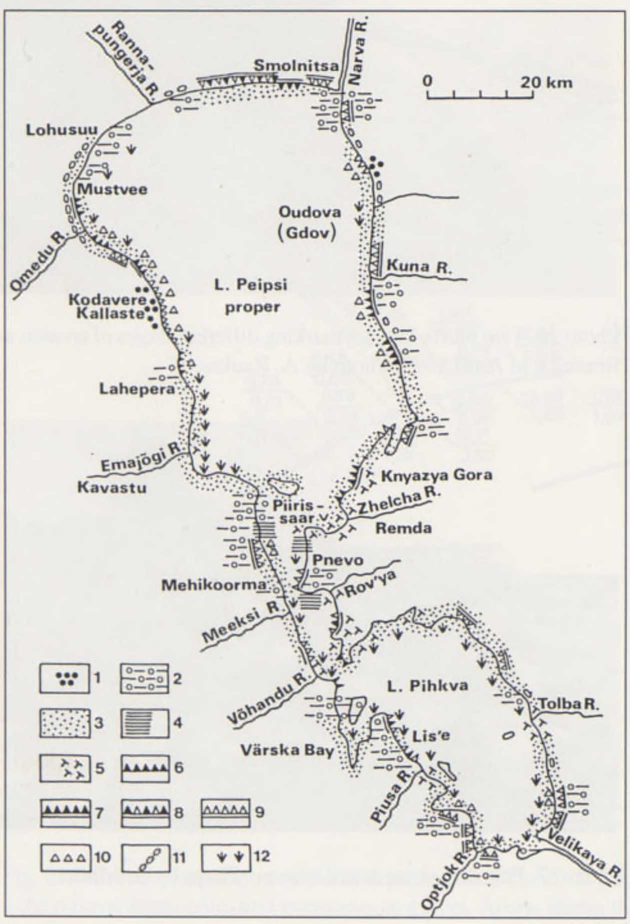
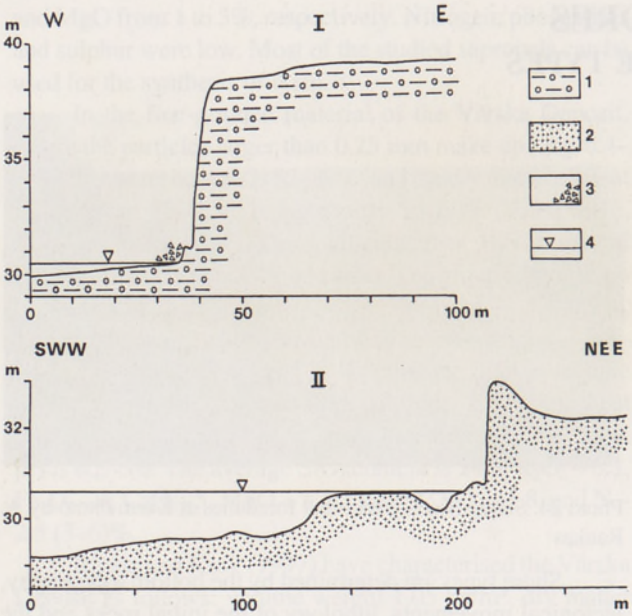


Fig. 84. Lithology of the shore types of L. Peipsi: 1 – Devonian sandstone; 2 – till; 3 – sand; 4 – clay; 5 – peat; 6 – cliff; 7–9 – bluffs in Quaternary deposits: 7 – active; 8 – passive; 9 – dead; 10 – erratic boulders; 11 – beach ridges; 12 – bulrush and reed zone.



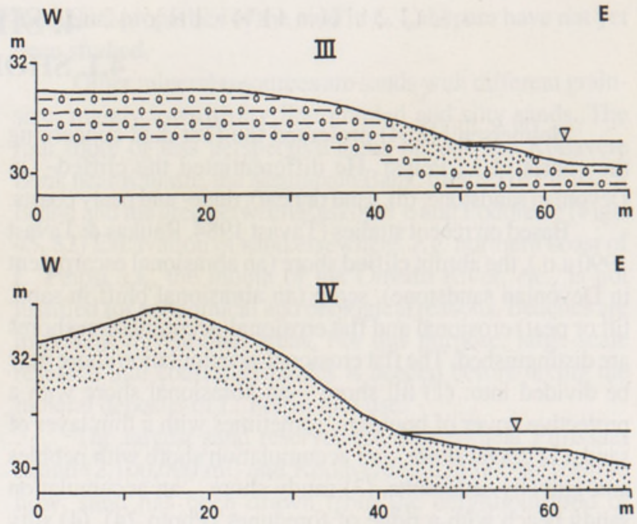


Fig. 85. Geological sections of the most characteristic shore types: I – bluff in till at Krivsk; II – scarp in sand at Baglitsy; III – flat till shore near Lyubnitsa; IV – flat sandy shore at Tammispää. Lithology: 1 – till; 2 – sand; 3 – boulders; 4 – water level.

washed out from till, are usually scattered all over the shore. Boulders have often been carried to the shore by the lake ice. Bulrush and reed grow on the foreshore and shrubs on the backshore.

The flat **swampy sandy** and **silty** shores (Photo 27) have much in common. The backshore is covered with bushes, and the bulrush and reed grow on the foreshore. A typical swampy sandy shore extends from the mouth of the Avijõgi River to the Town of Mustvee. The foreshore is bordered by a low beach ridge, 0.5–1.5 m in height. The upper part of the ridge has been blown into small dunes. A low swampy sandy shore, overgrown with shrubs, also stretches from the Avijõgi River to the Kauksi Brook. Bushes cover the shore sporadically. The foreshore is generally flat, and the zone of bulrush or reed is 40–50 m wide. One or two low beach ridges border the foreshore.

The **flat peaty** shore is usually covered with willow bushes and common alder on the backshore. A wide zone of bulrush spreads on the foreshore. A most typical flat peaty shore is encountered in the surroundings of the mouth of the Emajõgi River, where reed and bulrush form an extensive and very dense zone. Irregular neotectonic movements, causing the lake waters to advance to the south and southwest, favour the development of this kind of shore.

To sum the above up, it may be said that due to the prevailing westerly and southwesterly winds erosional-accumulative shores dominate in the eastern and northern parts of the lake, and accumulative shores on the western coast of the lake.

Photo 26. Two bluffs in sand, marking different stages of erosion at Remniku in April 1997. Photo by A. Raukas.

Photo 25. Till shore with a protective cover of boulders and with bluff on the backshore at Klenno. Photo by A. Miidel.

Photo 27. Flat silty shore near Lohusuu. Photo by A. Miidel.

4.2. GRANULOMETRY AND MINERALOGY OF SHORE DEPOSITS

The shore deposits of L. Peipsi show wide variations in the mineral composition, often due to the granulometry of the sediments. Silt and sandy silts are richer in feldspars, micas and other minerals characteristic of calm sedimentological conditions with limited transport of material.

The grain-size (Fig. 86) and mineralogy (Fig. 87) of shore deposits have been studied in particular detail in the northern part of the lake. In the northwestern part of L. Peipsi the shore deposits are mostly represented by sand. Sand is distributed on the backshore, and till overlain by some tens of centimetres up to several metres of sand occurs on the foreshore. Well-sorted (the coefficient of sorting is usually 1.2–1.3) fine and medium-grained sand (Md 0.15–0.47 mm) is spread on the stretch of the beach between the Avijõgi River and the outflow of the Narva River. Fine, well sorted sand (Md 0.15–0.20 mm; S_0 1.26–1.14) occurs in the coastal zone between Avijõgi and Kauksi Brook and at the outflow of the Narva River. Due to river input and erosion of the backshore, medium-grained, less sorted sand (Md 0.38–0.44 mm; S_0 1.29–1.7) is encountered in an area east of the Kauksi Brook.

The mineral composition of the light subfraction (density lower than 2.89 Mg/m^3) is dominated by quartz (84–94%) and feldspars (7–16%). Other minerals are present in amounts less than 1%. In the heavy subfraction (density higher than 2.89 Mg/m^3), amphiboles (max 75%), magnetite and ilmenite (up to 49%) and garnet (up to 36.4%) dominate. The quantity of valuable components, mainly zircon, is rather low (1.2–5.6%) in the heavy subfraction. Less resistant minerals are practically absent. Micas-chlorites and pyroxenes are present in small quantities (about 4%).

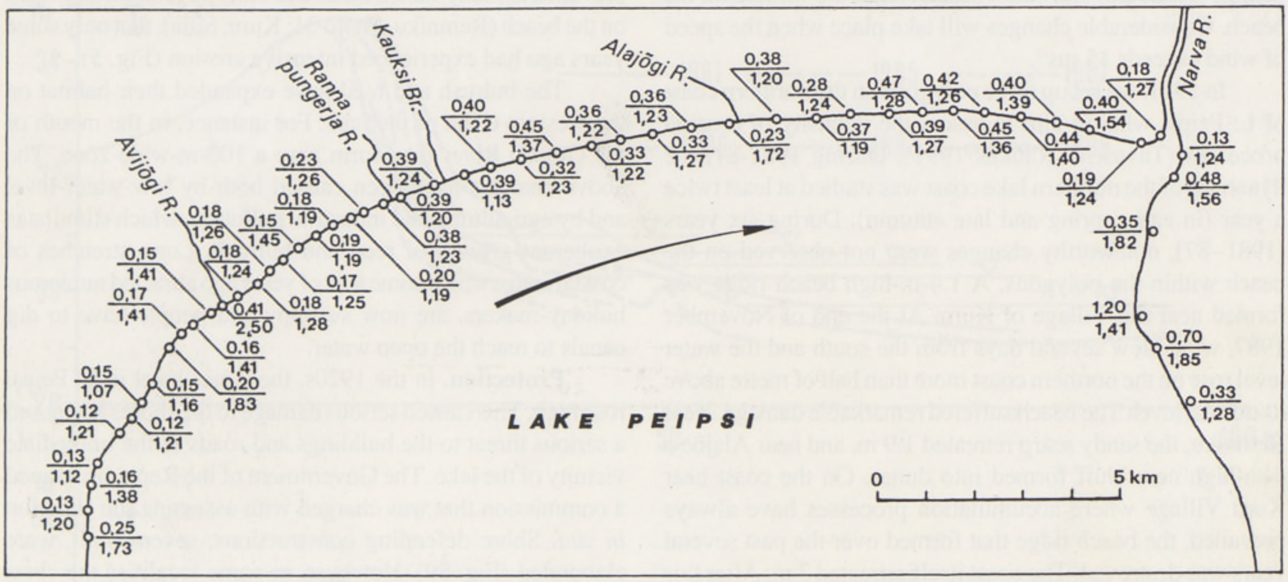


Fig. 86. The grain-size coefficients of the beach deposits in the northern part of the lake (after Tavast & Raukas 1996): denominator – the coefficient of sorting, numerator – the median. Arrow shows the direction of the drift.

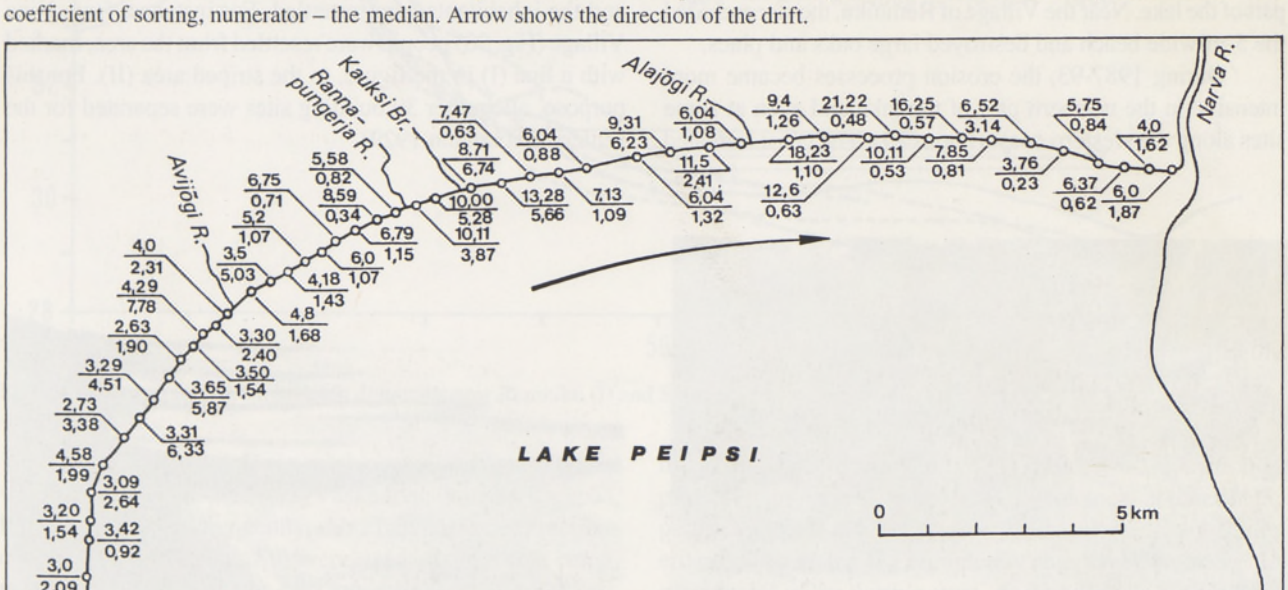


Fig. 87. Mineral composition of beach deposits (after Tavast & Raukas 1996): denominator – the ratio of quartz and feldspars, numerator – the ratio of amphiboles and pyroxenes to garnet. Arrow shows the direction of the drift.

4.3. SHORE EROSION AND PROTECTION

Erosion. Damages have always been greater on the eastern and northern coasts of L. Peipsi, which are more open to winds than the western coast. During the high water level, the effect of deep and short waves on the coast and buildings is considerable. G. Helmersen wrote already in 1864 that over a period of 20 years the water level had been high in L. Peipsi and, as a result, the coast suffered remarkable damage. K. Baer (1860) and G. Helmersen (1864) supposed that the high water level was caused by the clearance of forests and land improvement. They suggested to lower the water level in the lake by one metre. Many authors touched upon this subject at the beginning of the century.

The profile of equilibrium is formed at the coast by winds with a normal velocity of 4–5 ms^{-1}. When the speed of the wind is more than 6–7 ms^{-1}, the bluffs (50–60 cm high), festons (Photo 28) and small beach ridges are formed on the beach. Considerable changes will take place when the speed of wind exceeds 15 ms^{-1}.

In 1981, we set up three polygons on the northern coast of L. Peipsi with an aim to assess the intensity of erosion processes (Tavast & Raukas 1991). During 1981-87, the dynamics of the northern lake coast was studied at least twice a year (in early spring and late autumn). During six years (1981–87), noteworthy changes were not observed on the beach within the polygons. A 1.4-m-high beach ridge was formed near the Village of Kuru. At the end of November 1987, wind blew several days from the south and the water level rose on the northern coast more than half of metre above its normal level. The beach suffered remarkable damage. Near Remniku, the sandy scarp retreated 1.9 m, and near Alajõe a 4-m-high new bluff formed into dunes. On the coast near Kuru Village where accumulation processes have always prevailed, the beach ridge that formed over the past several years was destroyed. The coast itself retreated 3 m. After this storm, intensive erosion was registered over practically the entire length of the northern coast, excluding the northwestern part of the lake. Near the Village of Remniku, the storm eroded the 5-m-wide beach and destroyed large oaks and pines.

During 1987-93, the erosion processes became more intensive in the northern part of the lake and even at some sites along the western coast (Photo 29). The coast retreated several metres during those years. New wave-cut notches were formed in the Devonian cliff at Kallaste. An old layer of peat (Photo 30) was exposed in several places on the north coast of the lake. Both water erosion and hummocky ice affected the coastal processes. In some places, *e.g.* in the dune field at Rannapungerja, the zone of bulrush and bushes was slightly damaged by the hummocky ice in winter time.

The western coast, which is protected from the prevailing westerly and southwesterly winds, is swampy. Bushes grow on the shore, and the beds of bulrush and reed on the foreshore may be 40 or even more metres wide as, for instance, in the mouth of the Omedu River.

During the low stand of water in 1995-96, the rate of erosion decreased. There were practically no changes within our polygons in the northern part of the lake. Accumulation prevailed at many sites, bushes and other vegetation inhabited on the beach (Remniku, Photo 31; Kuru, Silla), that only some years ago had experienced intensive erosion (Fig. 5).

The bulrush and reed have expanded their habitat on the western coast of the lake. For instance, in the mouth of the Omedu River they form now a 100-m-wide zone. The above changes have been caused both by low water-level and by agricultural and industrial pollution, which stimulates exuberant growth of reed and bulrush. Long stretches of coastal areas, which some tens of years ago attracted numerous holiday-makers, are now swampy and people have to dig canals to reach the open water.

Protection. In the 1920s, the water level in L. Peipsi rose high. This caused serious damage to the shores and posed a serious threat to the buildings and roads in the immediate vicinity of the lake. The Government of the Republic formed a commission that was charged with assessing the situation *in situ.* Shore defending constructions, seven in all, were elaborated (Fig. 89). However, in some localities the coast had already suffered thus great damage that the foundation of shore protecting constructions was considered inexpedient and the inhabitants were resettled. For instance, from Nina Village (Fig. 90) people were resettled from the area, marked with a line (I) in the figure, to the striped area (II). For this purpose, altogether 30 building sites were separated for the settlers (Vichmann 1929).

Photo 28. Festons near Kauksi. Photo by A. Miidel.

Photo 29. Cliff at Kallaste with broken stairs in June 1998. Photo by A. Miidel.

Photo 30. Beach at Remniku in May 1990 after heavy storms. A buried organic layer in the lowermost part of the section. Photo by E. Tavast.

Photo 31. During low stand of water the bushes and types of vegetation inhabited on the earlier erosional beach at Remniku in June 1998. Photo by A. Raukas.

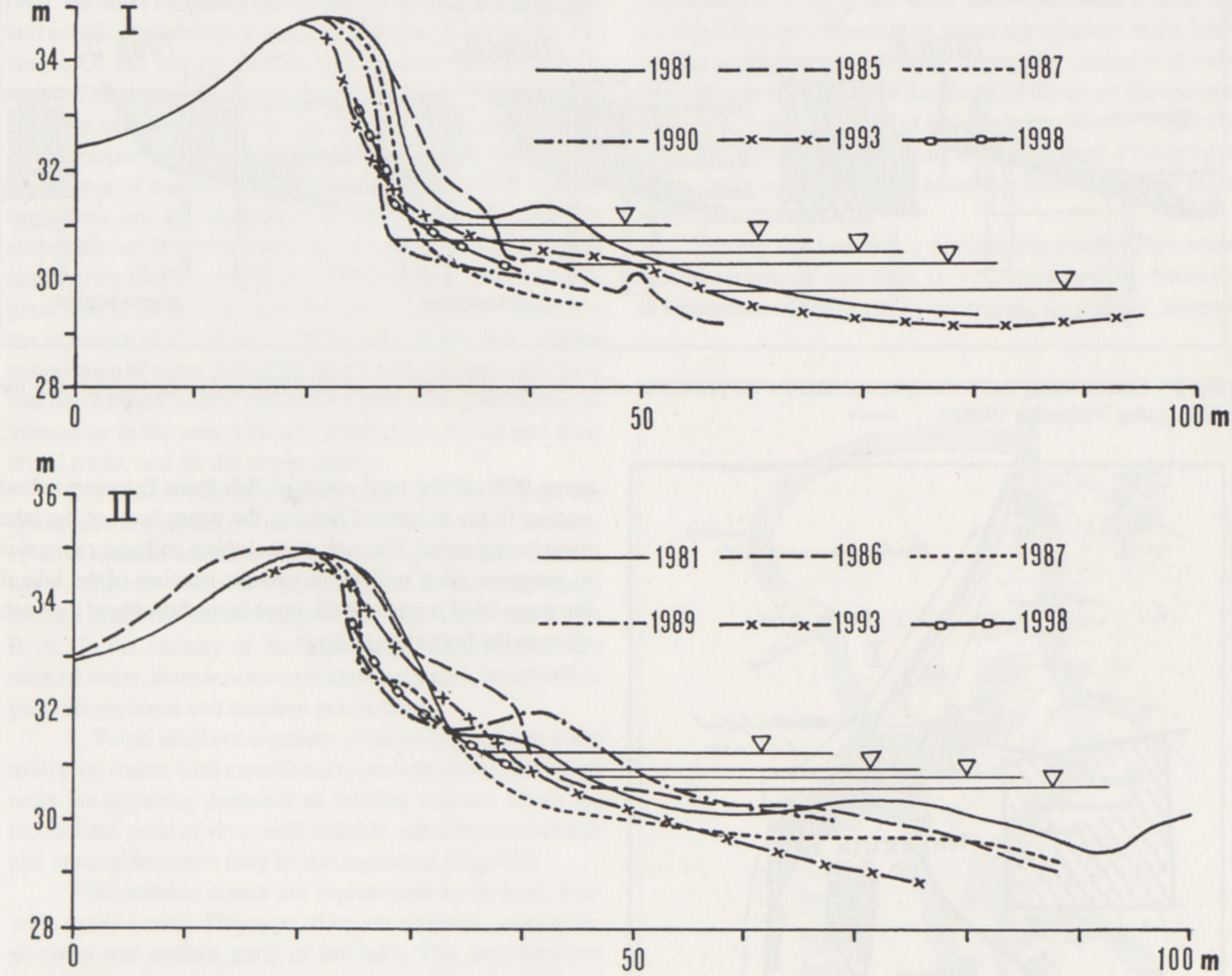


Fig. 88. Curves demonstrating beach dynamics near Remniku (I) and Silla (II).

In 1929, shore defense work started at 11 sites on the western coast of the lake — at Vasknarva, Omedu, Kasepää, Piirissaar, *etc*. Most frequently, shore defending constructions of types A and B (Fig. 89) were used. At the time being, these constructions do not exist any more.

The longshore drift carries the mass of sand, eroded during the storms from dunes, to the outflow of the Narva River. Even nowadays, some engineers have proposed to build defending constructions to protect the coast. This action is probably not necessary and from a geological and aesthetic point of view is not acceptable. In the more distant future erosion will stop here completely due to the neotectonic movements of the Earth's crust. As demonstrated in Chapter 1, the low and high water stands are regularly alternating. The low water level will be an unsuitable environ for fisheries. Peipsi ranks among the best fish lakes in Europe and gives

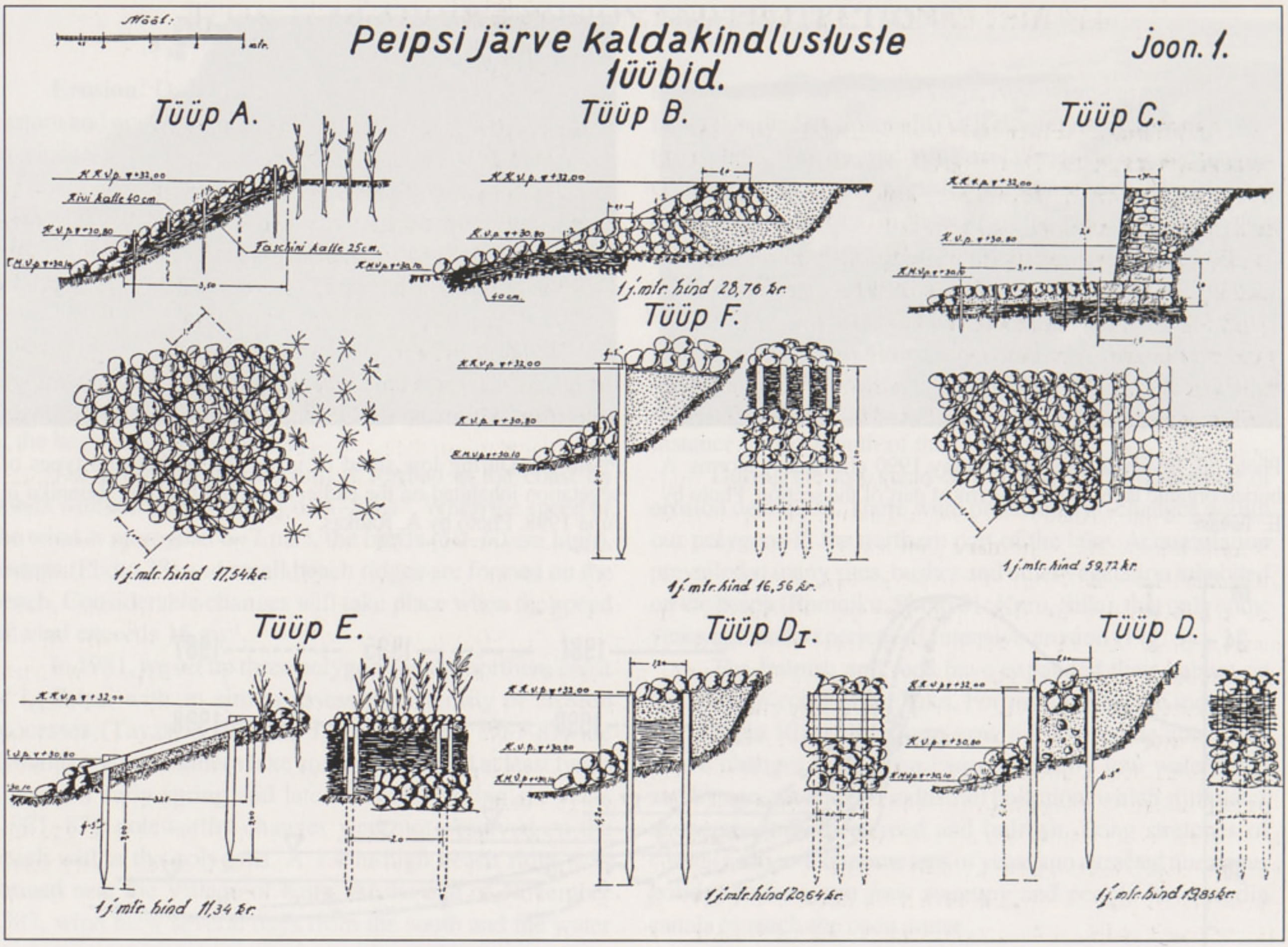
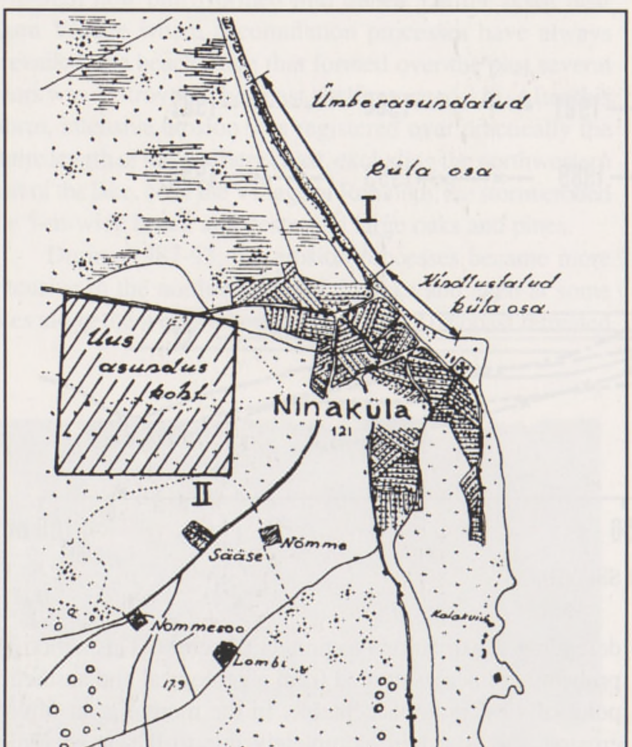


Fig. 89. Constructions that were recommended for the protection of the Peipsi coast at the end of the 1920s and in the beginning of the 1930s (after Vichmann 1929).



some 95% of the total catch of fish from Estonian inland waters. In the interest of fishing, the water level of the lake must be regulated. The task of geologists and geographers is to prognose what will happen on the beaches of the lake if the water level remains at the same level throughout the year, close to the long-time average.

Fig. 90. The plan of the replacement of the most endangered buildings on the lake coast near Nina Village in the late 1920s and early 1930s (after Vichmann 1929). I – settlement site before replacement, II – new settlement site.

4.4. RECREATIONAL POSSIBILITIES IN THE COASTAL ZONE

In Estonia, there are relatively many areas suitable for recreational purposes. The Republic has some 3780 km of coastline indented by numerous bays, lakes, rivers, islands and islets. Its varied landscapes offer good possibilities for hunting and fishing, hiking and boating, mushroom and berry picking. However, despite the dense network of roads and good traffic conditions, about half of the holiday-makers still prefer traditional recreation areas which make up less than 0.59% of Estonia's territory. Large crowds of people cause serious damage to nature and disturb the work of transport and commodity offices.

Whether the coastal zone is suitable for recreational purposes or not, depends on several factors. These may be divided into natural or primary, and artificial or secondary ones, the latter including the vicinity of industrial enterprises and mines, availability of roads, catering establishments and shops. Of the natural factors, the greatest significance is attached to geological-geographical conditions: the geological structure of the coastal zone, the fitness of the coastal zone for recreation (*i.e.* the availability of sandy beaches), the occurrence of curative mud and mineral water. But equally important are the hydrogeological conditions, and the suitability of the area from the standpoint of geological engineering (for the construction of buildings and roads). A great role is played by local climatic conditions, especially the exposure of the shore to the winds and the sun, and the temperature of water. All artificial and some natural conditions can be changed within less than a generation. Besides, the vegetation in the area depends on the types of soil and their initial rocks, and on the water regime.

L. Peipsi is rich in curative mud. In Värska Bay, the predicted reserves are estimated at 44.5 million m^3. Besides, curative mud occurs in the Peipsi Basin together with mineral water. Mineral water may be produced in the area, which extends from the southern tip of the lake up to the Narva River. In the vicinity of the deposits of curative mud and mineral water, there is often a picturesque beach, which offers good recreational and curative possibilities.

L. Peipsi displays a variety of shore types, from scarp to sloping shores with transitional types between them, which meet the differing demands of holiday-makers. From the recreational point of view, well-suitable, suitable, less-suitable and unsuitable coasts may be distinguished (Fig. 91).

Well-suitable coasts are represented by at least 3-m-wide sandy beach. This type of beach abounds only on the northern and eastern parts of the lake. The area between Kauksi and Alajõe in the northern part of L. Peipsi has become very popular. The area owes its popularity to the picturesque landscape – sandy beach covered with fine pine forest and exposed to the sunshine, slightly tilting coastal slope furnishing an excellent place for bathing of children, much warmer water than in the sea.

Suitable areas include the slightly or moderately swampy sandy beaches and slightly sodden scarp morainic or cliff coast. It is not easy to move along the naturally picturesque, high and dry cliff, but the situation may be improved if to lay out stairs and other accesses. This kind of shore occurs on the western coast of the lake, between the Sassukvere Brook and L. Lahepera.

Swampy low shores, usually wet and overgrown with bushes, and with the beds of reed and bulrush on the foreshore, are less suitable for recreation. This kind of shore is mainly found between Lohusalu and Mustvee in the western part of the lake and in the eastern part of L. Pihkva.

Wet and strongly overgrown peaty and shingle coasts are unsuitable for recreational purposes, unless hunters or fishermen, who have to dig long ditches for pushing boats into the water, use them.

In planning recreational areas on the shore, attention must be paid to the peculiarities of the water regime. As is known, water-level fluctuations in the lake may be rather impressive (Ch. 1). If the water level rises even a little bit over the average (30 m a.s.l.), the areas adjacent to the lake will be immediately inundated. Especially extensive floods take place at the outflow of the Emajõgi River, on the eastern coast of L. Lämmijärv and at the whole length of L. Pihkva's coast. In 1956, 647 km^2 of land were inundated. Floods take place every year, especially extensive ones during the high water periods.

During the low water periods the coastal processes would decrease and this is accompanied by intense development of sod-forming processes. As a result, several

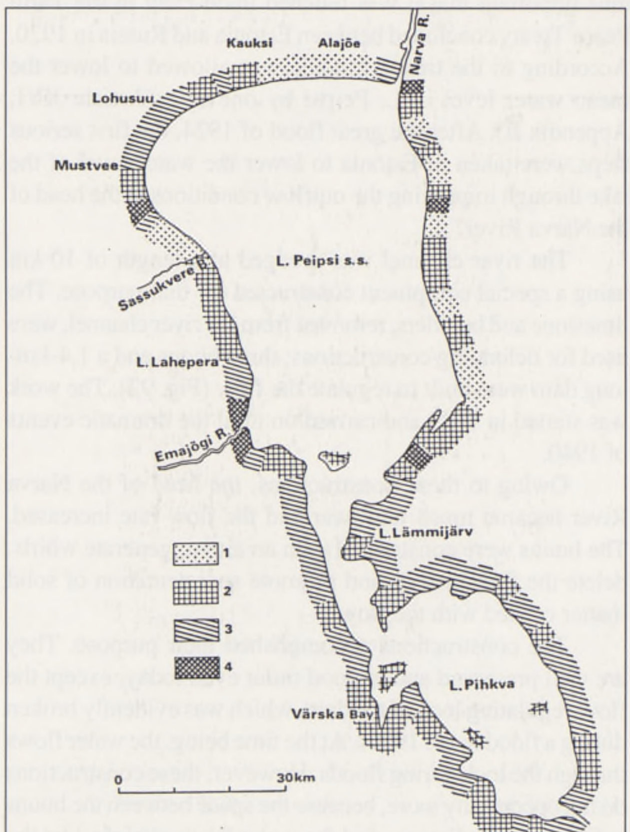


Fig. 91. Suitability of the coast for recreational purposes: I – well suitable; 2 – suitable; 3 – less suitable; 4 – unsuitable (after Raukas & Tavast 1989, with complements).

Photo 32. Till exposes on the coastal slope during low water period. Photo by A. Miidel.

fine sandy beaches may become less suitable for holiday-making. On the coastal slope till is covered by a relatively thin sandy layer. If the water-level drops, till may become exposed (Photo 32) instead of sand. With the formation of a new equilibrium profile, sand will accumulate again on the beach, but it takes several years.

From the above follows that the general state of L. Peipsi's coasts depends on natural factors and man's activities. Agricultural and industrial pollution must be reduced to protect and save the curative muds and mineral waters for medicinal purposes, and the water for drinking.

4.5. PROBLEMS RELATING TO THE REGULATION OF WATER LEVEL IN L. PEIPSI

As a result of the almost periodically recurring water level rises in L. Peipsi, extensive coastal areas amounting to several hundreds of hectares are inundated. In the past, the floods were a serious obstacle to the economic activities of local inhabitants in the surroundings of the lake. Unusually high water levels occurred in 1833, 1840, 1841, 1847 and 1867. Researchers of that time did not entirely agree as to the agents responsible for extremely high rises in the lake level. However, it was clear that the water level in L. Peipsi would progressively rise. In Estonia, the problem was considered thus important that it was touched upon even in the Tartu Peace Treaty concluded between Estonia and Russia in 1920. According to the treaty, Estonia was allowed to lower the mean water level in L. Peipsi by one foot (Article XVI, Appendix II). After the great flood of 1924, the first serious steps were taken in Estonia to lower the water level of the lake through improving the outflow conditions at the head of the Narva River.

The river channel was dredged at a length of 10 km using a special equipment constructed for that purpose. The limestone and boulders, removed from the river channel, were used for defending constructions; three buuns and a 1.4-km-long dam were built to regulate the flow (Fig. 92). The work was started in 1929 and carried on until the dramatic events of 1940.

Owing to these constructions, the head of the Narva River became much narrower and the flow rate increased. The buuns were constructed with an aim to generate whirls, delete the flow energy and promote sedimentation of solid matter carried with the flow.

The constructions accomplished their purpose. They are well preserved and in good order even today, except the flow-regulating lock of the dam, which was evidently broken during a flood in the 1950s. At the time being, the water flows through the lock during floods. However, these constructions do not operate any more, because the space between the buuns is filled with sediments and the water flows straight past the constructions and carries sediments into the Narva River channel.

In front of buun 1, there is a sandy beach. The total amount of sediments exceeds 600,000 m^3, the area above the mean lake level (30 m) is more than 12 ha. Between buuns 2 and 3 there are some 300,000 m^3 of sediments. Thus, in the above-mentioned area alone, the total amount of sediments is estimated at ca 1 million m^3. The area of "dry land" rising above the mean water level is *ca* 20 ha. However, the land is excessively damp and unsuitable for use (except the beach in front of buun 1). It would be purposeful to determine the total amount and the character of the accumulated sediments once again, and to decide where these sediments can be used. To return the constructions into use, the space between the buuns should be cleared from the sediments and dam 4 must be provided with a lock. Currently, the area between the buuns is subject to paludification. To a certain extent, waterfowl inhabit it. Therefore, ornithologists have to assess the significance of the area from this aspect.

The regulation of water level in L. Peipsi is not a topical problem any more. But, keeping in view the further use of the river and the Narva Reservoir, the problems relating to the outflow of the Narva River may become acute once again.

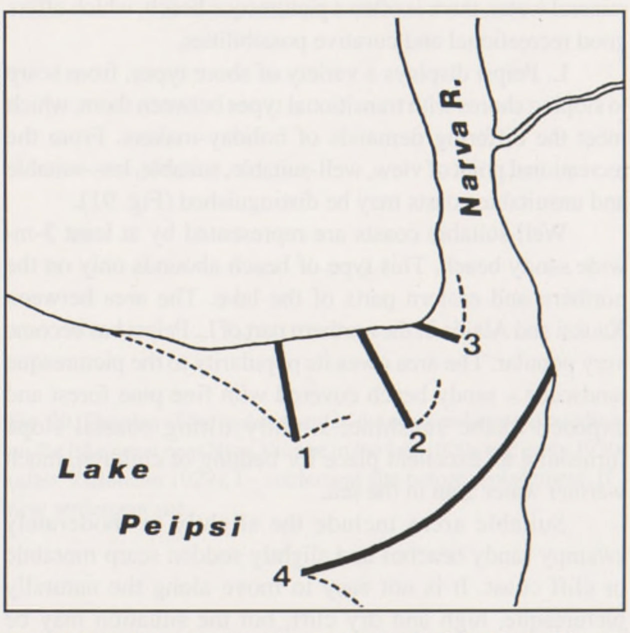


Fig. 92. Regulation constructions at the head of the Narva River.

And for the reason that the solid matter, which evidently originates from the sands on the northern coast of the lake, is carried straight into the river, where it accumulates in the river channel and in the upper portion of the Narva Reservoir.

According to earlier estimates, the protective constructions improved the conditions of outflow from L. Peipsi, but the hydrological effect of these constructions has not yet been analyzed. In all likelihood, this effect is overshadowed by the much more extensive natural fluctuations.

Another problem is the overgrowing of the head of the Narva River. Currently, there are no indications of noteworthy sedimentation in the lake, immediately in front of the riverhead. In all likelihood, sediments do not accumulate there, because the flow rate is much higher than prior to the construction of regulating devices.

5. MAIN FEATURES OF THE MORPHOLOGY AND FORMATION OF THE RIVER VALLEYS

The northern and southern parts of the Peipsi Basin differ in terms of the geology, topography and tectonic movements. These differences are reflected in the morphology of the river valleys, *e.g.* the changing shape of transverse profiles and the depth of the valleys, the existence and geology of the river terraces, the shape of longitudinal profiles. All these features suggest that the formation and evolution of the valleys took place under different conditions.

Longitudinal profiles. According to the shape, the longitudinal profiles of the rivers flowing in the Peipsi Basin could be divided into straight and concave. The only exception is the Narva River which falls into the Gulf of Finland. Like the other North-Estonian rivers, it has a convex longitudinal profile. A more or less straight longitudinal profile is characteristic to the rivers flowing into the northern part of L. Peipsi (Rannapungerja, Avijõgi, Utroya, Gdovka, Omedu). The stream gradient of these rivers ranges from 20 to 70 cm km^{-1}, only occasionally exceeding 100 cm km^{-1} (Miidel 1966a). The low gradient and straight longitudinal profiles are due to the bedrock topography which is lowering monotonously from the Pandivere Upland towards the lake. In some parts of the longitudinal profiles the gradient increases up to 200 cm km^{-1}, mainly due to the changes in the lithology of Quaternary deposits. In some cases, anomalously high stream gradient may also be caused by crustal movements. For instance, the higher gradient of the Avijõgi and Rannapungerja rivers has been induced by the Viivikonna zone of tectonic disturbances (Miidel 1966a).

The longitudinal profiles of the rivers rising on the heights and flowing into the southern part of L. Peipsi (Piusa, Võhandu, Ahja, Chernaya, Tolba, Lochkina, Optjok) are typically concave. As an exception serves the Velikaya – the longest river in the Peipsi Basin – which has a straight longitudinal profile reflecting a northward tilting of the bedrock topography in the Pskov - Velikoretski Lowland (Isachenkov 1969). Some 30 km before the mouth of the river, the stream gradient starts to decrease giving a concave shape to this section of the longitudinal profile. In the lower courses, the stream gradient is only 2-17 cm km^{-1} (Miidel 1996a).

The decrease of the stream gradient downward, particularly in the lower reaches of the rivers flowing into the southern part of L.Peipsi, has been explained by uneven tectonic uplift leading to the opening of the whole depression southwards and following the slow rise of the base-level.

Morphology of the valleys. In the northern part of the Peipsi Basin, the river valleys (Rannapungerja, Avijõgi, Tagajõgi) have mainly been eroded into the glaciolacustrine fine-grained sediments, less often into the low-lying till or Middle-Devonian rocks (Cherma, Gdovka, the upper reaches of the Narva River). In the plain topography, the fall of the rivers is small. As a result, the valleys are relatively small, being seldom deeper than 10 m (Rannapungerja) but, nevertheless, they are clearly expressed in the topography. The valleys of this kind are mostly 100 to 500 m wide, except the lower reaches of the Rannapungerja River and the upper section of the Narva River, the latter having a width of 2-3 km near the outflow from the lake.

The above-described flood-plain valleys have a symmetrical shape of the cross-profile; terraces occur only in the Rannapungerja Valley. In the upper part, where the rivers have cut into the till, V-shape cross-profile is common. The clearly expressed morphology of the valleys with a steep angle of slopes is due to the fine-grained sediments or till into which the rivers under discussion have cut. At the same time, the rivers flowing on the bedrock, e.g. the Narva River, have not been able to erode remarkable valleys and, therefore, they flow in the channels full of rapids.

The width of the floodplain is usually up to 500 m, except the lower reaches of the Rannapungerja River (Photo 33) and the Narva River near its outflow from the lake. The floodplain has a height of 1–2 m above a normal water level. The flat floodplain surface is cut by dry or partly filled oxbows, less often by overgrown oxbows. On the floodplain of the Narva River, a number of oxbows are still connected with the present channel. The meandering Narva River has formed remarkable point bars with wet areas filled by fine-grained alluvial deposits between them.

The floodplain is usually built up of fine-grained alluvial deposits, the total thickness of which in the Rannapungerja Valley is 4–5 m. The floodplain facies of fine-grained sand or silt with organic interlayers dominates. The somewhat coarser channel facies also consists of fine sediments; this is quite natural if to consider that the varved clay is its source material. In the Narva Valley, the alluvial deposits are 6–6.5 m thick. On the one hand, thus great thickness is related to the hydrological regime controlled by the crustal uplift and outcropping sedimentary rocks in the channel. On the other hand, the rapidly changing water level in L. Peipsi has played a significant role. These factors induce intensive meandering and sedimentation in front of rapids as local base levels. The same conclusion was reached by the Russian investigator Mozhaev (1973, 1973a), who mapped a local dome-like tectonic anticline between Permisküla and Omut on the right bank of the Narva River. If this anticline is still active, as Mozhaev supposed, the wide floodplain, great thickness of alluvial deposits and intensive lateral migration of the channel in the upper reaches of the Narva River, could be explained with the crustal uplift.

Photo 33. The low and wide floodplain near the mouth of the Rannapungerja River. Photo by A. Miidel.

The rivers in the southern part of the Peipsi Basin often flow above the buried pre-Quaternary valleys (Tavast & Raukas 1982). According to Miidel (Miidel 1966b, 1982), the direction of the valleys is connected with the orientation of tectonic faults. For instance, on several occasions the Võhandu Valley changes rapidly its course, and the direction of its different sections coincides with the orientation of tectonic joints. The situation is much the same in the surroundings of Izborsk and Pechory where the valleys form an orthogonal network (Bekker 1924, Isachenkov 1969a). The valleys are prevailingly of NW-SE and NE-SW orientation. The different oriention (E-W) of the Piusa Valley section on the northern slope of the Lokno Anticline is probably caused by a tectonic disturbance.

In the southern part of the Peipsi Basin, the rivers have cut deeper valleys there where they descend from the heights and plateaux into the depression. Thus, the Ahja, Võhandu and Tolba valleys are 20–30 m deep, while the Piusa Valley is 40–45 m, and in the surroundings of Pechory even 50–60 m deep (Photo 34). These valleys are eroded mainly into Middle Devonian terrigenous rocks. The southernmost rivers falling into L. Peipsi (Optjok, Velikaya) have their valleys in Upper Devonian carbonate rocks. In the lower reaches, the valleys turn shallower but, still, the Võhandu Valley is *ca* 10 m and the Velikaya Valley *ca* 10–15 m deep (Isachenkov 1969). Valleys eroded into the Quaternary deposits or bedrock end usually at some distance from the lake, at the boundary of the lake basin. Less often, for example, in the area where the Devonian Kuesta joins the lake from the south, deep valleys (Optjok, Kamenka, Velikaya) reach the lake. In their lower reaches, rivers usually flow over boggy lowland areas forming only channels. Near the river mouth, the bottom of the valley or channel is often below the present lake level: in the Ahja Valley near Lääniste 7–8 m (Fig. 93), in the Võhandu Valley near Räpina 10–11 m (Fig. 94) and near Võõpsu at least 7–8 m, in the Optjok Valley about 10 m (Miidel 1966a, Miidel & Tavast 1981, Miidel *et al*. 1995). A number of river mouths (Värska, Kul'e) are flooded and turned into estuaries (Photo 35). The only delta is situated in the mouth of the Velikaya River. The small boggy islets, forming this delta, have their surfaces below the present lake level (Isachenkov 1969).

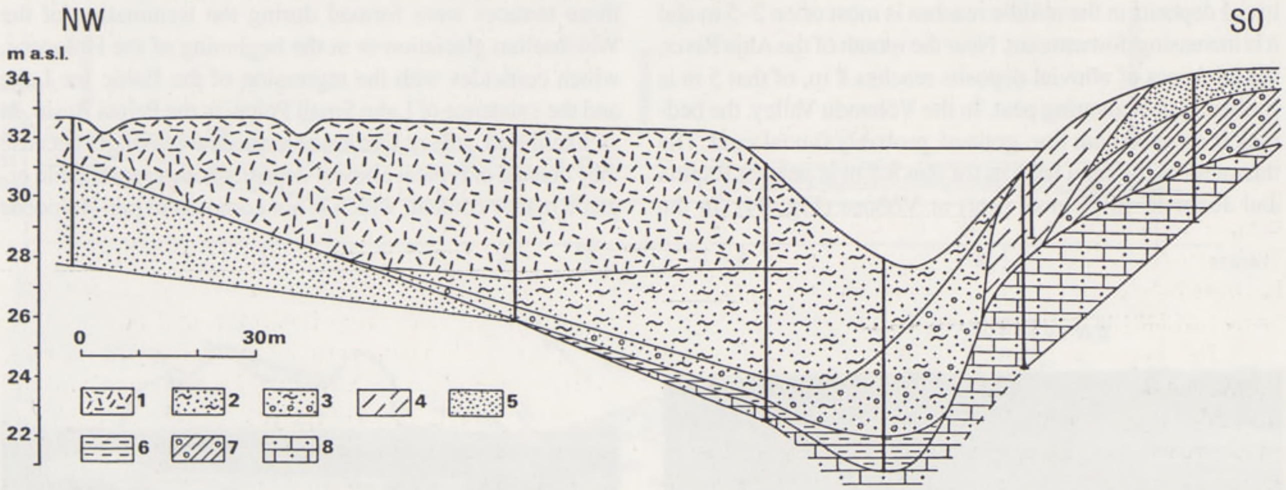


Fig. 93. Cross-section of the Ahja River, near Lääniste (after Pärna, with complements of Miidel & Tavast 1981): 1 - peat; 2 - sandy and 3 - gravelly river deposits with plant remnants; 4 - slope wash; 5 - glaciolacustrine sand; 6 - glaciolacustrine silt; 7 - till; 8 - Devonian sandstone.

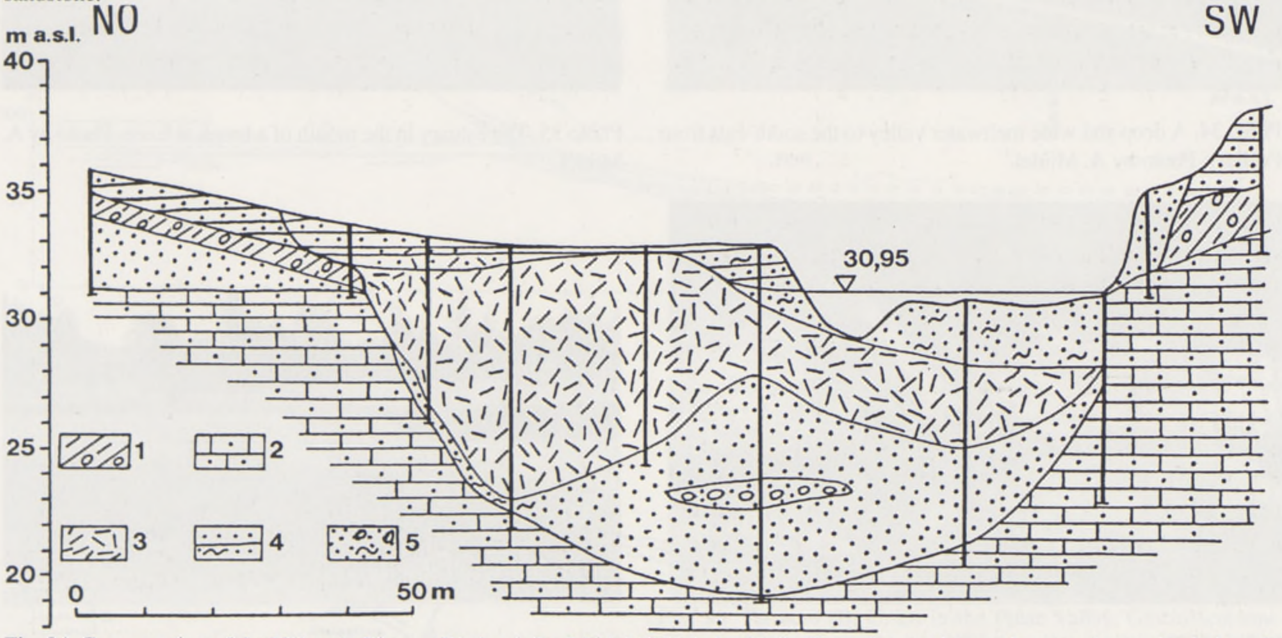


Fig. 94. Cross-section of the Võhandu River at Räpina (after Miidel 1966a): 1 - till; 2 - Devonian sandstone; 3 - peat; 4 - overbank deposits with plant remnants; 5 - channel deposits with gravel or plant remnants.

Formation of deep valleys was determined by the topography in the southern part of the Peipsi Basin. Total fall of the rivers rising on the surrounding heights is remarkable, and may exceed 200 m (Piusa 215 m, Võhandu 78 m). Downcutting of rivers was also stimulated by outcropping of terrigeneous Devonian sandstone and tectonic uplift of the area during the Late Weichselian.

The valleys under discussion are generally 100–500 m, in the lower reaches up to 2 km wide (Piusa Valley). In the middle reaches, they are usually 100–150 m wide with a commonly asymmetric transverse profiles, especially in the parts eroded into the bedrock (Võhandu, Ahja, Piusa). Frequently, the undercut slope is almost vertical (Photo 36), while the opposite slope is gently descending and often with terraces. The valley width increases downstream, the banks turn more gentle and asymmetric transverse profiles more symmetric. The 1–2-m-high floodplain has a various width and numerous overgrown oxbows. Somewhat different is the floodplain in the lower reaches of the Velikaya River, where the height of the rock-cut floodplain is 3–3.5 m (Isachenkov 1969). Accumulative alluvial floodplains prevail. The thickness of alluvial deposits in the middle reaches is most often 2–5 m and it is increasing downstream. Near the mouth of the Ahja River, the thickness of alluvial deposits reaches 8 m, of that 5 m is formed by the covering peat. In the Võhandu Valley, the bedrock is covered with fine-grained, probably fluvial sand. The thickness of the sand is 13 m (of that 8.5 m is peat) at Räpina and about 15 m (5 m of peat) at Võõpsu (Fig. 94). In the mouth of the Optjok River, peat forms a 10-m-thick layer; however, the genesis (fluvial? lacustrine?) of the floodplain (fine-grained sand) is not clear.

Terraces. The existence and distribution of river terraces is one of the most noteworthy difference in the morphology of valleys between the northern and southern parts of the Peipsi Basin. As mentioned above, in the northern part of the basin terraces are found only in the Rannapungerja Valley (Photo 37), while in the southern part they are present in all bigger valleys.

Tuhkanen (1968) distinguished and mapped terraces at seven different levels in the Rannapungerja Valley (Fig. 95). The altitude of these terraces in their downstream edge is 32.4, 33.0, 34.0–34.6, 36.2 and 37.0 m. Two higher terraces are expressed only in the upper reaches of the valley at an altitude of 40.5 m. Relative height of the terraces is small ranging from 0.5 to 1.8 m. Higher terraces are narrower (width 15–18 m) than the lower ones (60–300 m). The terraces have preserved as relatively short segments with their surface tilting towards the channel. According to Tuhkanen (1968), all these terraces were formed during the termination of the Weichselian glaciation or at the beginning of the Holocene, which coincides with the regression of the Baltic Ice Lake and the existence of Lake Small Peipsi in the Peipsi Basin. In all likelihood, at least five lower terraces are younger because they consist of typical alluvial deposits with a remarkable organic matter content, which is not common to pre-Holocene

Photo 34. A deep and wide meltwater valley to the south-east from Pechory. Photo by A. Miidel.

Photo 35. The estuary in the mouth of a brook at Lis'e. Photo by A. Miidel.

Photo 36. The steep undercut bank in the Devonian sandstones at the Ahja River. Photo by A. Miidel.

Photo 37. Terraces with dry channels in the Rannapungerja Valley. Photo by A. Miidel.

fluvial deposits in Estonia. V and IV terrace (36.2 and 37 m a.s.l.) have been linked to the Raadna ancient coastal formations which were formed in the second half of the Atlantic Chronozone (Miidel & Tavast 1981).

In the southern part of the Peipsi Basin, river terraces are widespread in the Ahja (Muru 1970), Võhandu (Sokman 1971), Piusa (Hang *et al.* 1964, Liblik 1966) and Velikaya (Isachenkov 1969) valleys. Probably, there are terraces in the Optjok, Zhelcha, Chernaya and Tolba valleys as well but, unfortunately, they have not been studied.

Liblik (1966) distinguished more than ten terraces in the Piusa and Võru-Hargla valleys. These terraces are concentrated into four main groups lying on different levels (Fig.96). The highest group A is situated in the Võru-Hargla Valley, but is palaeogeographically (genetically) linked with the terraces in the Piusa Valley. The distinguished terrace levels in this group are 95, 84, 82, 78, 75, 73 and 71.5 m a.s.l. As the horizontal terraces consist of glaciolacustrine and glaciofluvial sand, Liblik (1966) identified them as glacial outwash plains. The altitude of lower terraces in the Piusa Valley is 62 and 60 m (B); 51, 48 and 45 m (C); 41, 37, 35 and 33 m (D) (Fig. 96). The terraces of group B and C are well developed. Their width is increasing downstream up to 500 m and their relative height is 2–3 m. Their inside structure is not well known. The highest terrace in group B is an erosional rock-cut terrace in the upstream part which consists of fine-grained sediments in the lower reaches of the valley. The terraces of group C and D consist of fine-grained sand, the grain-size is decreasing downstream (Liblik 1966). Of the terraces of group D, only short and narrow segments have preserved. The terraces of both C and D group are slowly tilting towards the channel.

In the Ahja Valley, 16 terraces are concentrated into five groups (Fig.97) (Muru 1970). The altitude of terraces in the lower reaches is 85, 84, 79 and 76 m (group A); 62.5 and 61.5 m (B); 54, 52 and 50 m (C); 44 and 41 m (D). The width of the terraces varies from a few metres to several hundred metres, while the width of the second terrace in its lower reach is 800-900 m. In places, terrace surfaces are tilting towards the river. Usually, terraces have preserved only on one bank, less frequently they are placed symmetrically on both sides. Unlike the Piusa Valley, the terraces in the Ahja Valley are

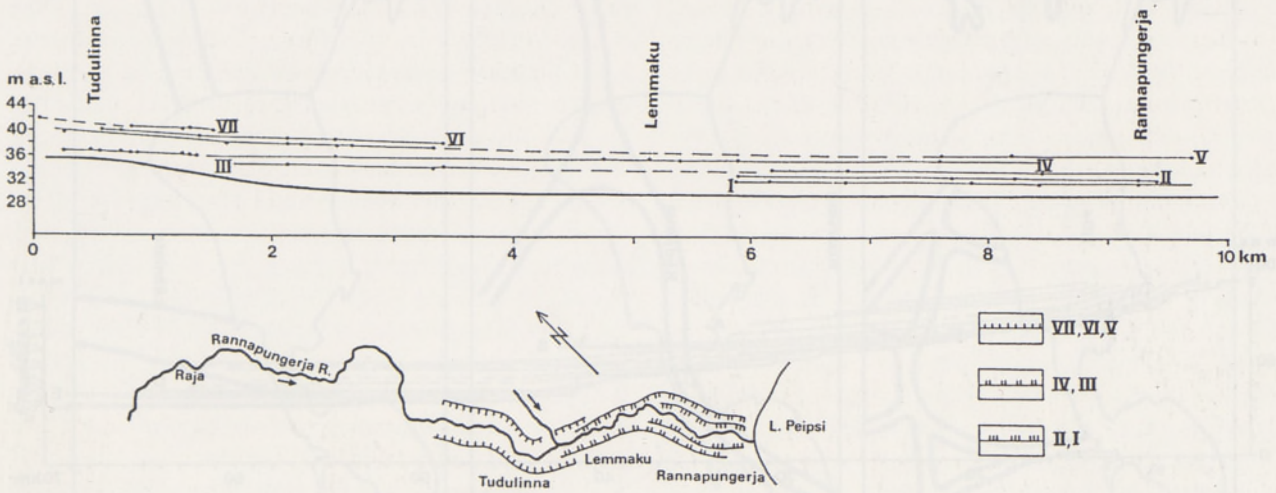


Fig. 95. Terraces (I - III) in the Rannapungerja Valley (after Tuhkanen 1968).

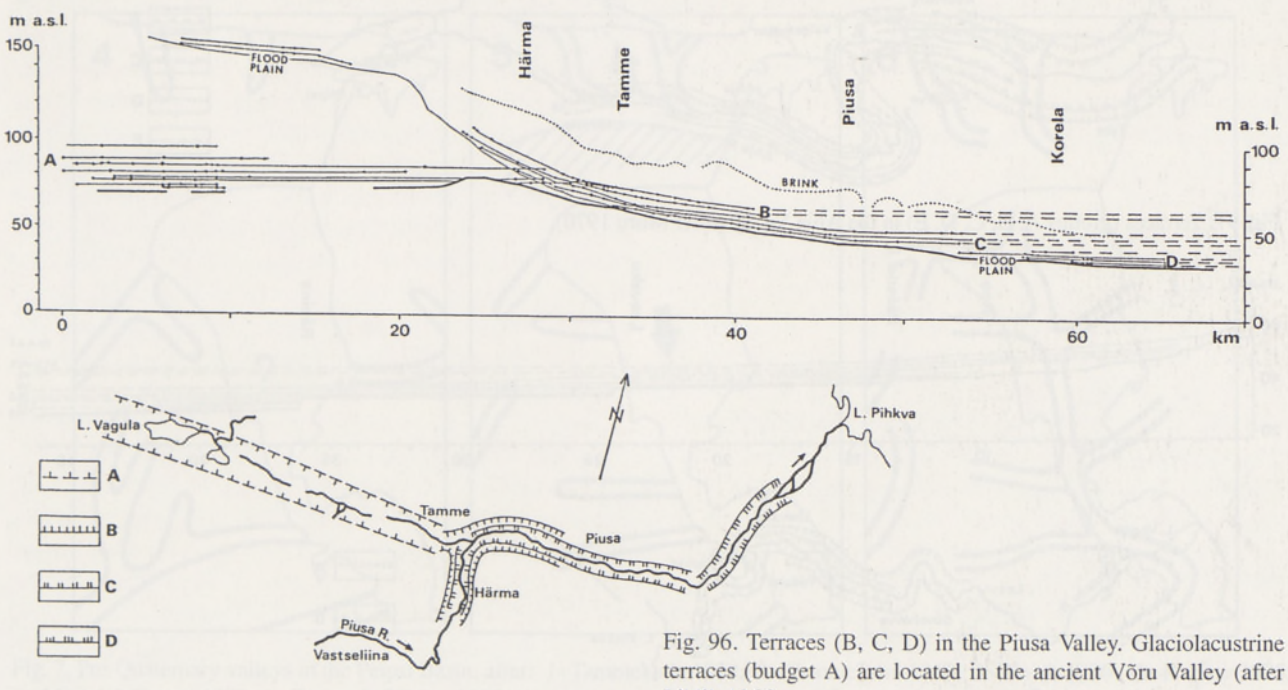


Fig. 96. Terraces (B, C, D) in the Piusa Valley. Glaciolacustrine terraces (budget A) are located in the ancient Võru Valley (after Liblik 1966).

mainly rock-cut. The sockle of terrigeneous Devonian rocks is covered by a 1–2-m-thick sandy and gravel layer, the latter being thicker in younger low-lying terraces (Muru 1970).

According to Muuga (1966) and Sokman (1971), six terraces in the Võhandu Valley are divided into two groups (Fig. 98). Group A consists of two terraces at an altitude of 47.5–46.5 and 45 m, and group B of four terraces with an height of 39, 36.5, 35 and 32.2 m a.s.l.. The width of the terraces ranges from 20 to 200 m. Two higher terraces are rock-cut (Photo 38), while the lower terraces consist mainly of sand.

Terraces have been studied also in the Velikaya Valley where, according to Isachenkov (1969), they have preserved at six levels. The higher-lying five terraces are probably not connected with the evolution of L. Peipsi. Judging by the altitude, they were formed during the time when a proglacial lake existed south of Pskov, before the Pihkva proglacial lake came into being. Downstream from Podmogil'e, only one terrace and a floodplain have preserved. The height of that rock-cut terrace increases from 5–6 m to 11–12 m in the surroundings of Pskov. According to Isachenkov (1969), the terrace joins there with a glaciofluvial delta formed during the existence of a proglacial lake in which the water level was about 45 m a.s.l. Downstream from Pskov, Isachenkov distinguished a 3–3.5-m-high (relative height from the main river-level) rock-cut floodplain and a 5–6-m-high rock-cut terrace.

Summarizing the geology, morphology and distribution of river terraces, the following could be concluded:
(1) terrace spectra are opening downstream (differences in altitude between terrace groups are increasing) ;
(2) higher terraces have preserved in the upstream part of valleys only;
(3) relative height of terraces is increasing downstream;
(4) terraces are erosional; thin alluvial cover is often difficult to separate from the low-lying sediments because of similar grain-size;
(5) it is not possible to distinguish floodplain and channel facies in the alluvial cover of terraces;
(6) sediments, building up terraces, do not contain organic matter;
(7) there are no signs of ancient channels on the terrace surfaces.

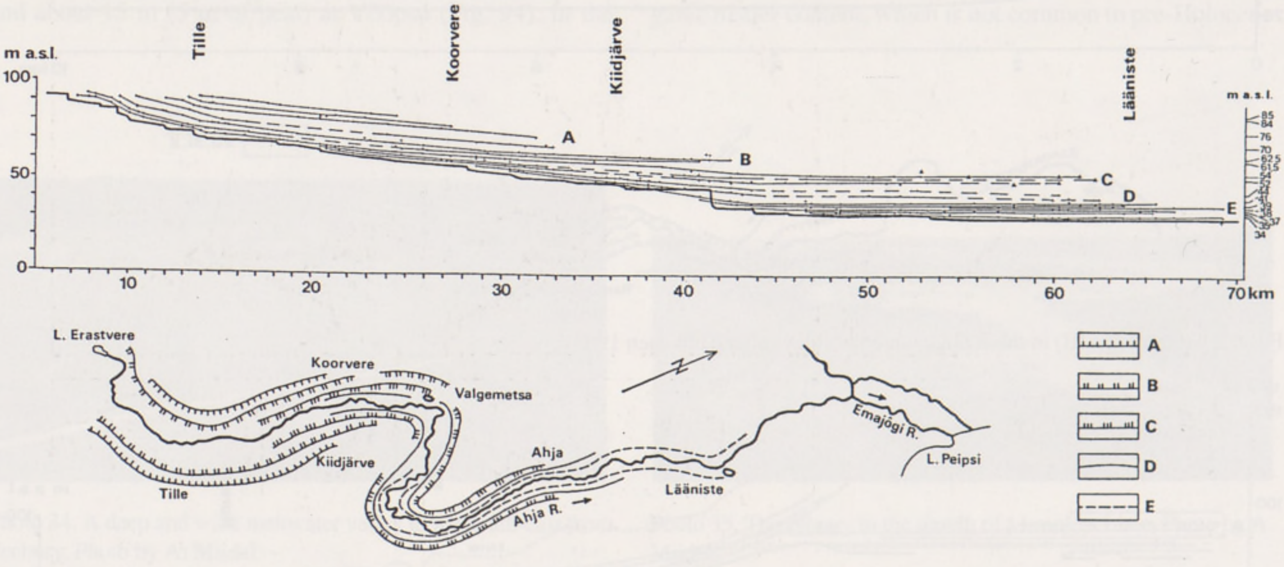


Fig. 97. Terraces (groups A, B, C, D, E) in the Ahja Valley (after Muru 1970).

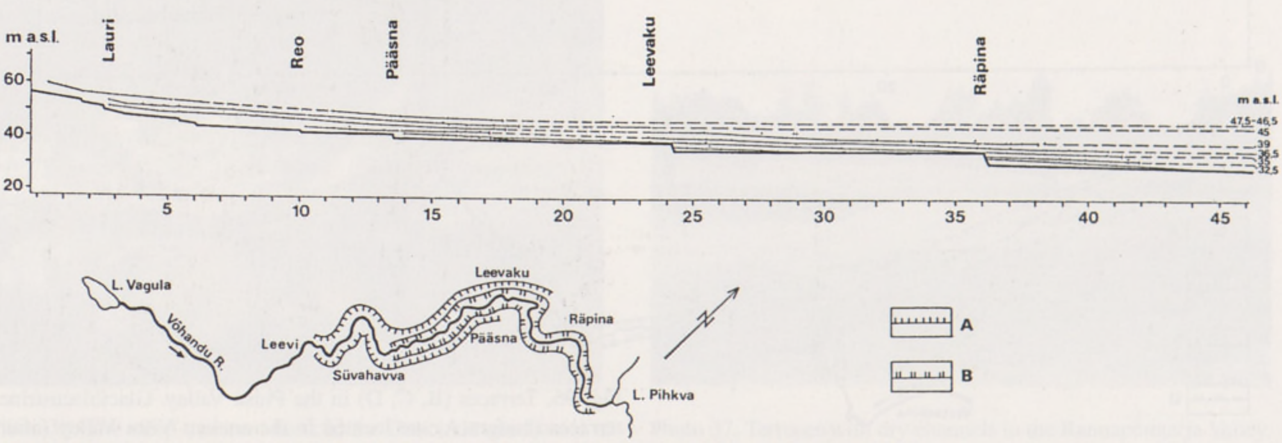


Fig. 98. Terraces (groups A and B) in the Võhandu Valley (after Muuga 1966, and Sokman 1971)

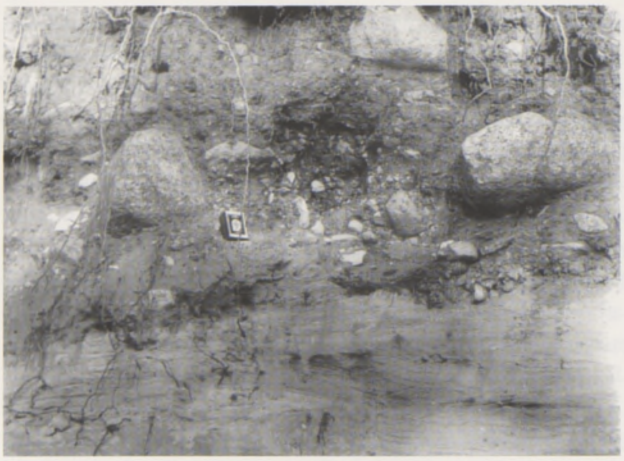


Photo 38. Devonian sandstones in a rock-cut terrace of the Võhandu Valley are covered by very coarse channel deposits. Photo by A. Miidel.

Formation of valleys. In the formation of the river valleys in the Peipsi Basin four stage have been identified (Miidel & Tavast 1981): Pre-Pleistocene, Pleistocene, Late-glacial and Holocene.

Pre-Pleistocene Stage

In defining this stage it has been assumed that deep depressions in the bedrock topography, mapped by drillings (Ch. 2), are parts of ancient valleys. Connecting these deeps by extrapolation, the network of probable pre-Quaternary valleys was reconstructed. The age of the valleys is difficult to determine, because only Middle-Pleistocene till as oldest is known from these buried valleys.

As early as 1928, Tammekann postulated that two wide valleys, one oriented to the north and the other to the west, start from the Peipsi Basin (Fig. 99-1). According to him, meltwater discharge from the Central Russian Upland went through these valleys. Since then, a number of investigators

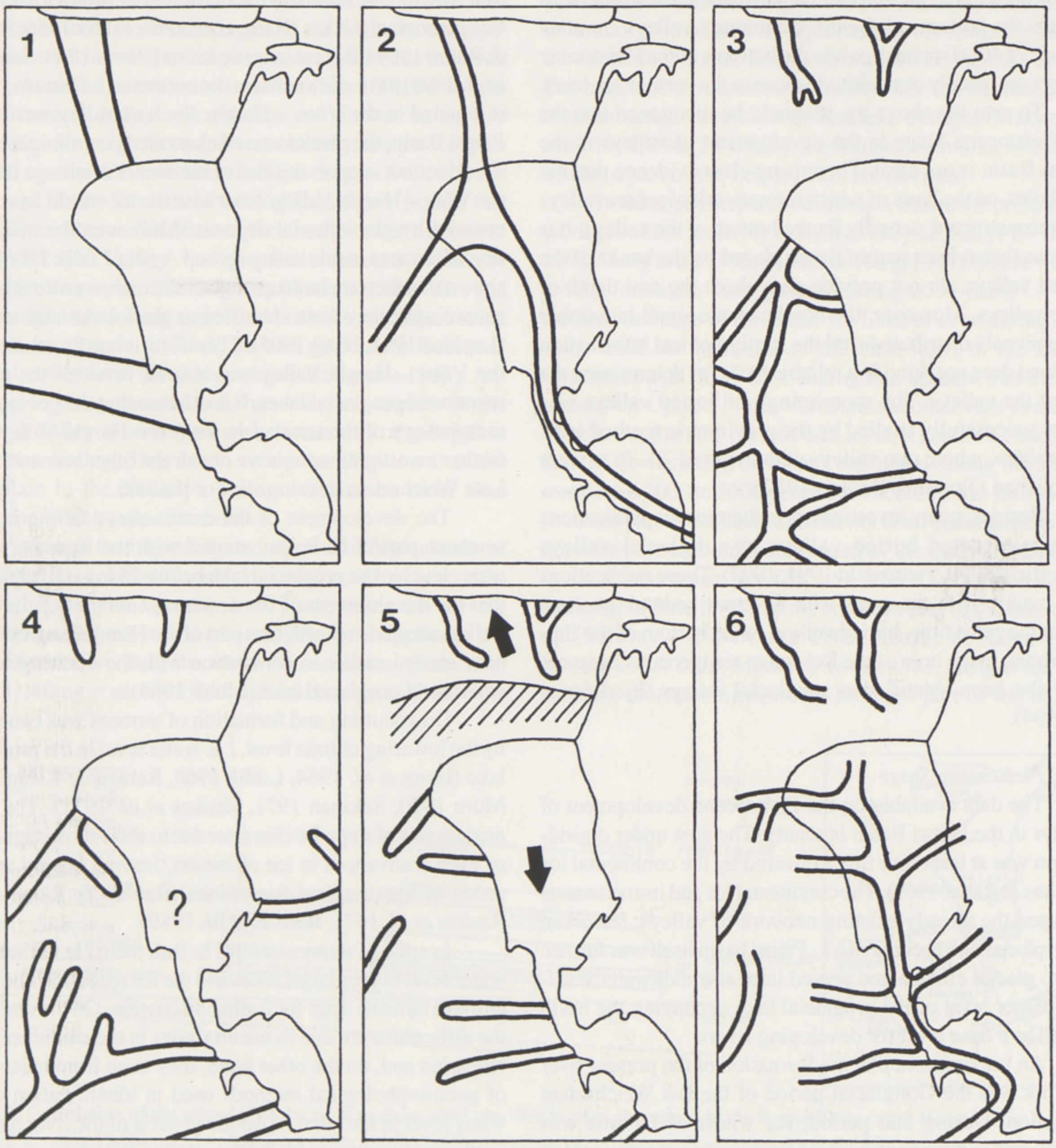


Fig. 7. Pre-Quaternary valleys in the Peipsi Basin after: 1- Tammekann 1928; 2 - Tammekann 1949; 3 - Kajak 1970; 4 - Kvasov 1975; 5 - Miidel & Tavast 1978; 6 - Tavast & Raukas 1982.

(Tammekann 1949, Verte 1962, Kajak 1970, Kvasov 1975, Miidel & Tavast 1978, Rähni & Tavast 1981, Tavast & Raukas 1982) have expressed different opinions as to the network of ancient valleys and the conditions of meltwater discharge in the area. The main point of the discussion has been the orientation of valleys – either along (Tammekann 1928, 1949, Miidel & Tavast 1978) or across (Kvasov 1975) the lake basin (Fig.99). Thereby, the glacial erosion must have been the main factor responsible for the formation of the basin (Tammekann 1928, Tavast & Raukas 1982). The latter opinion has been supported by the results of seismoacoustic investigations in the northern L. Peipsi (Noormets *et al.* 1998, Miidel *et al.* 1999). Tavast and Raukas (1982) hold the opinion that the buried valleys under discussion are of pre-Pleistocene age which have lost a lot of their morphological features. However, in all likelihood, there occur also valleys which were formed during the Pleistocene. Kajak (1970) distinguishes between the valleys of different age, using the depth of downcutting as a criterion. He postulates that in southern Estonia the bottoms of the pre-Quaternary valleys are at an altitude of 40–60 m b.s.l., while the bottoms of the Pleistocene valleys are usually at an altitude close to the present sea level.

To sum the above up, it should be mentioned that the pre-Pleistocene Stage in the development of valleys in the Peipsi Basin is not clear. There is no clear evidence that the boreholes, on the basis of which the network of palaeovalleys was reconstructed, actually fix the bottom of the valleys. It is obvious that at least part of them, placed on the banks of the buried valleys, do not provide data about the real depth of these valleys. Moreover, the boreholes are small in number and unevenly distributed, and the stratigraphical information obtained does not provide a reliable basis for determining the age of the valleys. The morphology of buried valleys was rather successfully studied by the gravimetric method in L. Lämmijärv, where two valleys with depth of 70–76 m were established (Tavast & Raukas 1982).

Besides, many investigators in their recent publications have interpreted buried valleys as subglacial valleys (Wingfield 1990, Pietrowski 1994, 1997). These publications deal mainly with the areas with different bedrock geology but, at the same time, buried valleys in the bottom of the Baltic Proper, in the area of the Devonian terrigeneous bedrock, have also been identified as subglacial valleys (Bjerkéus *et al.* 1994).

Pleistocene Stage

The data available on the Pleistocene development of valleys in the Peipsi Basin is scanty. The area under consideration was at least four times covered by the continental ice (Raukas & Rähni 1969). The continental ice and its meltwaters reshaped the already existing network of valleys, reworked and replaced older sediments. L. Peipsi basin itself was formed due to glacial erosion and served later as a meltwater reservoir. Water level of the proglacial lake occupying the basin served as a base level for developing rivers.

An important step in the formation of the present river network was the Gotiglacial period of the last Weichselian glaciation. During that period, the whole of Estonia was deglaciated and the continuously enlargening proglacial lake occupied the Peipsi Basin (Ch. 6). The Late-glacial Stage in the development of valleys was distinguished by Miidel and Tavast (1981) and it covered the time span from the Otepää (12,600 ^{14}C yr BP, uncalibrated) Stage of the Weichselian glaciation to the beginning of the Holocene. Later, it was shown (Lepland 1991, Hang 1995) that the formation of valleys could have been started earlier by the erosion of subglacial meltwaters. The idea was based on the fact that in the bottom of the lower reaches of the Piusa Valley, coarse-grained glaciofluvial deposits are lying at a very low (about 20 m) altitude. As there are no signs of such a low base-level in the area, it should be concluded that the fracture system was formed in the glacier and used by meltwaters.

It has been assumed that the glacial meltwaters from the proglacial lake in the southern the Peipsi Basin discharged through the Võru – Hargla Valley westwards into the proglacial lake in the Gauja Basin (Hausen 1913, Ramsay 1929, Liblik 1966, Raukas & Rähni 1969). Westward discharge terminated when the water-level in Pihkva proglacial lake lowered to 75 m a.s.l. and the watershed was formed between Pihkva and Gauja proglacial lakes (Hang *et al.* 1964, Liblik 1966, Raukas & Rähni 1969). Recent investigations (Pirrus 1969, Sandgren *et al.* 1998) have shown that in the bottom of L. Tamula, which is situated in the Võru – Hargla Ancient Valley west of the Peipsi Basin, the glacial varved clays are lying straight on till. This does not support the idea of meltwater discharge through the Võru – Hargla Valley, otherwise the till would have been covered by glaciofluvial deposits. Moreover, the investigation of terraces in this valley (group A after Liblik 1966) have shown that they are built up of glaciolacustrine sediments and, accordingly, they were classified as glaciolacustrine terraces (Lepland 1991, Hang 1995). Therefore, it was concluded that the Võru – Hargla Valley was a strait between the above-mentioned proglacial lakes. It is obvious that the geology and morphology of the terraces in the Võru-Hargla Valley need further investigation to prove one or the other scenario of the Late Weichselian development of the area.

The development of the river valleys falling into the southern part of L. Peipsi, started with the lowering of the water level in the proglacial lake below 75 m a.s.l. It might be that the development of the uppermost section of the Piusa Valley, situated in the highest part of the Haanja Heights, could have started earlier in connection with the discharge of the small local proglacial lake (Liblik 1966).

Downcutting and formation of terraces was controlled by the lowering of base level, *i. e.* water level in the proglacial lake (Hang *et al.* 1964, Liblik 1966, Raukas & Rähni 1969, Muru 1970, Sokman 1971, Raukas *et al.* 1971). The latter process was of stepwise character due to short-term stagnations or even readvances in ice recession, leaving behind several zones of ice-marginal formations (Raukas & Rähni 1969, Raukas *et al.* 1971, Raukas *et al.* 1989).

In spite of many attempts to link the river terraces, the water level in proglacial lakes and the ice recession, the problem still remains open for further discussion. On the one hand, the difficulties are due to uncertainties in the character of ice recession and, on the other hand, they arise from inaccuracy of geomorphological methods used in identification of the water level in ice lakes. Also the genesis of the river terraces has been long discussed. It is suspicious to link every terrace or terrace group with certain water level standstill, while it is hard to believe that the changing base level, which caused erosion, could design a certain terrace along the whole valley, and that with the next base level lowering the whole process could have been repeated again. Relatively fast ice recession (Hang *et al.* 1995) and the resultant frequent base-level changes did not leave enough time for such quite a slow process. Therefore, the terraces in these valleys are not cyclic terraces. Probably, the influence of base level change was first felt in the lower reaches of rivers, while in the other parts of the valley the formation of terraces at different levels continued simultaneously. Accordingly, only the downstream part or rather the downstream end of terraces or a group of terraces represent approximately the altitude of the base level. Based on that, Hang (Hang 1993, Hang *et al.* 1995) presented a new scheme showing the relationship between the terraces, the water level in the proglacial lakes and the ice recession from the Peipsi Basin (Fig. 100). He used the altitude of ancient coastal formations distinguished on the western coast of L.Peipsi (Liblik 1969) and transformed them together with the terraces from the southern part of the basin, to the baseline oriented towards the fastest upheaval of the Earth's crust (326°). According to such a correlation (Fig. 100), the formation of terraces started when the ice margin was on the line Piirissaar - Knyazya Gora and terminated when the ice margin had reached the line Männikvälja - Iisaku. Still, it is not clear how do the terraces of the Velikaya River fit into that correlation scheme. The first terrace in the lower reaches of the Velikaya Valley has been correlated with the glaciolacustrine plain (38-40 m a.s.l.) south of the lake, and the floodplain has been correlated with the glaciolacustrine plain at an altitude of 34-35 m and dated as a Holocene (Boreal) relief form (Malakhovskij & Bakanova 1969). According to the presented scheme, the first terrace in the Velikaya Valley is correlated with the terraces of group D in the Piusa Valley. This leads to the assumption that the floodplain in the Velikaya Valley can also be older than Holocene, which is hard to believe.

Although the beginning and the end of deglaciation of the Peipsi Basin are still under discussion, it has been concluded that the whole process took place during a relatively short (about 500 yr) period of time. The same applies to the terrace formation. Both the varve chronological (Raukas *et al.* 1971, Hang *et al.* 1995) and radiocarbon dates (Raukas *et al.* 1971) suggest that the terraces were formed during the course of 200–300 years (Hang 1993, Hang *et al.* 1995). During that period, the water-level of the proglacial lake occupying the Peipsi Basin, lowered *ca* 30 m and, based on the terrace levels in the Piusa Valley, even 60 m (Hang *et al.* 1995). This shows that the correlation of terraces and the water level in the proglacial lake is problematic, because the downcutting and terrace formation were rapid and continuous processes without long-term standstills of the water level. Downcutting was probably faster in the areas with steep inclination of underlying topography, and the following lateral erosion destroyed partly the already existing terraces. The above-described characteristic features of terrace formation have been supported by numerous tests in laboratory and nature (Makkaveev *et al.* 1961, 1962, Bylinskij 1957). In L. Sevan the water level dropped about 10 m within 23 years. This gave rise to intensive downcutting and lateral erosion, especially in the areas with steep inclination (similar conditions exist in South-Estonian heights and plateaux) (Bylinskij 1957). Laboratory simulation of rapid base level lowering caused the formation of a terrace spectrum with an increase in relative height of terraces downstream, while the number of terraces exceeded the number of base level changes (Makkaveev *et al.* 1961, 1962). Thereby, the tendencies similar to those in southern part of the Peipsi Basin are obvious.

In the southern part of the Peipsi Basin, the present network of valleys probably existed already 12,500 yr BP (Hang 1993, Hang *et al.* 1995, Miidel & Hang 1997). There was no terrace formation during the following 2000 years, *i.e.* until the final drainage of the Baltic Ice Lake or even during 3000 years until the next rise of the base level in the Boreal Chronozone (9000 yr BP). Taking into account that the bottoms of the Optjok, Piusa, Võhandu and other valleys are 8 to 13 m lower than the present lake level, it could be concluded that the downcutting of valleys continued during the above-mentioned 2000 to 3000 years. Downcutting was faster in the southern part of the basin and has not been documented in the northern part. It could be explained by the faster crustal uplift in the southern part (see Ch. 2.5), which led to such a great lowering of base level that the southern part of the lake dried up (Ramsay 1929, Orviku 1960, Raukas & Rähni 1969, Kvasov 1975, 1979, Miidel & Tavast 1981, Hang *et al.* 1995 a.o.). At the same time, in the northern part of the basin a

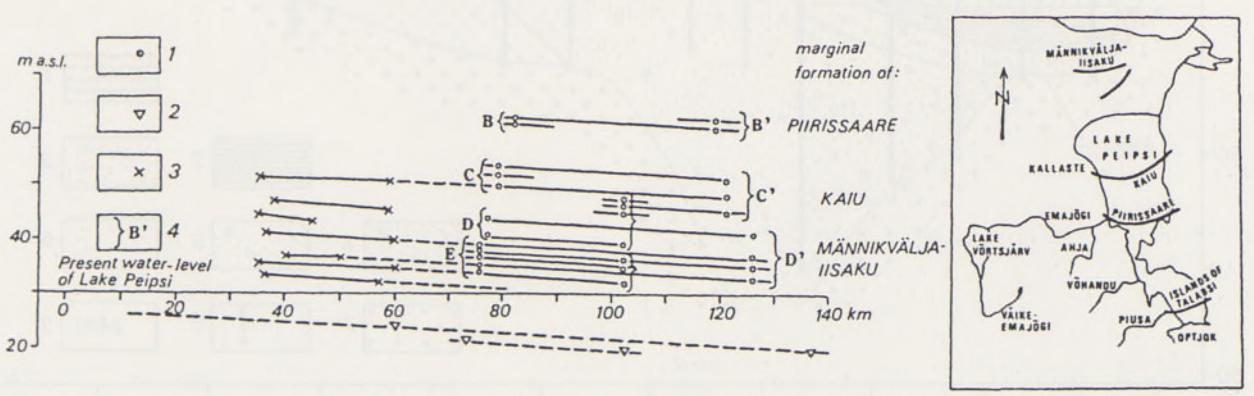


Fig. 100. Correlation of the altitudes of the river terraces and Late-Weichselian shorelines in the Peipsi Basin (after Hang 1995): 1 - river terraces; 2 - valley bottoms; 3 - shorelines (after Liblik 1969); 4 - group of terraces with similar ages and their correlation with the ice-marginal formations (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.

small body of water called Small Peipsi was formed. It has been postulated that during that period, the Velikaya with its tributaries (Piusa, Zhelcha, Võhandu a.o.) flowed in the southern part of the basin and fell into Small Peipsi north of Piirissaar Island (Hang *et al.* 1964, Raukas & Rähni 1969, Rähni 1973). Unfortunately, there are no straight signs of the valleys in the bottom of the lake. Seismoacoustic profiling, provided in the lake, didn't give indisputable results either (Noormets *et al.* 1998). The results of this study showed that there is a wide, up-to-1-m deep depression 2-3 km east of the mouth of the Emajõgi River, and a 2-m-deep and 1-2-km-wide depression to the north from Piirissaar Island. However, it is not known whether these depressions are parts of a buried valley or not. Recent biostratigraphical investigations in the southern part of the lake suggest that during the discussed time span there still existed remains of small lakes (Pirrus *et al.* 1985, Davydova & Kimmel 1991, Miidel *et al.* 1995); the Velikaya had to pass through several small lakes before its waters reached Small Peipsi.

Holocene Stage

During the Holocene, the lower reaches of the valleys in the Peipsi Basin were filled with alluvial and organic deposits due to continuous base-level rise. According to pollen data, fluvial and lacustrine sediments started to accumulate at different times in different parts of the basin: in the mouth of the Optjok River in the Early Boreal (Miidel *et al.* 1995), in the mouth of the Võhandu River in the Boreal (Pirrus & Tassa 1981) and in the mouths of the Emajõgi and Rovya rivers in the Late Atlantic (Thomson 1939, Sarv & Ilves 1975, Miidel *et al.* 1975). At the beginning of the Atlantic Chronozone, the accumulation of organic deposits started in the mouth of the Kuna River and at the beginning of the Sub-Boreal Chronozone in the Samolva River (Miidel *et al.* 1975). It could be assumed that the formation of the Velikaya delta started simultaneously with the accumulation of sediments in the Optjok Valley, but there are no direct evidence supporting this conclusion.

The accumulation process spread upstream in the valleys according to base level change. The floodplain of the Piusa River is covered with a swamp extending up to 5 km from the river mouth. Within the distance of 15–20 km upstream, the oxbows on the floodplain are filled with an up-to-10-m-thick layer of organic deposits. The accumulation of oxbow sediments started already in the Pre-Boreal Chronozone, which is demonstrated by the non-calibrated radiocarbon dates (9360±70 (TA-2230) ^{14}C yr BP) from the middle reaches of the Piusa River (Fig. 101) (Hang 1993). Dates from the same cross section, yielding a younger age, are connected with the younger generation of oxbows.

There is only scanty data available on the Holocene development of the valleys in the northern part of the Peipsi Basin. Terraces in the Rannapungerja Valley point to the base-level lowering, which, opposite to the conditions in the southern part of the basin, led to the formation of younger terraces and to the downcutting of rivers. As the higher terraces at an altitude of 37 and 36 m, possibly also at an altitude of 34-34.6 m, are correlated with the Raadna ancient coastal formations, formed in the Atlantic Chronozone, the formation of lower terraces and the downcutting of the valley took place when the coastline of Small Peipsi regressed from Raadna, *i.e.* at the end of the Atlantic Chronozone or later (Miidel & Tavast 1981). Geological data from the mouths of rivers do not indicate the overdeepening of valleys on the northern coast of L. Peipsi (Figs. 102, 103).

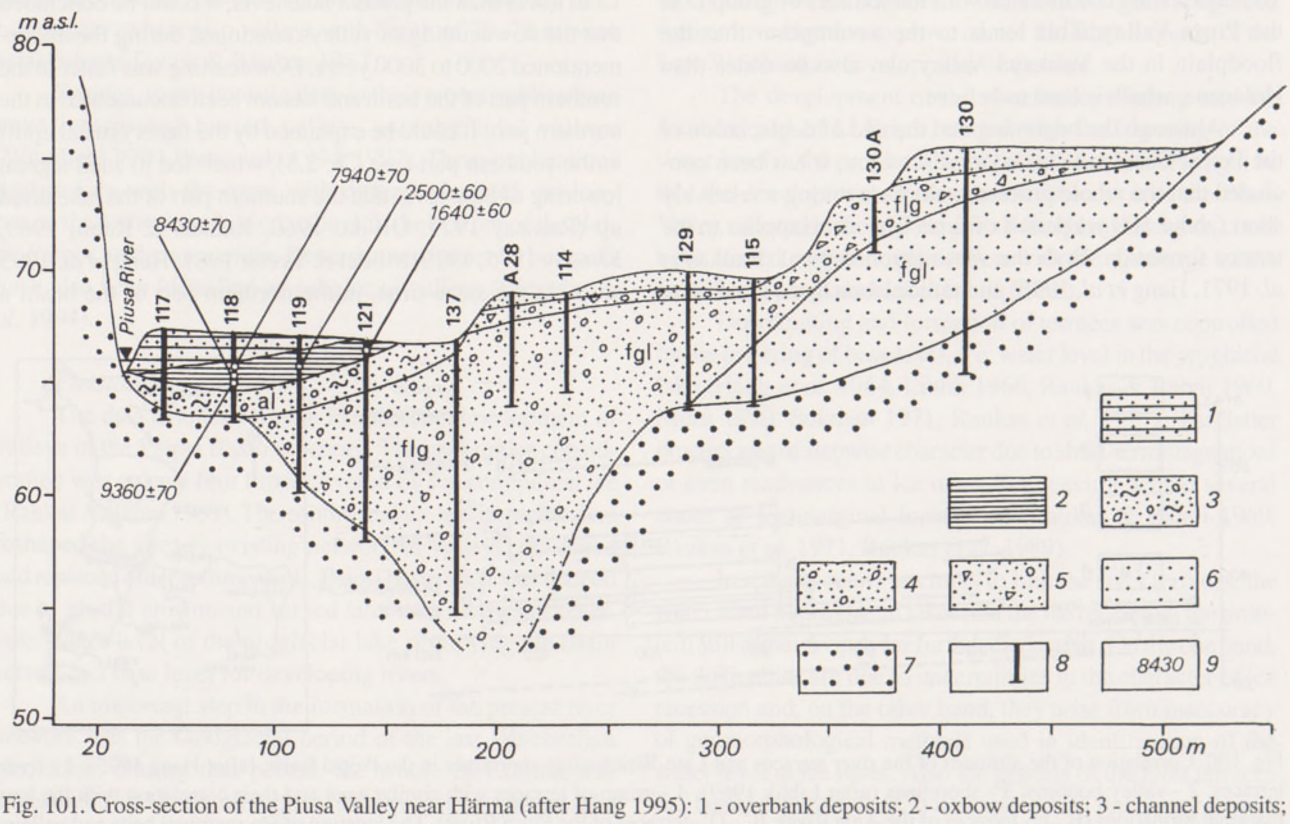


Fig. 101. Cross-section of the Piusa Valley near Härma (after Hang 1995): 1 - overbank deposits; 2 - oxbow deposits; 3 - channel deposits; 4 - glaciofluvial sand and gravel; 5 - glaciofluvial gravel with pebbles; 6 - glaciolacustrine sand; 7 - terrigenous Devonian rocks; 8 - borehole; 9 - ^{14}C datings (yr BP).

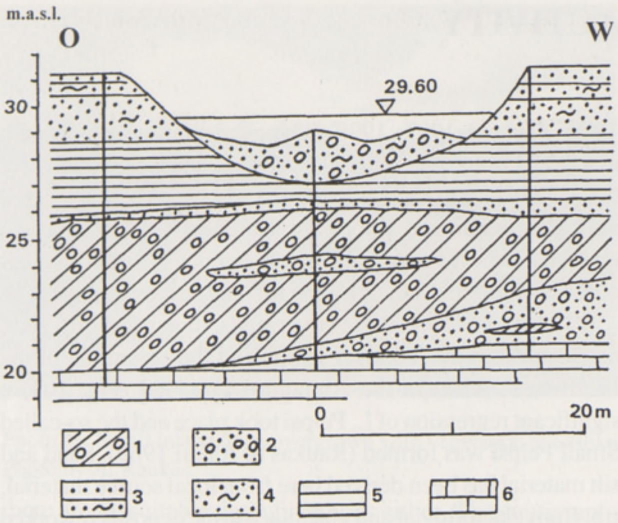


Fig. 102. Cross-section in the mouth of the Avijõgi River: 1 - till; 2 - channel deposits (gravelly sand with organic matter); 3 - overbank deposits (silt with organic matter); 4 - channel deposits (sand with organic matter); 5 - glaciolacustrine varved clay; 6 - Ordovician carbonate rocks.

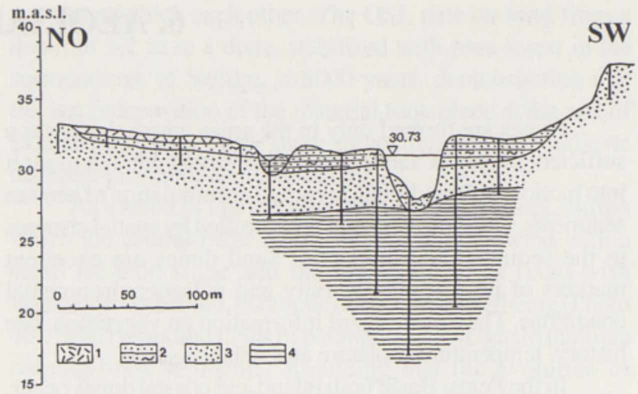


Fig. 103. Cross-section in the mouth of the Rannapungerja River (after Dzilna with complements of Miidel & Tavast 1981): 1 - peat; 2 - overbank deposits (silty sand with organic matter); 3 - channel deposits (sand); 4 - varved clay.

Still it remains unsolved when the outflow from L.Peipsi through the Narva Valley started. According to Liblik (1966), the formation of Small Peipsi was caused by the water level lowering in the proglacial lake at the site of the present-day Gulf of Finland which caused the outflow from the lake. It is also possible that the outflow was formed after the final drainage of the Baltic Ice Lake. At the same time, some researchers maintain that the outflow was opened much later. The palynologically studied oxbow sediments close to the outflow from the lake, have demonstrated that the accumulation of oxbow sediments started in the second half of the Atlantic . The beginning of the outflow has been correlated with that event (Rähni 1973, Miidel *et al.* 1975, Miidel 1976). This coincides with the regression of the lake from the Raadna coastal formations and with the terrace formation in the Rannapungerja Valley. But this assumption needs further verification as well.

6. AEOLIAN ACTIVITY

Dunes are formed only in the areas where there was a sufficient supply of sand, winds were strong enough to set it into motion and conditions favoured accumulation of aeolian sediments. Dune morphology is controlled by spatial changes in the sediment transport rate. Sand dunes are excellent markers of past aeolian activity and palaeoenvironmental conditions. They may impart information on vegetation, fire history, temperature-moisture and wind regimes.

In the Peipsi Basin both inland and coastal dunes occur. Due to the deficiency of sand and high precipitation rate, the dunes here are relatively low (mostly 5–15 m) and rendered stationary by vegetation. At present, the movement of dune sands is very limited (Photo 39). However, there were times, when the moving aeolian sands posed a great threat to the environment, devastating fields and buildings.

There are three primary controls of wind erosion in the basin: ground frost in wintertime, gravel lag in outcrops of till and glaciofluvial deposits, and vegetation. Even light covers of rooted vegetation, such as moss and low sparse grass can reduce erosion to negligible levels. Therefore, dune stabilization with vegetation plantings has been successful in Estonia, including the northern coast of L. Peipsi (Tiismann 1924).

The continental dunes in the Iisaku-Illuka region in an area of *ca* 50 km^2 are parabolic (Photo 40) or transversal, and indicate a westerly-northwesterly palaeowind direction (Rähni 1959, Zeeberg 1993, 1998). Dune orientations agree with modern NW-to-NE resultant drift directions. The dunes are 0.8–2.7 km long and up to 15–20 m high. Their west and northwest windward slopes are slanting (3–18°), while the opposite leeward sides are much steeper (18–24°). Small coversand hillocks occur on the top and slopes of dunes.

Photo 39. Aeolian sands moving towards the forest at Remniku. Photo by A. Raukas.

Photo 40. Parabolic inland dunes near Iisaku-Illuka. Photo by A. Miidel.

Most probably, these coversands, dunes and drift sand originate from the Younger Dryas and the beginning of the Pre-Boreal, when in the Alutaguse Lowland (Fig. 104) a significant regression of L. Peipsi took place and the so-called Small Peipsi was formed (Raukas & Rähni 1969). Sand and silt material has been derived here from local source material, *e.g.* from glaciofluvial and glaciolacustrine deposits reworked by ancient Lake Peipsi. The formation of dunes in this region started immediately after source deposits became available for aeolian redistribution, *i.e.* after the groundwater level had lowered and soils had dried. The process came to an end when the dunes became overgrown with vegetation (Photo 41). However, the dune crests have been cyclically active as evidenced by the buried soil horizons that crop out in the active crestal scours.

Owing to the short-lived character of aeolian activity in this area, the transition from glaciolacustrine to aeolian silty sands or sandy silts has not been distinguished in the sedimentological facies or by grain-size analysis. Probably, active dune forming here came to an end already in the second half of the Pre-Boreal when in summer the temperature was higher and the climate was more humid. This created favourable conditions for a rapid invasion of shrubs and trees, preventing extensive aeolian processes (Zeeberg 1998). Currently, the areas of inland dunes within the Peipsi Basin are surrounded by peat bogs and covered by pine forests. In the localities within the pine forest, the number of buried soil horizons is greater than elsewhere.

The dating of aeolian deposits is extremely difficult, because the sediments have been recurrently redeposited and

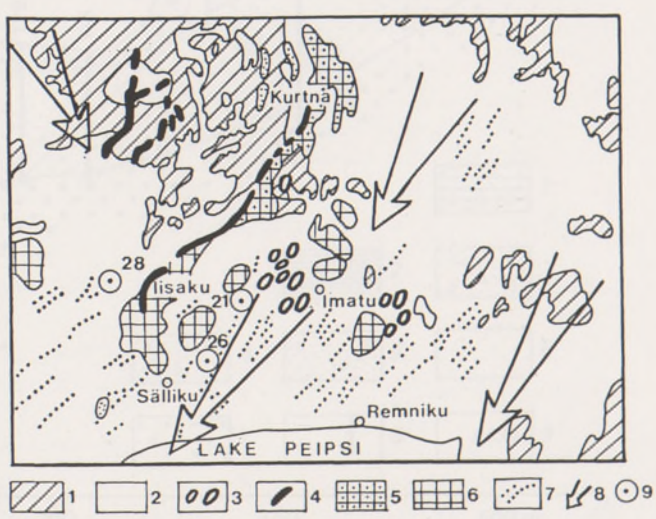


Fig. 104. Glacial topography and inland dunes in the Iisaku-Illuka area: 1 – till plain; 2 – glaciolacustrine plain; 3 – drumlin; 4 – esker; 5 – kame; 6 – limnoglacial kames; 7 – dunes; 8 – ice movement; 9 – boreholes (after Raukas *et al.* 1988).

Photo 41. Small inland dune overgrown with vegetation at Sälliku. Photo by A. Raukas.

there are no suitable dating methods either. The stratigraphy of the aeolian deposits in the coastal area of the Baltic Sea is based mainly on the dating of buried organic layers and geomorphological evidence, which allow to establish the approximate age of the aeolian sediments and related relief forms. In several cases in Estonia, the thermoluminescence (TL) method has also been used in dating of aeolian sands, but the application of the method has been limited by uncertainties associated with the residual TL signal detected in contemporary sediments (Raukas *et al.* 1988). Recently, a dating method based on new luminescence redout technique using a laser (Huntley *et al.* 1985) of infrared (IR) diodes with λ=880 nm (Godfrey-Smith *et al.* 1988) to stimulate electrons from light-sensitive traps in quartz and feldspars was developed by Galina Hütt (Hütt & Jaek 1996, Hütt *et al.* 1988) in the Institute of Geology at Tallinn Technical University. Almost all other steps in the infrared optically stimulated luminescence (IR OSL) dating procedure are similar to those used in the conventional TL dating of unheated sediments. However, the new redout technique can utilize light-sensitive traps for dating because of the loss of electrons, which takes place shortly after the sediment has been exposed to direct sunlight. Hence, the residual signals for modern sediments can be easily zeroed. This, in principle, will enable to date aeolian and other young sediments in a rather reliable way (Raukas & Hütt 1998). Experiments of Wintle *et al.* (1994) have shown that the method enables to date very young samples; *e.g.* she has dated a surface sand sample with an age of 40±15 years.

In the area of inland dunes in the Peipsi Basin, the samples for TL dating were taken from 3 profiles (Fig.104, 105). The obtained results confirmed Holocene age of aeolian deposits and suggested recurrent redeposition of initial Younger Dryas glaciolacustrine and aeolian sediments (Raukas *et al.* 1988). Dune sands, 2 km to the north from Lake Imatu, gave two consistent dates: 4000 years at a depth of 0.4 m and 5900 years at a depth of 2.5 m (Fig. 105). Dune sands west of the Iisaku Settlement yielded three dates: 5200 years at a depth of 0.6 m, 6900 years at a depth of 3.5m and 3000 years at a depth of 11 m (Fig. 105). The latter date is not reliable, but the reason for that is not clear. In the third profile near Valgesoo, the date of 13,680 years from a depth of 0.4 m is not reliable (Fig. 105), all other dates – 4700 years (depth 2.4 m), 5400 years (3.7 m) and 7100 years (4.8m) are consistent which each other. The OSL date on sand from a depth of 3.2 m in a dune, stabilized with pine forest in the surroundings of Sälliku, is 8000 years, demonstrating that the last redeposition of the material took place at the end of Boreal time. The dates seem rather reliable and indicate different phases of aeolian activity.

As a result of uneven migration of inland dunes, during which the central ridge blowouts moved downwind with a steep lee-side slope and the low-lying arms fixed with vegetation lag behind, parabolic dunes came into being (Photo 40). Both ridges and fields of parabolic dunes contain multiple reaping-hook segments. Assuming that the evolution of parabolic dunes takes place through blow-out migration, the long axis of parabolic dunes or the line that bisects the angle of the covering dune arms, can be taken as pointing in the direction of the main dune-forming wind (Galloway & Carter 1994).

Coastal dunes occur in different parts of the lake basin, more often at the eastern and northern coasts, *i.e.* on the west-facing shores where the prevailing winds are westerlies and south-westerlies. The size of a dune here is a function of many factors, *e.g.* the length of the period of consistent wind regime under which it grew. Dunes in the Peipsi Basin are mostly ephemeral and adjusted to diurnal wind regimes.

The northern lake coast abounds in the so-called basket-trap dunes with specific morphology (Orviku 1933). These dunes were formed behind rather huge parabolic wind ditches (Photo 42), which developed at the sites where the vegetation cover had been locally breached as the result of damage to vegetation by wave erosion or by clearing, grazing, burning or trampling (Photo 43). Onshore winds have developed

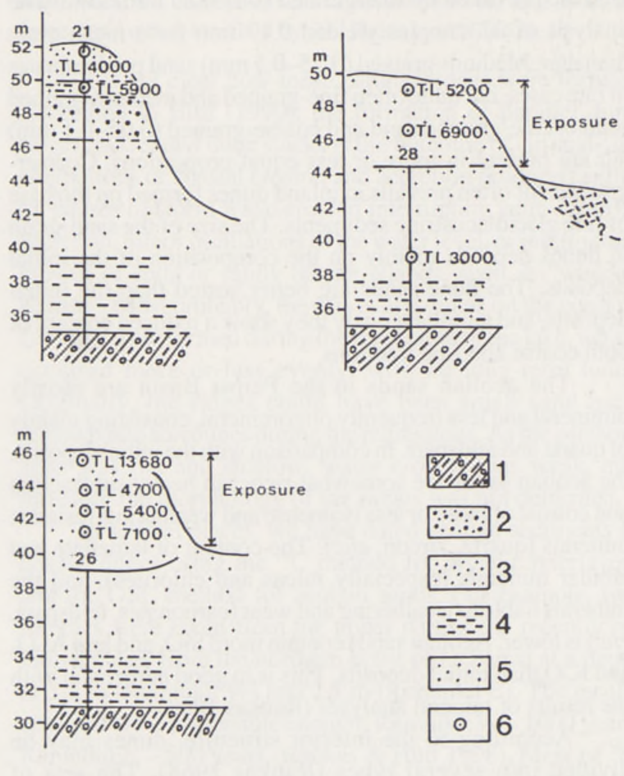


Fig. 105. Sections dated with TL method. For location of sections and boreholes see Fig. 104: 1 – till; 2 – glaciofluvial deposits; 3 – glaciolacustrine sand; 4 – glaciolacustrine silt; 5 – aeolian sand; 6 – location of dated samples (after Raukas & Hütt 1998).

Photo 42. Basket-trap dune formed behind a parabolic wind ditch at Kuru. Photo by A. Raukas.

Photo 43. Man-induced wind ditch at Remniku. Photo by A. Raukas.

blowouts, hollows cut through a foredune or ancient dune, with a nose of sand spilling inland. Most blowouts feature an asymmetric profile, with moisture and frost producing over-steepened south-facing slopes that are subject to seasonal collapse. Detailed studies of blowout morphology have revealed a complex pattern of sediment redistribution involving both aeolian and slope processes. Migration of sand manifests itself most clearly from June to September, when the load of holiday makers is high and much trampling is done.

During long-shore migration, sandy sediments were subject to mechanical and mineralogical differentiation as a result of which well-sorted (S_o=1.2–1.3) sand accumulated on the shore (Tavast & Raukas 1996). The dunes in the Peipsi area are prevailed by fine-grained (0.1–0.25 mm) sand. The analysis of 27 samples yielded 0.17 mm for a mean grain diameter. Medium-grained (0.25–0.5 mm) sand predominates in rare cases, but quite often fine-grained and medium-grained sand or fine-grained sand and coarse-grained (0.05–0.1 mm) silt are present in more or less equal proportions. Coarser-grained silt often prevails in inland dunes formed on the base of fine glaciolacustrine sediments. The size of the sand-grain in dunes depends mainly on the composition of the initial deposits. The dune-sands are better sorted than the initial deposits, and simultaneously they show a reduced content of both coarse and fine fractions.

The aeolian sands in the Peipsi Basin are mostly bimineral and less frequently oligomineral, consisting mainly of quartz and feldspars. In comparison with the initial deposits, the aeolian sands are somewhat richer in heavy subfraction and consist of 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. Aeolian sands contain more SiO_2 and less Al_2O_3 and K_2O than initial deposits. This is in good agreement with the results of mineral analyses (Raukas 1968).

According to the interior structure, dunes may be divided into several types (Raukas 1968). The sets of transversally situated laminae are more or less symmetrical in relation to both slopes. A wedge-like inclined stratification according to the classification by L. Botvinkina (1962) prevails. Concavo-convex wave-like sets of laminae are rare. This type of lamination is more frequent in longitudinal sections. Figure 106 and Photo 44 demonstrate the longitudinal section of a dune close to the mouth of the Alajõgi River. Figure 107 shows a typical stratification across the dune at Sälliku. Most of dunes in the Peipsi Basin are relatively low and rendered stationary by vegetation (Photo 45).

The water level in L. Peipsi undergoes considerable fluctuations, being lowest in summer and autumn/winter periods. Long-term observations have revealed a distinct rhythmicity in lake-level fluctuations, which may be due to solar activity (see Ch.1). During low stands of water like it was, for instance, in 1996, small foredunes will form on the coast. Soon these dunes become overgrown with plants (Photo 46) and are eroded during high stand of water.

Lake hummocky ice may also cause great damage to dunes, even at a distance of tens of metres from the lake shore (Photo 47). The accumulation of snow and sand together with spring thaw may break trees growing on the slip-face of dunes.

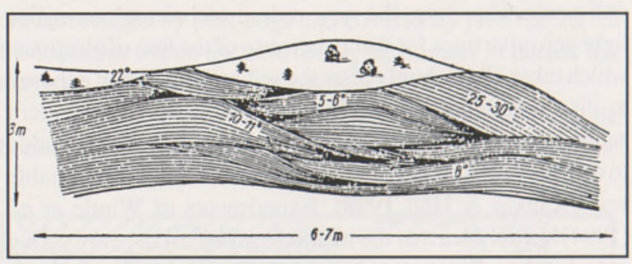


Fig. 106. Longitudinal section of the recently formed coastal dune near the mouth of the Alajõgi River (after Raukas 1968).

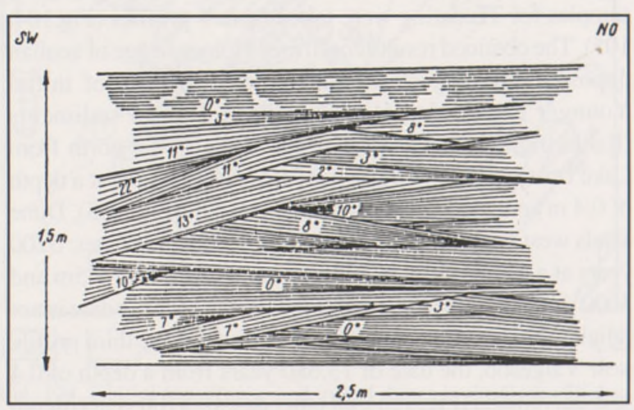


Fig. 107. Cross-section of the continental dune at Sälliku (after Raukas 1968).

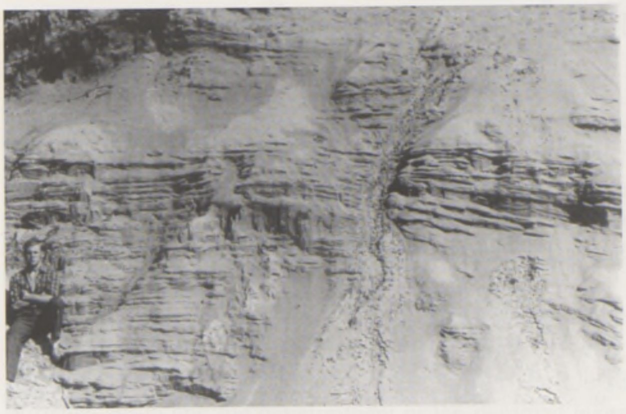


Photo 44. Stratification in the longitudinal section of the dune near the mouth of the Alajõgi River. Photo by A. Raukas.

Photo 45. Planted pines on the top of the recently moved parabolic dune at Kuru. Photo by A. Raukas.

Photo 46. Small foredunes formed during low stand of water in 1996-97 at Remniku. Photo by A. Raukas.

Photo 47. Hummocky lake ice can cause great damage to coastal dunes even at a distance of tens of metres from the lake shore. In foreground a tree scrached by ice. Photo by A. Raukas.

Dunes build up behind sandy beaches in the Peipsi Basin typically form one foredune ridge, colonized and stabilized by grassy and shrubby vegetation. But where the coastline has prograded as a result of sand accumulation, there is often a series of sucessively formed parallel foredunes (primary dune ridges). In the most classical way they are assembled in the small (2.5 km^2) Järvevälja (Rannapungerja) Dune Field, west of the Rannapungerja River mouth on the north-west coast of L. Peipsi *s.s.* This dune field consists of 10–14 parallel ridges. The inclination of their opposite sides displays only small variations (2–8°). The tilt of the windward slopes is generally 4–9°, and leeward slopes 6–17°. The dunes are mostly 1–2, seldom 3 m high; the width ranges from 20 to 30 m (Fig. 108). The morphology and structure of the dune ridges are quite similar. They consist mostly of well-sorted (S_0=1.15) fine to medium (Md=0.18–0.29 mm) quartz (83.5–97.5, in average 89.4%) sand, in which the average feldspar content is 2.0–15.7% (Martin 1984).

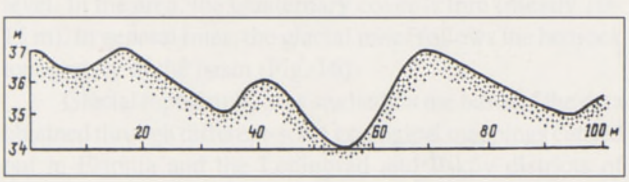


Fig. 108. Profile of the Järvevälja dune field (after Raukas & Tavast 1991).

It is not clear how these parallel ridges were formed. According to Bird (1990), the formation of parallel dune ridges and related dune slacks is the outcome of cut-and-fill, alternations of coastal erosion and accretion produced either by phases of storm wave erosion interrupting calm-weather accretion, minor oscillations of the water level, or intermittent growth of spits or sandy barrier islands, such as "wadden islands". Most probably, the Järvevälja dunes on the coast of L. Peipsi were formed during the regression of the lake, which occurred more or less evenly, with no long-term halts. Originally, the ridges could have been small sand bars, reblown into foredunes during the regressions. As the offshore zone was flat and shallow, water erosion was weak, and therefore the morphology of the ridges was not deformed.

In several places, we have dated aeolian sediments in coastal dunes using the ^{14}C method for organic interlayers and the OSL method for aeolian sands. For example, two samples were taken from the upper part of the recently re-blown dunes from the northern coast of L. Peipsi at a depth of 1.5 m (Remniku W) and 2.5 m (Remniku E). The results obtained were reasonable enough: Remniku W – 1000 years, Remniku E – 500 years (Raukas & Hütt 1998). Most of ^{14}C dates obtained from organic intersand layers are around 1000-2000 years and demonstrate repeated resedimentation of aeolian sand. The results also indicate a peak in aeolian activity approximately 150–200 years ago. *In situ* organic layers and ancient soil horizons typically indicate periods of dune stability (Photo 36).

The Basin of L. Peipsi has always been sparsely populated. Therefore, human impact on aeolian processes has been limited. Currently, areas of recent deforestation, some military areas on the eastern coast and tourist regions, mainly on the northern and eastern coasts of the lake, may display uncovered surficial sand (Photo 48).

Photo 48. Aeolian sands at Spitsyno set into motion as a result of human activity. The blowout area with buried tree stumps. Photo by A. Miidel.

7. DEGLACIATION OF THE LAKE BASIN

Hausen (1913) was the first to study the geomorphology of L. Peipsi Basin. He distinguished several glacial marginal formations and glacial lakes related to these features. He paid attention to the uneven uplift of the Earth's crust in the lake basin during Late Weichselian and Holocene.

Tammekann (1926) studied the Late Weichselian and Holocene levels of Pihkva and Peipsi lakes in more detail. Ramsay (1929) did a great job in reconstructing the course of deglaciation and the development of ice-dammed lakes in Estonia. His contribution was recognized by later investigators (Orviku 1960, Kvasov 1975, Raukas *et al.* 1971).

Markov (1931) and Krasnov (Markov & Krasnov 1929) studied the Neva and Luga glaciolacustrine basins and induced the varve chronological research in North-East Estonia. The studies of this kind were carried on by Rähni (1963), Karukäpp (Bakhmutov *et al.* 1987) and Hang (1997).

In the 1960s, the glacial geology and deglaciation history of the L. Peipsi Basin were dealt with in the publications by Kajak (1964, 1967), Rähni (1967), Raukas and Rähni (1969, 1981), Raukas *et al.* (1971) a.o. Bakanova and Malakhovskij (1969) and Isachenkov (1969, 1972) described the geomorphology and geology of the eastern part of the basin. Kvasov (1975) presented a wider treatment of the history of the ice-dammed lakes in this area. He proposed to name the glacial lake, which had existed there during the Neva Stage, after Ramsay. This lake was the first stage of the Baltic Ice Lake (Kvasov & Raukas 1970).

In the 1970s and 1980s, the researchers of the Quaternary Department of the Institute of Geology of the Estonian Academy of Sciences studied the geology of the lake and lake basin. The results obtained were presented in several reports and publications (Miidel *et al.* 1975, Raukas 1981b, Karukäpp 1985, 1987, Raukas *et al.* 1989).

The conclusions drawn on the genesis of the lake basin were based on both geological data and satellite imagery (Mozhaev & Mozhaeva 1978).

In spite of the long history of investigation and the great number of investigators involved, many problems of topical interest in the deglaciation history of the Peipsi Basin have still not been solved.

Glacial topography and palaeoglaciological reconstructions. The area needed for the reconstruction of the deglaciation history includes the palaeoarea of the Peipsi Glacier Stream which was situated between the Baltic and Ladoga streams (Fig. 109). For this reason, the border of the basin does not coincide with any certain hypsographic level. In the area, the Quaternary cover is thin (mostly 10–15 m). In general lines, the glacial relief follows the bedrock topography of the basin (Fig. 16).

Glacial topography was studied on the basis of the data obtained through different-scale geological mappings carried out in Estonia and the Leningrad and Pskov districts of Russia. Topographic maps (scale 1:25 000), aerophotos (1:18 000 and 1:37 000) and satellite imageries were other important sources of information.

During the fieldwork, instrumental topographic measurements were carried out at several sites, including dunes north of L.Peipsi, the glacial valley of the Kuna River, Mustoja Kame Field, the Piusa, Võhandu, Ahja and Optjok river valleys, *etc.*

Clast fabric was measured in glacial deposits to study the lithology and glacial dynamics. The till fabric analyses were performed on the horizontal shelves in the section walls. In elongated clasts, the orientation and dip of the longest axis were measured. During each analysis, at least 100 clasts were measured.

The till plains have usually a levelled surface. However, quite frequently they are undulated or drumlinized. The drumlins are associated with the Peipsi Glacier Stream and Lobe; they are practically absent in the north-west of the Iisaku – Illuka Interlobate Complex (Fig. 110).

Like in the Saadjärv Drumlin Field, the drumlinized surfaces near Kallaste are of SE orientation (140°), following the bedrock topography (Fig. 16). In the vicinity of Pala, the orientation of clasts in till coincides with the orientation of the landforms. The same direction of the oriented elements in the till surface topography can be found east of the Plyussa River and further to the south. As an exception serve the 15 kilometres of the coastal zone of L. Peipsi, were another direction (110–120° to SW) of drumlinization is dominating (Fig. 110). This natural border between two different directions of drumlinization could be taken as the eastern limit of the Peipsi Glacial Lobe Depression.

In the Kokora – Alatskivi area on the western coast and between Beshkino – Kuna –Torokhovo on the eastern coast, the moraine topography is more dissected. The hummocky moraine topography of both sides associates with glacial valleys (Kuna Valley), glaciofluvial plains or eskers (Torokhovo Esker). This belt has been considered to mark the Sakala Stade of the glacier retreat (Raukas *et al.* 1971). However, it should be stressed here that end moraines are not found in this complex of glacial landforms (Fig. 111, see inside back cover).

The till surface of the Ugandi Plateau is cut by valleys and valley-like features. The lineated elements of the topography, including drumlins, drumlin-like elevations, oriented slopes, valleys and elongated depressions have two main directions: from north-west to south-east or from north-east to south-west (Fig. 111). The orientation of the relief becomes more distinct in a southerly direction (between the Ahja and Piusa rivers).

The glacial marginal formations along the line Mehikoorma – Pnevo – Samolva – Remda are low (less than 5 m), gently sloping and relatively wide elevations, partly buried in mires. Abrasion by lake water has stressed the topography of drumlins on the coastline of islands and peninsulas (Mtezh, Kolpino, Lis'e, Kamenka) or end moraines (Talabskij Islands) forming escarpments that rise 15–19 m above the lake level (Fig. 111).

Interpretation of the glacier marginal positions (Fig. 112) shows that there is a great diversity of opinions as to the location of the ice margin. According to Malakhovskij

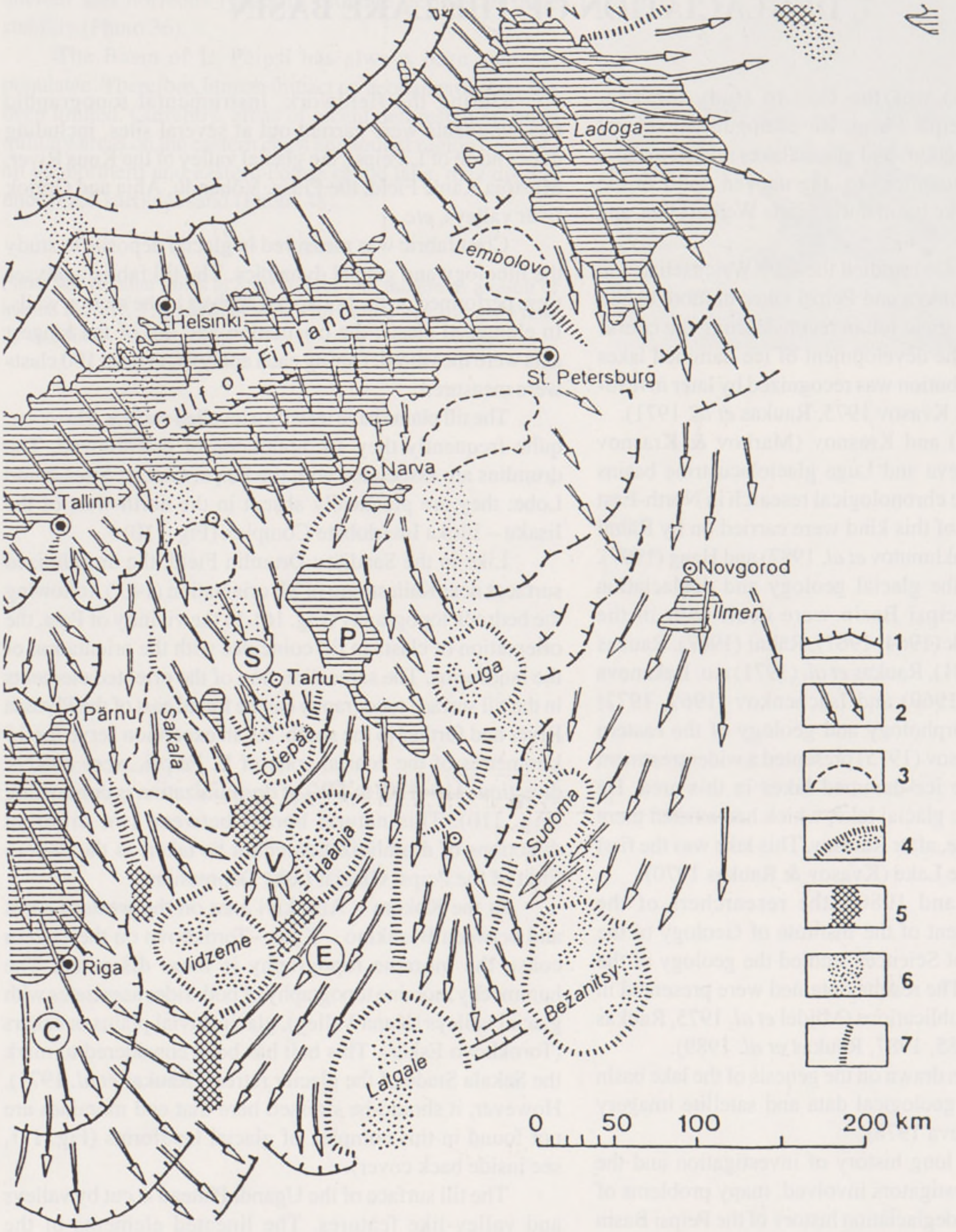


Fig. 109. Glacial dynamics and morphogenesis in the southeastern sector of the Scandinavian glaciation in Gotiglacial time (after Karukäpp 1996): 1 – glacier margin; 2 – ice flow direction in the final stage of its activity; 3 – plinth-type bedrock upland of glacier erosion; 4 – accumulative insular heights; 5 – interlobate complex of landforms; 6 – ice shed between glacier flows; 7 – local ice shed. The numbers in circles mark lobe depressions: (E) – Eastern Latvian, (V) – Võru–Hargla, (P) – Peipsi, (C) – Central Latvian, (S) – Saadjärv Drumlin Field.

and Bakanova (1969), a new extensive readvance of the ice sheet took place in the Neva (Pandivere) Stage. Apukhtin and Krasnov (1967) and Isachenkov (1972) stated that there was a far-forward-extending glacier lobe of the Neva (Pandivere) Stage. Recent interpretation of the GIS-based data and satellite imagery (Boulton & Hulton 1998) indicates big variations in the flow directions on the Fennoscandian Shield, but subparallel streaming on the sedimentary bedrock in the northwestern part of the East-European Platform, where the local streams were unstable and short-living. Based on this data, the pattern of the glacier retreat was reconstructed. As a result, the more or less straight sub-parallel lines (normals) of the glacier margin position were drawn (Boulton & Hulton 1998).

Local directions of the Late Weichselian ice movement are clearly recorded by the drumlin formation pattern (Fig. 110). Two well developed ice marginal belts are recorded by interrupted end moraines via Talabskij Islands in L. Pihkva and Mehikoorma – Pnevo – Jamm in L. Lämmijärv (Fig. 111). The complicated Iisaku – Illuka end moraine – esker – kame system formed as an interlobate complex between the Peipsi (moving to SW) and East-Pandivere (moving to SE) glacier lobes (Raukas *et al.* 1971) and could be conventionally connected with the Pandivere (Neva) Stage.

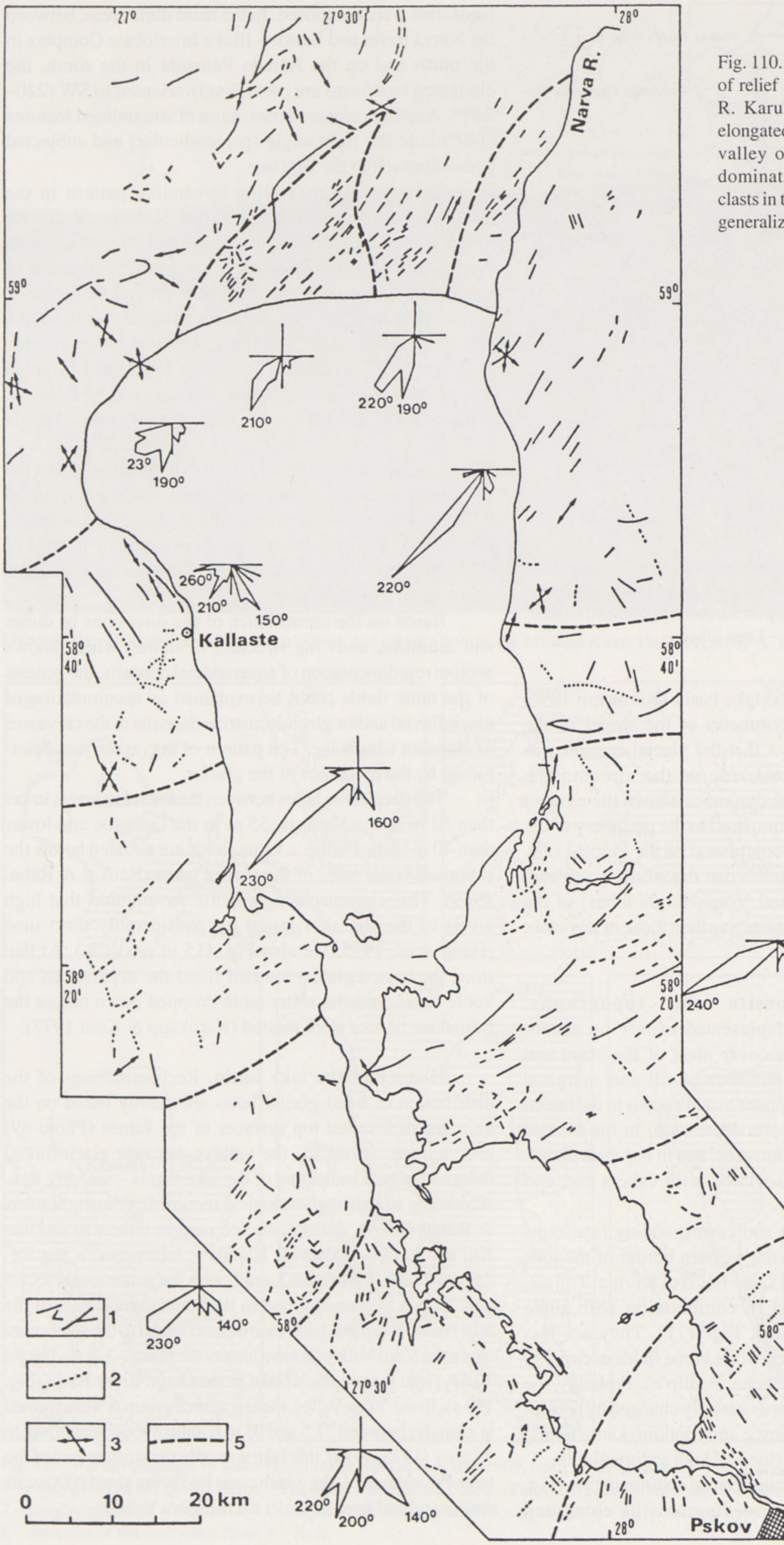


Fig. 110. Orientation of elongated forms of relief and clasts in till. Compiled by R. Karukäpp: 1 – crests of positive elongated forms; 2 – bottom line of a valley or elongated depression; 3 – dominating orientation of elongated clasts in till; 4 – border of the area for the generalization by a rose diagram.

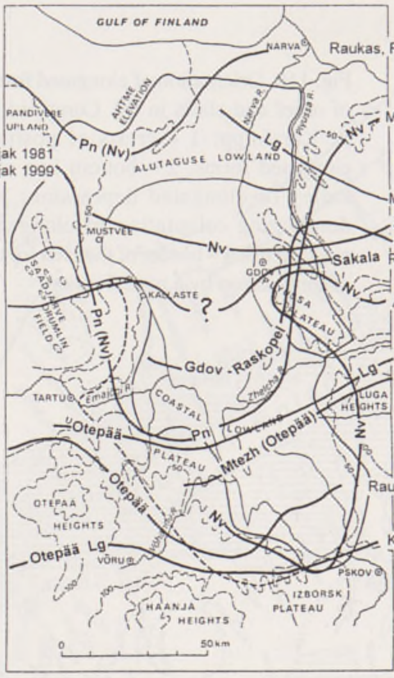


Fig. 112. Interpretation of glacial marginal zones in the area of the Peipsi Basin by different authors on the background of landscape regions. Compiled by R. Karukäpp on the basis of published sources. Pn – Pandivere, Nv – Neva, Lg – Luga Stage.

The cross-sections across the lake basin (Karukäpp 1996, Fig. 2) demonstrate the asymmetry of the slopes of the bedrock relief. This suggests that the glacial erosion was more intensive on the west side of the stream. The reconstruction of the glacial dynamics shows the evident activity of the Peipsi Lobe compared to the periphery of the Baltic Ice Stream west and south-west of the Peipsi Lobe. The reconstruction demonstrates that the westerly oriented portions (Eastern Latvian and Võru–Hargla lobes) of the Peipsi Stream were much more active than those of any other direction (Karukäpp 1999).

Aqueoglacial deposits and topography. Aqueoglacial deposits are represented mainly by glaciolacustrine sediments, which cover most of the basin and considerably levelled the surfaces of till and marginal formations. Glaciofluvial deposits usually occur in the lateral and marginal areas of the lobe depression, in the formed kame fields or glaciofluvial terraces, and in outwash plains. More seldom, they are associated with eskers and end moraines.

Glaciofluvial deposits and corresponding landforms are of wider distribution on the western border of the lobe depression (radial belt of the kame fields of Kurtna, Tõljase, Kaiu, Selgise and Mustoja in combination with some interlobate ridges and eskers, Fig. 111). They are less frequent on the eastern border, while kame fields occur only in the southern part (Gorodishche, Podlip'e). Probably, the topography of some of them was strongly changed by aeolian processes (Podlip'e, Podporov'e and Plotkino kame fields) and levelled by glaciolacustrine and lake sedimentation.

Streamlined areas of aqueoglacial sedimentation (Fig. 110) were analysed and rose-diagrams of elongated landforms were composed. In the main dune areas, between the Narva River and Iisaku – Illuka Interlobate Complex in the north and on the Remda Peinsula in the south, the elongated landforms are prevailingly oriented to SW (220–240°). Another measured maximum of streamlined features (140°) is at the right angle (perpendicular) and subjected (subordinated) to the first one.

The strongly prevailing streamline pattern in the topography refers to the controlled features of glacier disintegration. Whether the streamline pattern on the Remda Peninsula reflects the direction of the glacier movement or is perpendicular to it, has not yet been proved. In favour of radial orientation testifies the circumstance that the same direction (about 230°) of the streamlined topography, represented by drumlins (Karukäpp 1985) and oriented pattern of soil varieties (Raukas 1961), could be followed further to SW (Mehikoorma, Räpina and Võru – Hargla Valley).

Bakanova and Malakhovskij (1969) pointed out the SW direction of the glacier movement 40–50 km south of Pihkva. But between this area and the Remda Peninsula there is evidently the drumlinized southern coast of L. Pihkva (Figs. 110, 111) where the direction of the glacier movement was 140° to the south-east.

Based on the coincidence of the directions of dunes and drumlins, and the structure of dunes, which shows aeolian resedimentation of aqueoglacial deposits, the genesis of the dune fields could be explained as accumulation of glaciofluvial and/or glaciolacustrine deposits in the crevasses of stagnant (dead) ice. The pattern of crevasses was determined by the dynamics of the glacier.

The deep kettle holes between the kames at levels lower than 62 m in the Mustoja, 55 m in the Goristoe and lower than 40 m in the Podlip'e kame fields are situated below the supposed water table of the glacial lakes (Raukas & Rähni 1969). This circumstance supports the opinion that high levels of the ice lakes lasted for considerably short time (Hang *et al.* 1995, see also Fig. 115 in next Ch.). At that time, the buried glacier ice still filled the depressions and kettle holes, and the water table dropped down before the buried ice blocks were melted (Karukäpp & Kont 1977).

History of the lake basin. Reconstructions of the distribution of local glacial lakes are mainly based on the altitudes of levelled top surfaces of the kames (Photo 49) and terrace levels in the valleys, because glaciofluvial deltas – the best indicators of the lake levels – are very rare. According to palaeoglaciological reconstructions by Raukas & Rähni (1969), during the ice margin retreat to the line Kul'e – Lis'e – Talabskij Islands – Jelizarovo, a big ice-dammed lake (Pihkva Ice Lake I, with the water levels 95, 85 and 75 m a.s.l.) was formed in the southernmost part of the lake basin. From this lake, meltwaters flowed to the south-west along the Võru Valley (known also as the Piusa – Võru – Hargla Valley) into the proglacial lake in the Gauja River Basin (Fig. 113 a). In the Võru Valley, the terraces of group A were formed at altitudes between 71.5 and 95 m (Liblik 1966). According to Kvasov (1975, 1979), this lake was only the western part of the large Privalday Ice Lake which came into being about 2000 years earlier and had later an outlet via the Võru Valley.

Photo 49. Levelled surface of the kame hillock and a kettle hole in the Mustoja Kame Field. Photo by A. Raukas.

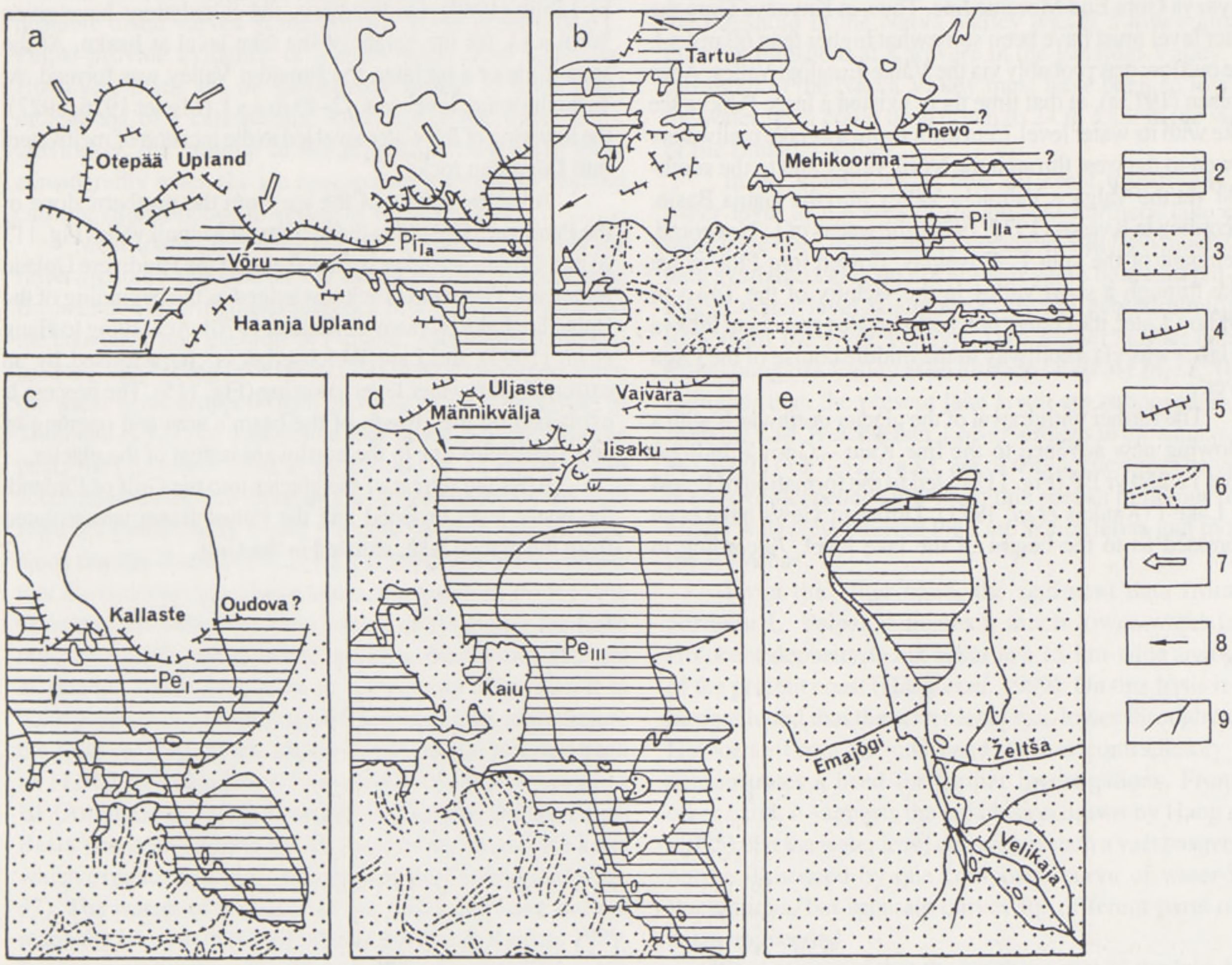


Fig. 113. The development of Lake Peipsi (after Raukas & Rähni 1969, from Miidel & Raukas 1997): a – Pihkva Ice Lake I (phase Pi_Ia); b – Pihkva Ice Lake II (phase $Pi_{II}a$; c – Peipsi Ice Lake (phase Pe_I); d – Peipsi Ice Lake (phase Pe_{III}); e – Lake Small (Old) Peipsi. 1 – glacier; 2 – proglacial lake; 3 – mainland; 4 – margin of the active glacier; 5 – dead ice; 6 – meltwater valley; 7 – direction of the ice movement; 8 – direction of the meltwater flow; 9 – river.

It is possible that the terraces in the Võru Valley are glaciolacustrine (Lepland 1991, Hang 1995) not glaciofluvial in origin, because they consist of horizontally bedded fine-grained silts and sands. Although the terraces have no inclination towards the supposed flow direction, the Võru Valley was a wide strait which connected ice lakes in the east and west (Fig. 113a). Further studies are needed to estimate the origin of the terraces in the valley.

Afterwards, when the lake level dropped to 75 m, the 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, the water level sank to 62–60 m a.s.l. and Pihkva Ice Lake II came into being (Raukas & Rähni 1969, Raukas *et al.* 1971, see also Fig. 113 b). The inflow was from the surrounding heights, occupied by dead ice. The formation of the Ahja and Võhandu valleys started. There was probably no outflow from the Pihkva Ice Lake II.

The next belt of ice marginal formations has been established on Piirissaar Island and at Knyazya Gora (east coast of L. Peipsi). The water level was at a height of 60 m a.s.l. (Raukas & Rähni 1969). According to Hang (1995), the terraces of group B in the Piusa and Ahja 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. After Hausen (1913a), 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 south-west via the Valga – Valmiera Valley into the Gauja Basin. According to Kvasov (1975, 1979), 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.

The further withdrawal of the glacier northwards with a following new advance to the line Kaiu – Gdov (Oudova) about 12,250 yr BP (Fig. 113 c) led to the formation of Peipsi Ice Lake I (Raukas *et al.* 1971). Different views have been expressed as to the height of the lake level. According to Raukas and Rähni (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. As is known, Peipsi Ice Lake corresponds to the South Peipsi Ice Lake by Hausen (1913a) where the water level was only 36–37 m a.s.l., but Hausen failed to establish any shoreline there. Hang (1995, Hang *et al.* 1996) thinks that the terraces of group C in the Ahja and Piusa valleys were formed when the glacier came to a halt on 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. (Miidel & Raukas 1997).

During the Pandivere (Neva) Stade when the glacier readvanced again and came to a halt along the Männikvälja – Iisaku – Vaivara ice marginal formations (Fig. 113 d), the third phase of the Peipsi Ice Lake (Pe_{III} after Raukas & Rähni 1969) formed a single body of water with glacial lakes in the east (Luga, Neva a.o.). Kvasov (1975, 1979) termed it L. Ramsay which corresponded approximately to Great Peipsi Ice Lake by Hausen (1913a). According to Raukas and Rähni (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. But after Hausen (1913a), 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, which was determined between Kallaste and Kavastu by Liblik (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); the lowering of the water level led to the 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. 113 d) and Uljaste, ice lakes west and east of the Pandivere Upland joined up. The event is acknowledged as the beginning of the Baltic Ice Lake (Kvasov & Raukas 1970). According to Hang *et al.* (1995), the Late Weichselian is characterised by an extremely fast water level lowering (Fig. 115). The process is explained by the growth of the basin's area and opening of new thresholds due to the northward retreat of the glacier.

After the retreat of the glacier into the Gulf of Finland, the water level dropped and the Peipsi Basin was isolated from the glacial lake, situated in the Gulf.

8. HOLOCENE HISTORY OF THE LAKE

The development of L. Peipsi in the Holocene has been controlled by glacioisostatic movements and the climatic factor. The vast lake basin is elongated approximately in the direction of the tilting of the Earth's crust. Therefore, the glacioisostatic factor has played the leading role in changing the water level and configuration of the lake. However, the data concerning crustal movements is very scanty and often contradictory, which hampers estimating the actual effect of this factor in the evolution of the lake (Ch. 2.5). Synchronous lake level changes in the small lakes of the adjoining area and in L. Peipsi (Miidel *et al.* 1995; Hang *et al.* 1995) provide evidence of the significant role of climate. In view of this, the reason for water level changes in L. Peipsi during the Holocene is postulated to be hidden in a total combined effect of climate and tectonic movements (Miidel *et al.* 1995).

After the retreat of the Late Weichselian inland ice into the Gulf of Finland the water level dropped and the Peipsi Basin isolated from the proglacial lake (Fig 113 e). The exact time of the event is difficult to determine. It is even harder to say whether the water level dropped abruptly or slowly. The character of the glaciolacustrine varved clay with the gradual transition from normal varves to diffused unstable distal varves in the bottom of the present-day L. Peipsi provide evidence of a slow water level lowering. However, there are no geomorphological or geological indications of a catastrophic lowering of the water level. Obviously, the volume of the proglacial lake basin grew considerably when the ice retreated to the Gulf of Finland causing simultaneous lowering of the lake level.

Most likely, the lowering of the water level and the faster isostatic rebound in the northern part of the Peipsi Basin ended up with the emergence of a watershed between the lake basin and the Gulf of Finland. As a result, L. Peipsi isolated. According to Markov (1931), this happened when the water level dropped from 38 to 32 m a.s.l. Referring to Thomson (1929) he mentioned that in the Auvere Bog the peat, with its lower limit at an altitude of 32 m, started to accumulate in the Pre-Boreal. This is supported by the data from the Puhatu Mire where the peat has also accumulated since the Pre-Boreal (Ch. 2.4.2.2). This data demonstrates that the isolation must have taken place earlier. Prior to the drainage, the transgressional waters of the Baltic Ice Lake reached the altitude of 40–35 m in the Narva area (Kessel & Raukas 1979) and 28 m at Kingissepp (Fig. 37). It is quite possible that the isolation of L. Peipsi took place before the final drainage of the Baltic Ice Lake that is dated back to 11,565 calibrated yr BP equivalent to 10,690 varve yr BP (Andrén *et al.* 1999). It has been postulated that the Peipsi Basin was a bay during the B_I stage of the Baltic Ice Lake which expanded in front of Salpausselkä I formations (Fig. 37). The waters of B_{III} stage of the Baltic Ice Lake did not reach the lake basin (Orviku 1960, Kessel & Raukas 1979, Svensson 1989) (Fig. 37). Thus, L. Peipsi isolated when the water level in the Baltic Ice Lake dropped from B_I to B_{III} level. This could be taken as the initial stage of L. Small Peipsi. The isolation of the Narva – Luga Lowland was completed with the emergence of the Baltic Klint during the final drainage.

But still, the lowering of the water level in the Peipsi Basin down to the lowest level of L. Small Peipsi (24 m a.s.l., *i.e.* 6 m lower than the present lake level) in the mouth of the Emajõgi River is not yet convincingly explained. There still remains the question in which way the water of L. Peipsi discharged from the lake, because at that time the southern part of the lake basin was at a higher altitude than the northern one and undisputable outlet is missing.

The waters could have moved in two possible directions – through the Narva Valley to the north or through the Emajõgi Valley to the west. The altitudes of the thresholds in both valleys – 26–28 m and *ca* 29 m, respectively (Fig. 37) – are poorly investigated. In the Late Weichselian, the whole of North-East Estonia was flooded by the waters of the proglacial lake. If the above-described situation holds true, then during the emergence of the Narva – Luga Lowland as a watershed the main amount of the proglacial water probably discharged through the lowland north of the threshold in the Narva Valley. This is supported by the shoreline data of the Baltic (Orviku 1960, Pärna 1960, 1962, Kessel & Raukas 1979, a.o.) which demonstrates that the threshold in the Narva Valley was at a lower altitude than the threshold at Kärevere in the Emajõgi Valley. The threshold in the Narva Valley must have been so low that the water level could drop down to 24 m a.s.l. in the mouth of the Emajõgi Valley.

In the conditions of thus different topography, the water level in the area of the present northern lake coast must have been higher than nowadays. But how much? The data available today do not allow it yet to be estimated. The higher water level is supported by the geomorphological data from the river valleys. Two higher terraces in the Rannapungerja Valley at an altitude of 40–41 m, *i.e.* 10–11 m higher than the present lake level, are supposed to date from the Late Weichselian. Lower terraces in the same valley at a height of 37–32 m a.s.l. are of Holocene age. The valley is not overdeepened (Fig 103); this speaks in favour of the higher water level during the Late Weichselian and most of the Holocene.

From the other hand, the sediment data from the northern L. Peipsi indicates a much lower water level. Holocene deposits do not occur in a 15-km-wide area south of the present coast (Hang *et al.* 1999). On this basis it may be concluded that the water level was lower throughout the Holocene. From the one hand, such a contradictory data demonstrates a need for further investigations. From the other hand, it supports the conclusion drawn by Hang *et al.* (1995) that the water level changes in such a vast basin could not be illustrated by one composite curve of water-level fluctuations, but separate curves for different parts of the basin are needed.

It is supposed that the southern part of the lake basin dried up at the end of the Late Weichselian and the beginning of the Holocene (Orviku 1960, Raukas & Rähni 1969, Miidel & Tavast 1981, a.o.). However, some researchers believe that it might have happened much earlier – about 12,000 ^{14}C yr BP (Kvasov 1975, 1979, Hang *et al.* 1995). Actually, it was already Ramsay (1929) whose palaeogeographical schemes depicted the southern part of the Peipsi Basin as dry land since the second stage of the Estonian Ice Lake when the ice margin was still close to the Kurtna Kame Field. According to Ramsay, L. Peipsi isolated already during the fourth stage of the Estonian Ice Lake when the ice margin was on the north-west slope of the Pandivere Upland. Kvasov (1975), too, was convinced that the southern part of L. Peipsi up to Piirissaar Island was dry land *ca* 11,900 yr BP although at that time nearly half of Estonia's territory was still covered with ice.

Almost at the same time, L. Small or Old Peipsi (Orviku 1960, Kajak 1964, Raukas & Rähni 1969 a.o.) occupied the northern part of the present-day Peipsi Basin. It is supposed that the Velikaya River with its tributaries (the Zhelcha, the Emajõgi, *etc.*) discharged into L. Small Peipsi (Fig. 113 e).

The southern part of L. Peipsi remained dry during a long period. However, according to Davydova (Ch. 3.2.4.1) there was a cold and deep body of water in L. Lämmijärv during the Younger Dryas. From the Allerød to Boreal Chronozone this part of the basin was dry or there was only a small lake in the deepest part.

The correlation diagram of geomorphological evidence proposed by Hang *et al.* (1995) (Fig. 100) shows that during the isolation of L. Peipsi the water level in the mouth of the Optjok River was *ca* 6–7 m lower than at present. At that time the water level was either 32 m a.s.l. or even lower south of the Baltic Klint. Biostratigraphic data both from L. Lämmijärv (Davydova & Kimmel 1991) and from the mouth of the Optjok River (Miidel *et al.* 1995) demonstrate a low water level during the Pre-Boreal. This suggests that the water level dropped earlier, before the onset of the Holocene.

At the beginning of the Pre-Boreal Chronozone, a shallow lake or rather the system of small lakes existed in the southern part of the basin (Pirrus *et al.* 1985, Davydova & Kimmel 1991, Miidel *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. It is not known what was the height of the water level and whether it differed between lakes, but in the mouth of the Optjok River it must have been 10 m lower than today (Miidel *et al.* 1995). Pollen evidence (Pirrus *et al.* 1985, Davydova & Kimmel 1991, Miidel *et al.* 1995) suggests paludified surroundings and shallow water. Nevertheless, the water depth in the deepest part of L. Lämmijärv must have been about 7–7.5 m (Miidel *et al.* 1995).

The curves of water level changes (Miidel 1981, Davydova & Kimmel 1991, Miidel *et al.* 1995, Hang *et al.* 1995) show that the water level was at its lowest in the Pre-Boreal Chronozone (Figs. 35, 114, 115). However, one has also to consider the unevenly distributed radiocarbon dates and the resultant inaccuracy of the curves. Therefore, an abrupt change in sediment stratigraphy in L. Lämmijärv, where silty deposits became overlain by organic rich silt or silt-bearing fen peat could be interpreted as an evidence of the lowering of the water level at the beginning of the

Photo 50. Endel Rähni (1917–1994) conducted a long-term research into the Quaternary geology of the Peipsi Basin. Photo by A. Miidel.

Photo 51. Reet Pirrus (1935 – 1997). During the course of many years she studied the stratigraphy of bog and lake deposits in the Peipsi Basin and the development of the lake. Photo by A. Miidel.

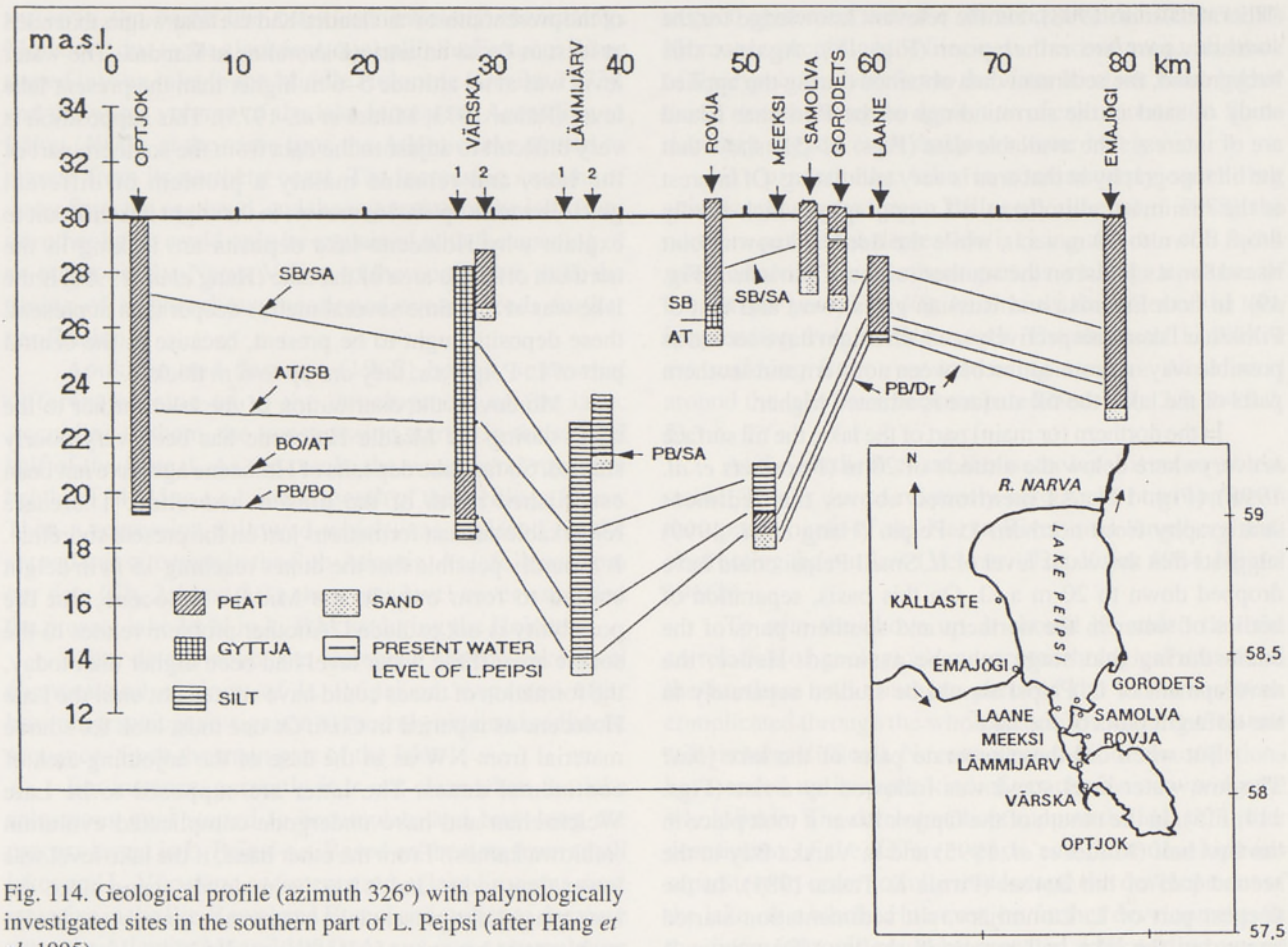


Fig. 114. Geological profile (azimuth 326°) with palynologically investigated sites in the southern part of L. Peipsi (after Hang *et al.* 1995).

Holocene (Davydova & Kimmel 1991, Miidel *et al.* 1995). At the time the fen peat started to form 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 lowermost water level in the Holocene between 10,000 – 9000 yr BP was marked by a break in sedimentation in L. Lämmijärv (Davydova & Kimmel 1991, Miidel *et al.* 1995, Hang *et al.* 1995). The lowest water level in the mouth of the Emajõgi River in the main part of L. Peipsi has been dated to the beginning of the Pre-Boreal Chronozone (Sarv & Ilves 1975). No more datings are available from the period of low water level from the main part of L. Peipsi. However, the recently discovered sediment stratigraphy from the latter area (Hang *et al.* 1999) suggests that during the existence of L. Small Peipsi the water level was even 8–10 m lower than at present in this part of the lake. As an evidence serve the sandy mollusc rich offshore sediments on top of the glaciolacustrine varved clay preceding the accumulation of calcareous gyttja during the period of L. Small Peipsi (Hang *et al.* 1999). According to Niinemets (Ch. 3.2.5), the ostracode fauna in sandy offshore sediments points to the groundwater feeding of the shallow and cold body of water. Thus, if the assumption about a very low water level in L. Small Peipsi (*ca* 20 m a.s.l.) holds true, the knowledge of glacial and Late-glacial topography of the basin is important in order to understand the palaeohydrological situation during that particular period. The glacial topography in the northern part of the lake has been rather well established

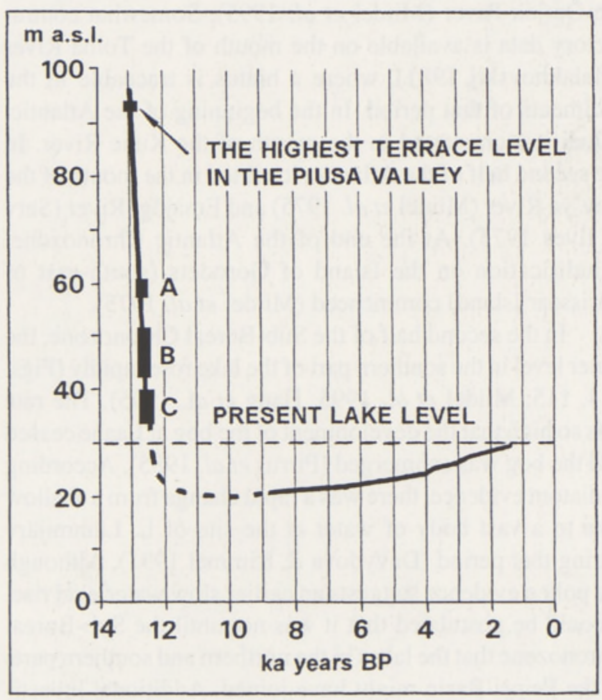


Fig. 115. Proposed lake level changes in the southern part of L. Peipsi. A, B, C – altitudes of the groups of terraces in the Piusa Valley (after Hang *et al.* 1995).

(Noormets *et al.* 1998), but the relevant knowledge for the southern part are rather poor (Fig. 18). Against this background, the sediment data obtained during the applied study of sand in the surroundings of the Piirissaar Island are of interest. The available data (Figs. 41–51) show that the till topography in that area is very undulating. Of interest is the minimum altitude in till topography which hardly drops down to 20 m a.s.l., while the deepest known point rises 18 m a.s.l. just on the southern slope of Piirissaar (Fig. 49). In both Estonian and Russian gates (west and east of Piirissaar Island, respectively), which might have served as possible ways of connection between northern and southern parts of the lake, the till surface is situated higher.

In the northern (or main) part of the lake, the till surface is everywhere below the altitude of 20 m (Noormets *et al.* 1998) (Fig. 18). As mentioned above, the sediment stratigraphy from northern L. Peipsi (Hang *et al.* 1999) suggests that the water level of L. Small Peipsi could have dropped down to 20 m a.s.l. On this basis, separation of bodies of water in the northern and southern parts of the basin during that stage may be assumed. Hence, the development of L. Peipsi should be studied separately in the different parts of the lake.

But when did these separate parts of the lake join? The low water level stand was followed by a rise (Figs. 114, 115). In the mouth of the Optjok River it took place in the first half (Miidel *et al.* 1995) and in Värska Bay in the second half of the Boreal (Pirrus & Tassa 1981). In the deepest part of L. Lämmijärv, silt sedimentation started anew, but the lake still remained shallow (Davydova & Kimmel 1991).

During the Atlantic Chronozone, the accumulation of reed peat with shell fragments continued at the mouth of the Optjok River (Miidel *et al.* 1995). Somewhat contradictory data is available on the mouth of the Tolba River (Malakhovskij 1981), where a hiatus is traceable in the sediments of that period. In the beginning of the Atlantic, paludification started in the mouth of the Kune River. In the second half of the Atlantic, it started in the mouth of the Rov'ya River (Miidel *et al.* 1975) and Emajõgi River (Sarv & Ilves 1975). At the end of the Atlantic Chronozone, paludification on the Island of Gorodets (south-east of Piirissaar Island) commenced (Miidel *et al.* 1975).

In the second half of the Sub-Boreal Chronozone, the water level in the southern part of the lake rose rapidly (Figs. 114, 115; Miidel *et al.* 1995, Hang *et al.* 1995). The rate was so high that the development of the bog at Laane ceased and the bog was submerged (Pirrus *et al.* 1985). According to diatom evidence, there was a rapid change from a shallow lake to a vast body of water at the site of L. Lämmijärv during that period (Davydova & Kimmel 1991). Although the pollen evidence suggests an earlier slow waterlevel rise, it could be postulated that it was not until the Sub-Boreal Chronozone that the lakes in the northern and southern parts of the Peipsi Basin might have joined. Additional investigations in the surroundings of Piirissaar Island are needed for further discussion.

We do not know what happened close to the present northern coast of L. Peipsi at that time. It has been supposed that during the Atlantic Chronozone even the area N–NW of the present coast was flooded and the lake waters extended at least as far as an ancient shoreline at Raadna. The water level was at an altitude 5–6 m higher than the present lake level (Rähni 1973, Miidel *et al.* 1975). This supposition is very difficult to adjust to the data from the southern part of the lake, and remains mainly a problem of different glacioisostatic uplift. Moreover, in that light it is difficult to explain why Holocene lake deposits are lacking in the northern offshore area of the lake (Hang *et al.* 1999). If the lake was at that time several metres deeper than at present, these deposits ought to be present, because in the central part of L. Peipsi *s.s.* they are up to 6 m thick.

Moreover, the distribution of the lake further to the north during the Middle Holocene has been very poorly studied. So far, lake deposits of Holocene age have not been established north of the present shoreline. There are remarkable aeolian formations just on the present shoreline. It is hardly possible that the dunes reaching 15 m in height started to form only in the Middle Holocene, but the possibility is not excluded. Another problem relates to the source area. If the water level had been higher than today, the formation of dunes could have started not until the Late Holocene as reported in Ch.6. Or one must look for source material from NW as in the case of the adjoining area of continental dunes. The latter are supposed to be Late Weichselian and have undergone complicated evolution (reblown kames). From the other hand, if the lake level was lower than today and it has been slowly rising since the Late-glacial, the present lake bottom as a source area is also problematic because of glaciolacustrine clays outcropping in the offshore area (Hang *et al.* 1999). Although these varved clays contain silt and sand accessible for aeolian activity, formation of the area of dunes within a relatively short period of time is doubtful. This also shows that additional studies are needed to study the genesis of the formations under discussion, but still leaves open the question concerning the water level changes in the northern part of the L. Peipsi.

The lake-level changes in L. Peipsi are associated with an intensive glacioisostatic uplift that was faster in the north (Hausen 1913b; Ramsay 1929). The lake-level rise, which has continued during the whole of Holocene cannot be ascribed to the glacioisostatic phenomenon alone. A rapid rise in the second half of the Sub-Boreal is correlated with the similar data from the Võrtsjärv Basin (Fig. 39; Pirrus *et al.* 1993) and also with the data concerning lake level changes in small lakes in the adjoining area (Saarse & Harrisson 1992, Miidel *et al.* 1995, Hang *et al.* 1995). It is obvious that the water level rise was controlled by both glacioisostatic and climatic factors (Miidel *et al.* 1995, Hang *et al.* 1995). But also the hydrology of the whole basin should be taken into account with a special emphasis to the tectonically controlled altitudes of thresholds in outlets and inlets. If to consider that at the end of the Late Weichselian the water level was up to 10 m lower than at present and the reasons or rather the ways of drainage are not yet clear, the reasons for the lake level rise during the Holocene must also be explained. One factor is the warmer and more humid climate and another – the changing tectonic situation, which caused an outflow from L. Võrtsjärv so that the amount of water in the basin has been continuously increasing. Moreover, due to the glacioisostatic uplift the Emajõgi River turned into an inlet in the Middle Holocene (Orviku 1973), and the water volume of the lake increased. According to Rähni (1973), at the same time the Atlantic lake started to regress from its northern coast. The increasing amount of water from the one hand, and the regression of the lake from the other hand, could only be explained with the opening of an outlet via the Narva Valley to the north. But, still, the timing of all the above-mentioned events needs further verification.

Apukhtin and Sammet (1967) hold a somewhat different opinion as to the development of the lake. According to them, the transgression started in the second half of the Boreal. As the result, the water level rose to a height of 34–35 m a.s.l. in the mouth of the Velikaya River. Then a regression followed which was replaced by the second transgression in the Sub-Atlantic. As it follows from our data (Ch. 2.4 and 2.3), the water level never exceeded the present lake level in L. Pihkva during the Holocene.

More details about the Holocene history, lake-level changes and ecology of L. Peipsi are expected after biostratigraphic investigation of recently obtained sediment sequences from the main part of the lake.

For instance, currently it is not clear when the lake calcareous mud started to accumulate and how long the process lasted in L. Peipsi *s.s.* Based on the data from small lakes and L. Võrtsjärv, sedimentation of lake lime was most intensive in the Pre-Boreal and Boreal and started to decrease since the Atlantic (Männil 1964). At any rate, accumulation of calcareous mud show that the lake was shallow which is in good agreement with the ostracode evidence (Ch. 3.2.5). Moreover, according to Niinemets, the continuous presence of *Ilyocypris bradyi* refers to the water level not much higher than 4 metres during the deposition of calcareous mud. Some data may suggest even shallower water levels.

At present the lake is continuously spreading southwards. According to Vallner (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 (Fig. 40). As a result, the water of the lake is spreading from north to south. The shores of L. Pihkva are suffering from ever increasing erosion. Wide areas around the lake have become paludified (Ch. 2.5, 2.4.2.2, 5).

At the northern coast, the water level is more or less stable. This is due to relatively hard rocks outcropping in the upper course of the Narva River, and longshore drift obstructing the outflow (Vellner 1928, Kajak 1964, Miidel 1966b).

To sum the above up, it should be added that the correlation of geological and geomorhological data from the northern and southern parts of the basin has been complicated through the whole history of investigation and still remains difficult. Nevertheless, most of investigations have ended up in some kind of conclusions covering the whole basin. The current authors support the idea of further description of the different parts and their development separately in order to fully understand the development of the lake, as a whole. However, no further developments in the discussed topic could be expected without discovering the bottom sediments of the lake.

9. ANCIENT HABITATION AND THE IMPACT OF PRIMITIVE SOCIETY ON NATURE

In the different parts of the area under consideration, human impact on the environment became apparent, strengthened and turned into the leading factor at different times, depending on natural conditions, increase in population and economic development. Man's interference with the environment has been studied from the historical point of view, but it is very difficult to draw distinct boundaries between the stages of this process. To date, the ever growing impact of human activities on nature is continuing unchecked.

According to archaeological evidence (Jaanits 1959, Jaanits *et al.* 1982, Jaanits 1983), the coasts of the well-stocked L. Peipsi and a variety of game-rich landscapes in its surroundings attracted man as early as the Mesolithic (the middle of the 8th up to the 5th millennium BC). In all likelihood, the first people arrived here from the south via the Velikaya River, but also from the basins of Lake Võrtsjärv and the Emajõgi River in the west. To the presence of fishermen and hunters in the area during that period refer archaeological finds. Bone implements (4 harpoon tips, an arrow head, a fish-hook) were discovered in the bottom of L. Peipsi at a distance of 1–10 km from the coast, *e.g.* close to Kuivloo Island, which does not exist there any more (Moora 1965, Jaanits 1983), and in the area of the Sahmen Bank. A bone harpoon was found from under the peat on the bank of the Omedu River. It is not excluded that the stone axe from Vtroya River in the northeastern coastal zone of the lake dates also from the Mesolithic.

Unfortunately, the data at our disposal do not show whether there were dwelling sites on L. Peipsi's coast during that period. As a matter of fact, remains of an old stone hearth were found in the sand under the cultural layer at the site of the Akali Settlement in the mouth of the Emajõgi River. These remains have yielded a radiocarbon age of 4300 yr BC (Punning *et al.* 1968), but the archaeologist Jaanits is convinced that this was only a temporary fire-place used by Mesolithic fishermen. Scraps of food, mainly bones of mammals, birds and fish, have preserved in the cultural layers of the 6th, 5th and 4th millennia BC. The remains of primitive fishing-nets (some 5000 yr BP) found from Siivertsi, a suburb of Narva Town, suggest that in the Mesolithic man already used fishing gear.

Mesolithic hunters and fishermen affected wild nature through hunting and forest fires. As in the Mesolithic, the number of inhabitants was small and they were concentrated in the vicinity of bodies of water. At that time, man's impact on the dynamics of therio- and ichthyofauna must have been negligible. Mesolithic man could have caused only local changes in the populations of some mammal, fish or bird species (Paaver 1965). The environment, as a whole, remained intact.

In the Neolithic (the 4th millennium – middle of the 2nd millennium BC), the number of inhabitants in the Peipsi area increased. Evidence is derived from the number of archaeological monuments associated with this area. The settlements were located mainly on the banks of the rivers flowing into L. Peipsi *s.s.* and L. Pihkva, and on the bank of the Narva River. Besides the already mentioned Akali, there was another settlement close to the Emajõgi River mouth. However, the available data (Jaanits 1959) suggest the existence of several more settlements in this area, which were later submerged by lake waters. According to Jaanits, the remains of Neolithic settlements have also been found on the bank of the Obdekh River flowing into L. Pihkva, and in the territory of the present-day Pskov Town. The finds of Neolithic stone chisels on the coast of L. Lämmijärv, not far from Mehikoorma Settlement, show that there was probably a settlement at that site. However, it was buried under the waters of the lake, which also reached several other settlements located immediately on the south coast of L. Peipsi or within the near-shore area of the present-day L. Pihkva. Occasional finds, first of all stone chisels, have also been found in the lower reaches of the Võhandu River, not far from the Town of Mustvee, and elsewhere.

In the Neolithic, a wider use was made of fishing-nets, and therefore the finds of fish-spears are much rarer than in the Mesolithic settlements. Compared with the Mesolithic, the fishing-nets were larger and better in construction. Fish-traps and fish-spears, which contributed to larger catches, were also used. The cultural layer of the Akali Settlement abounds in fish bones. The bones analysed belonged to sheat-fish or pike. The sheat-fish were often up to 1.5 m and the pike 40-60 cm in length (Jaanits 1959). Hunting was another important means of livelihood. Osteological evidence from ancient settlement sites shows that elk, wild boar, aurochs and beaver, to a lesser extent also bear, wild horse and roe, were the main game animals. Birds, particularly waterfowl, were also hunted, but their bones have not preserved in the cultural layer; only occasionally single wild duck and sea eagle bones have been found. Alongside fishing and hunting, man occupied himself with food-gathering. For instance, the site of the Akali Settlement abounds in crushed shells (Jaanits, 1959).

The transition from the Mesolithic to the Neolithic was accompanied by a slight increase in the population. Material culture developed, but man's activities did not exert any noticeable influence on the dynamics of the theriofauna or ichthyofauna; in the Neolithic, phytocoenoses remained much the same. Small tribes of fishers and hunters could only locally affect the plant cover and, with this, the living conditions of mammals (Paaver 1965). Nevertheless, human impact on wild nature in the Peipsi area undoubtedly increased, particularly in the immediate vicinity of the lower reaches and mouths of several rivers. But, as already said, this evidently did not cause any remarkable changes in the flora, fauna or other landscape components. Economic activities concentrated mostly in the area around settlements. More remote areas remained untouched, even by hunters.

Man's interference with the environment increased considerably at the end of the 3rd millennium BC, when the tribes of the boat-axe culture reached Estonia from the south. They were already familiar with animal husbandry and primitive tilling. Abundant animal bones excavated from burial places included also all kinds of domestic animals (except horses) raised in the Baltic region (Paaver 1965).

The tribes of the boat-axe culture roamed from place to place, and the traces of their settlements are rather weak. However, some burial places have still been found - one at Tartu and another at Haapsi on the west bank of L. Peipsi. Single boat-axes have been found in many locations in the western part of the Peipsi Basin, *e.g.* at Assikvere, Omedu, Avinurme, Tammetaga, in the surroundings of Kastre-Võnnu, on the lower reaches of the Võhandu River, *etc.* The inhabitants of the Akali and Kullamäe settlements continued to occupy themselves with fishing and hunting, but in the first half of the 2nd millennium BC cattle breeding and land cultivation evidently gained in importance as well. Stone axes, dating from that period, have been found in different places on the west coast of L. Peipsi, but also in the vicinity of Mustvee Town, on the lower reaches of the Võhandu River, and in several other localities. Pieces of ceramics dating from that period have been found at the sites of Akali and Kullamäe settlements, and on the bank of the Zhelcha River in the eastern part of the Peipsi area.

There is very little direct evidence to indicate that in the Late Neolithic the people in the Peipsi area occupied themselves with primitive tilling and animal husbandry. There are no data confirming that specially-shaped stones were used for crushing grains. A single sheep tooth from the Akali Settlement is not a weighty proof either. However, there are several indications (Jaanits 1959, *etc.*) of rudiments of primitive cattle-breeding and land cultivation in this region. But, nevertheless, fishing and hunting, were still the main means of livelihood, particularly in the surroundings of Neolithic settlements. From the 2nd millennium up to the present, human impact on the landscapes, particularly on the vegetation and zoocoenoses, has been been gradually increasing due to the development of progressive forms of economy. The cleared areas were used for animal husbandry and small plots of land also for primitive tilling. Branches of some broad-leaved trees, especially elm, were cut and used as fodder in winter. The decrease of forests caused some changes in the theriofauna. The disappearance of the aurochs, an inhabitant of forests and forest meadows on the coasts of water basins, and the wild hoarse, is one destructive effect of man's activities.

With time, the zone of human impact on wild nature extended gradually around the settlements, the inhabitants of which were mostly engaged in fishing, game hunting and food-gathering. This applies, first of all, to the densely populated area near the mouth of the Emajõgi River, where the zoocoenoses evidently changed in a great extent. The flora and fauna of the water basins impoverished considerably.

The Akali Settlement, which existed from the Early Neolithic (Late Mesolithic?) up to the Early Iron Age, offers interesting material to illustrate the evolution of L. Peipsi. The settlement was situated on a small sandy hill in the estuary of the Emajõgi River, on the south-west bank of the Akali River, which connects the Emajõgi and the Kalli. The cultural layer of the settlement is almost completely buried under peat. An up-to-2.5-m-thick peat layer (Jaanits 1959) covers the lower-lying parts of the cultural layer at Akali. At first, the settlement was situated on the riverside slope of the hill. However, a gradual rise in the groundwater level which was connected with the water-level rise in the southern part of L. Peipsi, caused accumulation of a thick layer of peat on the lower-lying part of the hill slope and made it unsuitable for habitation. The inhabitants could not stay there any longer and had to remove to the higher-lying parts of the slope. On the Kullamägi Hill, the situation was much the same (Jaanits 1959).

The movement of the settlement started in the second half of the 3rd millennium and continued slowly in the first half of the 2nd millenium BC; by that time the habitation centers in the lower-lying parts of the slope had already been abandoned. In the second half of the 2nd millennium, the movement somewhat intensified and the settlement center became located in the upper part of the hill. It remained there from the 1st millennium BC up to our era.

The above-presented data are in good agreement with the results of recent studies (1977–1978). According to these studies (Moora *et al.* 1988), the increment of peat in the surroundings of the settlement and the rise of water level in L. Peipsi takes place at a rate of 0.5–0.6 metres during the course of thousand of years. Some 7300 years ago, the mire surface was there 4 metres lower than to date. Some 6400 years ago, it was 0.5 metres higher and there was no mire around the settlement. During the following two millennia, the mire continued to advance slowly. About 3500 years ago, *i.e.* at the end of the Neolithic and in the beginning of the Bronze Age, the lowermost sites turned into swamp; the Akali Settlement was situated on a mineral "island" in the centre of the ever extending swamp. The settlement ceased to exist at the beginning of the Early Iron Age. This was due to profound changes in the social and economic development, *i.e.* primitive animal husbandry and land cultivation gained in importance, but natural conditions on a sandy hill in the middle of an advancing bog, did not support this kind of activity.

Archaeological finds of the Bronze Age (the middle of the 2nd millennium up to the 7th century BC) in the Peipsi area are scarce. The settlements continued to exist at Akali and on the Kullamägi Hill. Besides the finds from those sites, 50 bone axes were found from the vicinity of L. Peipsi during different works. The finds come from localities with highly variable natural conditions. These include the marshy lowlands at Alutaguse north of the lake, the environs of Värska on the sandy west coast of the lake, the undulating till plains of the South-East Estonian Plateau, which in the vicinity of Kodavere and Kallaste reaches immediately the west coast of L. Peipsi. In most cases, these finds are not associated with the settlements but, nevertheless, to a certain extent they reflect the distribution of settlements in the western part of the Peipsi region. The east coast was also inhabited, but the Bronze Age finds are much rarer there.

The transition from the Stone Age to the Early Iron Age was gradual: during a long period the food-gathering and food-producing economies existed side by side (Moora 1956, Jaanits *et al.* 1965). In general lines, man's influence on nature remained much the same as it had been during the Neolithic. The scanty data at our disposal suggest that during that period different ethnic groups intermingled in the Eastern Baltic, and animal husbandry and primitive tilling gained in importance. As a result, the population started to move to higher-lying fertile areas. And though hunting and fishing continued to play an important role in man's everyday life in those areas which supported that kind of activity (*e.g.* the mouth of the Emajõgi River), the cattle-raising tribes of the Early Iron Age had a greater impact on nature than the Neolithic tribes. In land cultivation, small plots of land, such as vegetable gardens, cleared by slash-and-burn methods, seem to have dominated. As earlier, the slash-and-burn methods were used to clear lands for grazing grounds. K. Kimmel analysed bottom deposits in the southern part of L. Lämmijärv on spore-and-pollen and discovered there pollen grains dating from the end of the Sub-Boreal. These serve as indirect evidence showing the expansion of tilling in the Bronze Age.

However, as earlier, the area under consideration was only partly inhabited. The population still preferred lower areas. The settlements were situated either in large river valleys or stood in the immediate vicinity to the sea. In general, both direct and indirect human influence on the landscapes was small in this region. In the areas adjacent to the settlements, the zone of human impact on wild nature expanded gradually. Forest lands decreased and animal grazing resulted in the meadows poor in plant species. Large carnivores, which posed a threat to domestic animals, were killed in large numbers. Species of southern origin were subject to indirect human impact (Paaver 1965). Man's interference with nature manifested itself clearly in soils of light mechanical composition, which were easy to cultivate by means of primitive implements.

The Early Iron Age (the 6th century BC – the 1st century AD) in the Peipsi area, like in most of central and southern Estonia, is characterised by a limited number of finds. In general lines, animal husbandry and primitive tilling had become the main occupation of inhabitants. Traces of man's economic activities at that time are still observable at Akali and on Kullamägi, but soon these sites were finally abandoned. In the first millennium BC, fishers and hunters left their old settlements. New settlements were not so closely related to water any more. During that period, a fortified settlement existed at Alatskivi on the west coast of L. Peipsi, and another probably in the territory of the present-day Pskov (Jaanits 1983). In search of better living and feeding conditions, most people moved away from the lake coast and settled in areas favourable for cultivation, *e.g.* Alatskivi. On the other hand, we do not know to which extent these scanty archaeological finds reflect the real habitation. But one is sure: the zone of human influence on wild nature had shifted to the lands with autochthonous and polyhydromorphic soils. The area under fields and forest meadows enlarged gradually and with this human impact on the plant cover and soils increased.

The first millennium AD, when land cultivation became the main occupation of inhabitants (Jaanits *et al.* 1982), marks the beginning of a qualitatively new epoch in the history of the relationship between man and nature. Evidence is derived from both archaeological monuments and the finds of domestic animal bones at the sites of the settlements, since animal husbandry was closely related with land cultivation systems (Paaver 1965). In the first millennium, the population increased, expanded over higher-lying areas and settled. Many of the present-day villages sprang up just during that period. The area of cultivable land increased abruptly and the zones of human influence expanded drastically. Man's ever growing interference with wild nature caused changes in the plant cover and in the Estonian landscape, as a whole. However, it seems that primitive implements did not allow cultivation of clayey soils until the second half of the first millennium.

During the last millennium, man essentially changed nature in the Peipsi area. As earlier, the forest area gradually decreased and the proportion of open landscapes, including meadows and forest meadows, increased as a result of the development of a spit. The composition of forests changed. Broad-leaved trees were felled and the area of secondary broad-leaved forests increased. New settlements sprang up and cultural biotops came into being, the hunting of several species of wild animals intensified (Paaver 1965).

In several regions, the forest land evidently reduced as early as the beginning of the first millennium when man moved and settled in localities most favourable for land cultivation. Already then, as a result of land cultivation, pine and mixed forests were destroyed at several sites on gravelly and sandy soils and partially in the areas with two-membered parent soils, *e.g.* at Alutaguse. The number of large forest mammals decreased, some species were destroyed. However, on the other hand, extensive forest areas were sparsely populated, *e.g.* Alutaguse, belts of kame fields on the boundary with the Saadjärv Drumlin Field, *etc.*, where most of mammals maintained their former abundances up to the middle of this millennium. The sites where trees had been cut down and which had become overgrown with broad-leaved species offered favourable conditions for some wild animal species whose abundance considerably increased (*e.g.* hoofed animals).

An increase in the population and the distribution of landscapes which had changed as a result of anthropogenic stress is revealed in the spread of archaeological monuments of the Roman Period of the Iron Age.

In the Roman period of the Iron Age (the 1st – 5th centuries AD), the Peipsi area was sparsely populated; single inhabited areas alternated with almost intact forests. The spread of stone burial places suggests an increase of the population on higher-lying sites, favourable for cultivation. This applies, first of all, to the areas south of the Emajõgi River where the number of archaeological monuments dating from the previous periods is very small (Jaanits 1983). Stone burial places have been found on the east slope of the South-East Estonian Plateau (Lääniste) and at the north foot of the Haanja Heights (Loosi). People had also settled on the lower reaches of the Võhandu River (*e.g.* the environs of the Räpina Settlement) and in the coastal zone of L. Lämmijärv in the vicinity of Mehikoorma. On the west coast of L. Peipsi, the inhabitants concentrated in the northeastern part of the South-East Estonian Plateau, first of all at Alatskivi and in its surroundings. Some burial places have preserved at Lahepera on L. Peipsi's west coast, at Alasoo and elsewhere. The practically unpopulated Alutaguse area was visited only by occasional hunters and fishermen.

The spread of fortified settlements, burial places and other archaeological monuments suggests that in the Middle Iron Age (the second half of the 5th century – the 9th century AD) the population of the Peipsi region continued to increase and colonized new areas (Jaanits 1983). Small fortified settlements were situated at Tarakvere south of Torma, and at Alatskivi, but also west of L. Lämmijärv – at Melliste, Lääniste, Kureküla, Kauksi and Võuküla. During this period, there were numerous settlements to the south-west and south from L. Pihkva (Lis'e, Gorodishche, Lezgi, Zakhnovo, *etc.*) (Jaanits 1983). The settlements at Kamno and Pskov on the lower reaches of the Velikaya River date from the same period. Archaeological finds from several fortified settlements (Kamno, Pskov) suggest that land cultivation and animal husbandry were the main branches of economy, but the inhabitants occupied themselves actively with fishing and hunting as well. The data showing the distribution of settlements in the eastern part of the Peipsi region is very thin, but Grodno on the bank of the Lochkino River is evidently one of the oldest settlements there. The settlements on the Tolba River in the vicinity of the Bolshoe Kryukovo Village and to the southwest from Gdov have not yet been dated (Jaanits 1983),

Alongside the Middle Iron Age fortified settlements, traces of burial mounds have also been found. In the Peipsi area, stone chist graves of that period are rather rare. They have been found at Lahepera and on the lower reaches of the Velikaya River. In L. Pihkva basin and on the banks of the rivers flowing into this lake, burial mounds – a new type of burial places in sands and sandy loam of different genesis – are most common. Early burial mounds occur also in the Peipsi Basin, but they are not so numerous there. In the eastern part of the Peipsi area, burial mounds are spread as far as Gdov, middle course of the Plyussa River and Lake Sampo, while west of L. Peipsi they reach the Kääpa River (Selirand & Tõnisson 1965, Jaanits 1983).

Judging from numerous archaeological monuments and occasional finds, in the Late Iron Age (the 10th–13th centuries AD) the distribution of the population and the spread of the landscapes affected by human activities in the Peipsi area changed only little. Like during the previous centuries, the densely populated areas with favourable agricultural conditions alternate with the areas with a very small number of permanent inhabitants, including the marshy estuary of the Emajõgi River, the sandy lowlands east of Värska (*e.g.* Palumaa, forest massifs and mires in the basin of the Mädajõgi River, *etc.*).

During that period, the Kodavere - Kallaste - Alatskivi area with its relatively fertile soils on undulating morainic plains was evidently one of the most densely populated areas. It was more or less permanently settled since the first half of the first millennium AD at the latest. Archaeological finds from the beginning of the 2nd millennium include stone chist graves and other grave types (Lahepera, Raatvere, Savastvere, Sassukvere) and ancient dwelling places at Ranna, Raatvere, Sääritsa, Kodavere, Tederküla (in the vicinity of Kallaste), Alasoo and elsewhere. Traces of iron melting have been found at Sepametsa, Punikvere and Raatvere. The number of single archaeological finds is high. Alatskivi, where two fortified settlements were situated close to one another, was evidently the center of this populated area.

In the 11th-12th centuries, inhabitants settled in several locations north of L. Peipsi, including the marginal glacial formations at Alutaguse, and on the banks of the Narva River. Burial mounds originating from that period are known from the upper and middle reaches of the Narva River (Olgin-Krest, Kuningaküla, *etc.*). There are more than 300 burial mounds of the latest type in the surroundings of Jõuga (Jaanits *et al.* 1982).

The northern part of the area to the south from the Emajõgi River was most densely populated, because soil con-ditions there were even better than in the southern part of the Peipsi area.

According to archaeological finds, permanent agri-cultural settlements were founded in the first millenium A. D. Such permanent centres of population were in the vicinity of Alatskivi and Gdov (on the coasts of L. Peipsi *s.s.*), and in the vicinity of Räpina (on the coasts of L. Pihkva). At the end of the 1st millenium Slavonic inhabitants reached L. Pihkva and the eastern coast of L. Peipsi *s.s.* The first town-type settlement in the Peipsi area arouse near the Pskov stronghold. In the 14th century, strongholds were built at Gdov and Vasknarva. Starting from the 15th–16th centuries, fishermen villages appeared on the coasts of the lake. During the course of time the number of Russian fishermen increased also in the western area of Lake Peipsi. In the second half of the 19th century and in the beginning of this century coastal villagers started cultivating vegetables and developed several branches of handicraft. A few villages outgrew into small county towns (Võõpsu in 1920-1938, Räpina in 1945) and later into towns (Mustvee and Kallaste in 1938).

To sum the above up, it may be said that at the beginning of the 13th century the spread of population and cultivated landscapes in the Peipsi region was much the same as to date. The west coasts of L. Peipsi and L. Lämmijärv were the most densely populated areas where, correspondingly, anthropogenic load on the plant cover, animals and other landscape components was the highest (the surroundings of Kodavere and Alatskivi). On the other hand, extensive areas, unfavourable for land cultivation, remained unchanged for a long time. In all likelihood, anthropogenic stress did not cause any serious ecological crises in that area. Neverthless, additional palynological and studies of other kind should be carried out before to draw a final conclusion.

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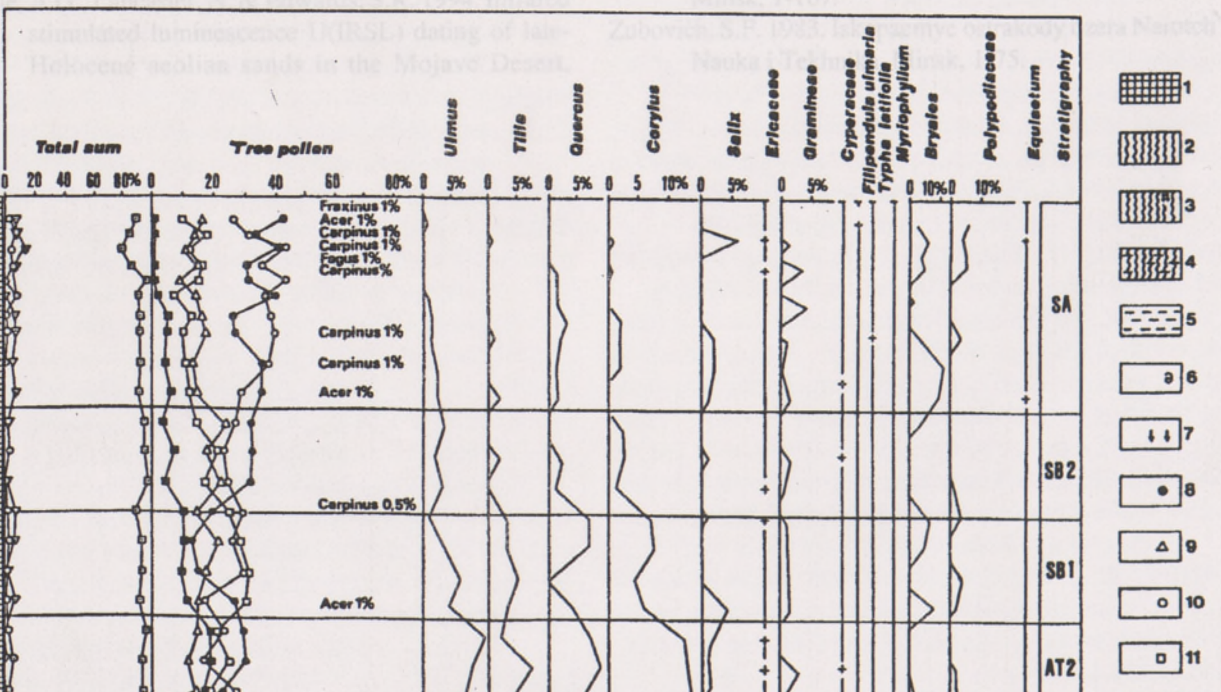
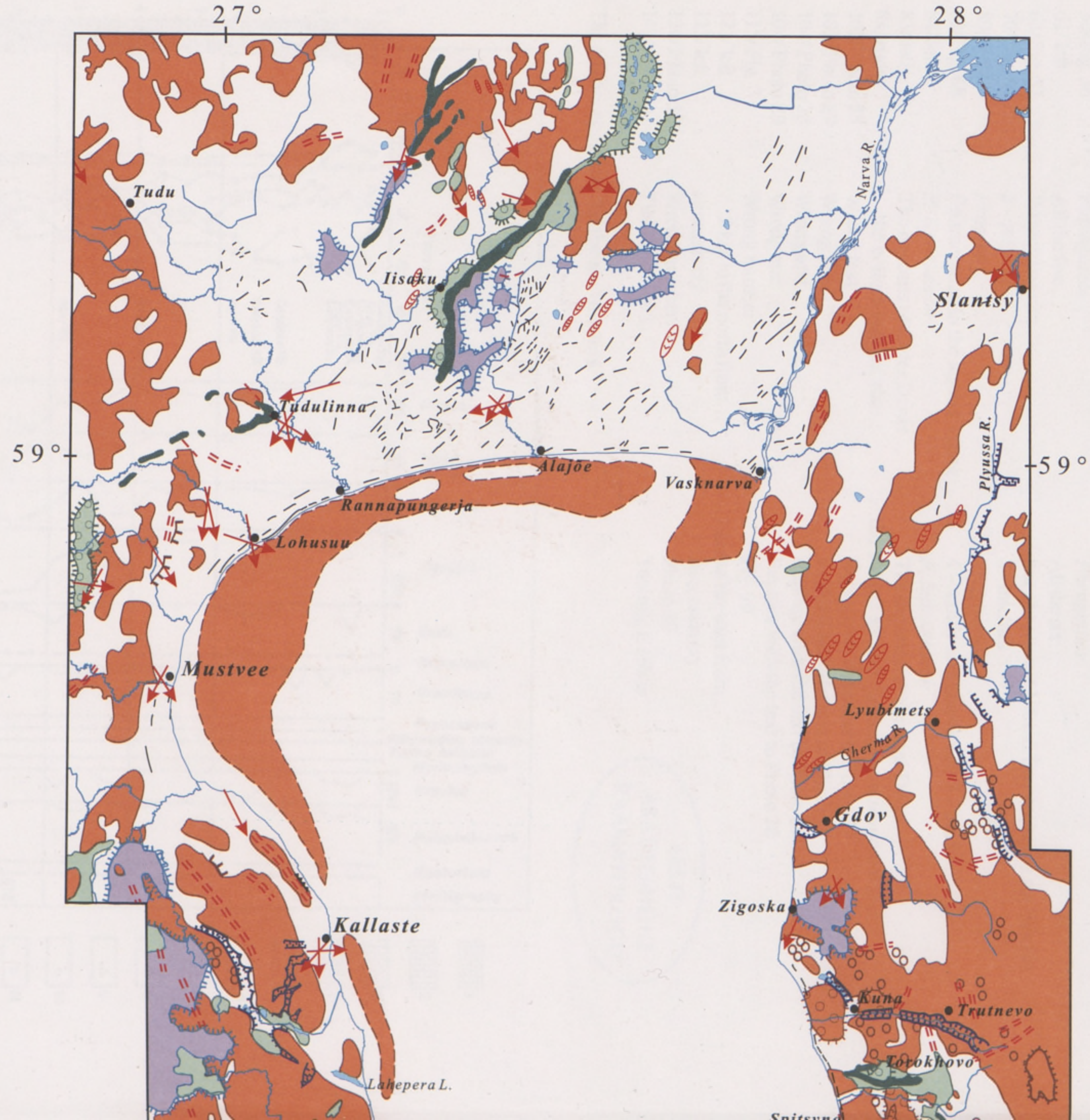
OÜ «Greif». Tellimus nr. 2366.

ERRATA

Page	Error	Correction
36/text to Fig.	Ilves & Sarv 1975	Sarv & Ilves 1975
41/ Table 5	Analyst V. Säga	Analyst L. Säga
44/ Fig. 29	Enumeration in the legend wrong	5 - *Carex-Hypnum* peat; 6 - sand
44/ Fig. 29	*Pharagmites*	*Phragmites*
45/ Fig. 30	*Pharagmites*	*Phragmites*
46/ Fig. 31	*Lycopedum*	*Lycopodium*
47 /Fig. 32	*Pharagmites*	*Phragmites*
50/ Fig. 36	Pharagmites	*Phragmites*
61/ left	All the five…	All the six…
63/ Fig. 52	Wrong diagram	For the correct Värska Bay sequence see this list
70/ Fig. 60	2 – miktites	2 - mictites
80/ right	*Fragila*	*Fragilaria*
81/ Fig. 68	Enumeration in the legend wrong	4 – sandy and clayey silt; 5 - sand
81/ left	*F. lepostauron*	*F. leptostauron*
83/ left	The deposits of Older Dryas…	The deposits of Younger Dryas…
84/ right	…the bottom species and…	…the bottom species *Gyrosigma attenuatum* and…
102/ Photo 25	Wrong text	Replace with the text to Photo 27
102/ Photo 27	Wrong text	Replace with the text to Photo 25
104/ Photo 28	Wrong text	Replace with the text to Photo 29
104/ Photo 29	Wrong text	Replace with the text to Photo 28
115/ Fig. 7	Wrong number	Fig. 99
121/ left	…2 km to the north from…	…2 km to the west from…
123/ left	sucessively	successively
124/ Photo 36	Wrong number	Photo 30
147/ right	Tuuling, I. 1998a	Tuuling, I. 1988a

The editors apologise for the above errors.




27°
28°
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Tudu
Iisaku
Tudulinna
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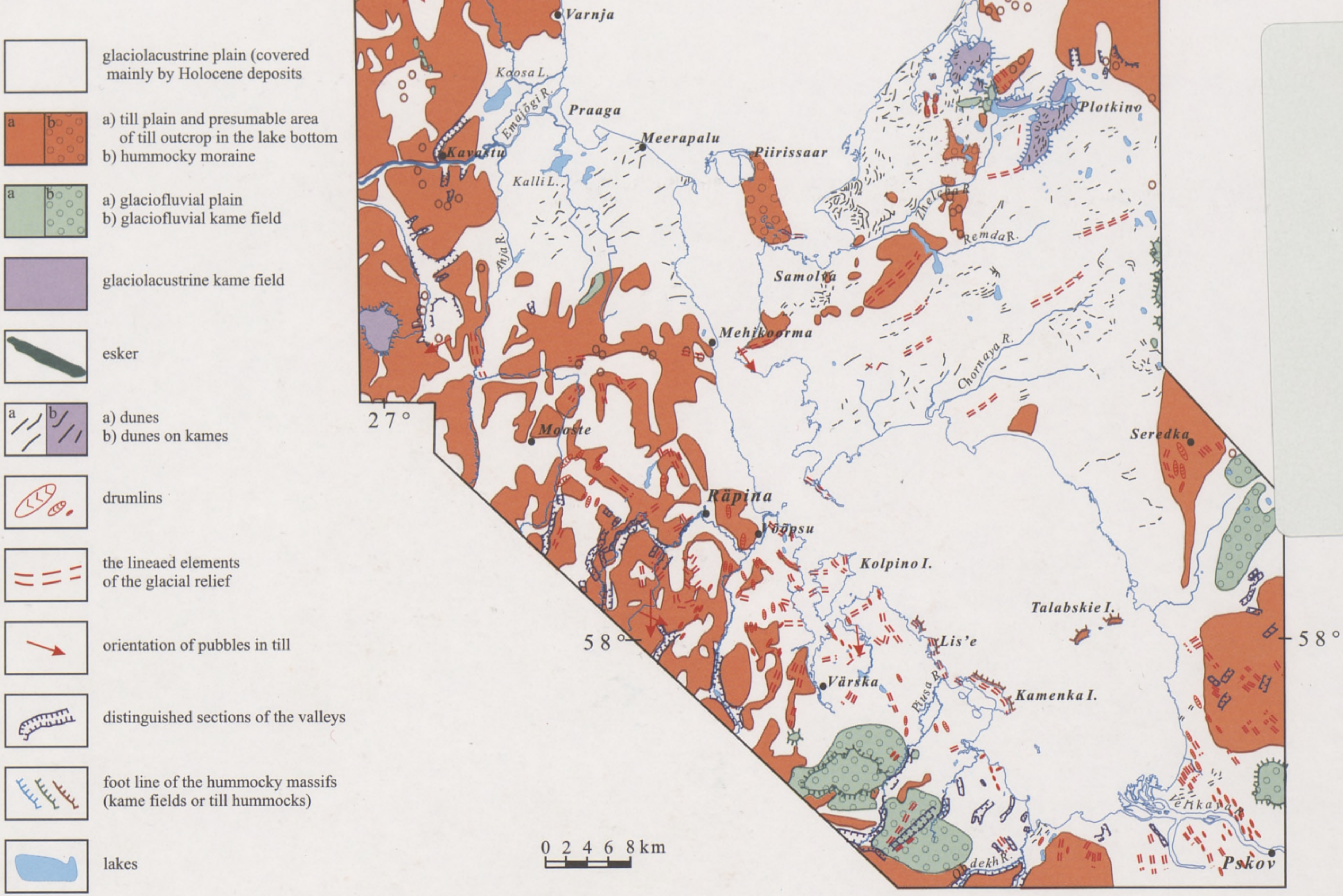
Vasknarva
Rannapungerja
Lohusuu
Mustvee
Kallaste
Lahepera L.
Narva R.
Slantsy
Plyussa R.
Lyubimets
Cherma R.
Gdov
Zigoska
Kuna
Trutnevo
Torokhovo



glaciolacustrine plain (covered mainly by Holocene deposits
a) till plain and presumable area of till outcrop in the lake bottom
b) hummocky moraine
a) glaciofluvial plain
b) glaciofluvial kame field
glaciolacustrine kame field
esker
a) dunes
b) dunes on kames
drumlins
the lineaed elements of the glacial relief
orientation of pubbles in till
distinguished sections of the valleys
foot line of the hummocky massifs (kame fields or till hummocks)
lakes
Varnja
Koosa L.
Emajõgi R.
Praaga
Meerapalu
Kavastu
Kalli L.
Piirissaar
Plotkino
Remda R.
Samolva
Mehikoorma
Chornaya R.
27°
Mooste
Seredka
Räpina
Kolpino I.
Talabskie I.
58°
Lis'e
Värska
Piusa R.
Kamenka I.
0 2 4 6 8 km
Pskov

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