Baltic Stratigraphical Association

THE FIFTH BALTIC STRATIGRAPHICAL CONFERENCE

BASIN STRATIGRAPHY – MODERN METHODS AND PROBLEMS

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I. Nestor

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Geological Survey of Lithuania
Institute of Geology and Geography
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The Fifth Baltic Stratigraphical Conference

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EXTENDED ABSTRACTS

Edited by Jonas Satkūnas and Jurga Lazauskienė

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The Conference is devoted to the all-kind problems and modern research methods of the **stratigraphy of the sedimentary basins**. The approach of quantitative stratigraphy as well as compilation and implementation of geological computer databases are highly emphasized.

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MILESTONES OF ACTIVITY OF THE BALTIC STRATIGRAPHIC ASSOCIATION

The Baltic countries have a long history and rich experience of cooperation in many fields of geology, first of all – stratigraphy, regional geology and geological mapping. This co-operation could be illustrated by many outstanding achievements, witnessing the value of joint efforts in the way of understanding of geology of the Baltic region and adjacent areas. In order to continue the cooperation in stratigraphic research, the Baltic Stratigraphic Association was established in 16th October 1990, replacing the former Regional Commissions on Stratigraphy of Estonia, Latvia, and Lithuania.

The activity of the Baltic Stratigraphic Association could be marked by the following milestones:

First Geological Conference of the Baltic Sea States, April 15–21, 1991, Tallinn – Lohusalu, Estonia. (**Abstracts** // Bulletin of the Geological Survey of Estonia. – 1991. – No. 1/1. – P. 1–54. – ISSN 1021-7428).

The Second Baltic Stratigraphic Conference, Vilnius, 9–14 May 1993. (**Abstracts** / Ed. by A. Grigelis, T.-R. Jankauskas, R. Mertinienė; Baltic Stratigraphic Association. – Vilnius: Geological Society of Lithuania, 1993. – 119 p.).

The Third Baltic Stratigraphical Conference, Tallinn, 8–12 October 1996. (**Abstracts, Field Guide** / Institute of Geology Tallinn, Institute of Geology University of Tartu, Geological Survey of Estonia, Baltic Stratigraphical Association. – Tartu, 1996. – 156 p. – ISBN 9985-60-225-2).

The Fourth Baltic Stratigraphical Conference "Problems and Methods of Modern Regional Stratigraphy". Jūrmala, Latvia, September 27–30, 1999. (**Abstracts** of a Joint Baltic Stratigraphical Association / IGCP 406 Project meeting / Ed. by E. Lukševičs, Ģ. Stinkulis, I. Kalnina; Institute of Geology, University of Latvia Faculty of Geographical and Earth Sciences, University of Latvia, State Geological Survey of Latvia, Latvian Museum of Natural History, Baltic Stratigraphical Association, UNESCO IUGS IGCP 406 Project. – Rīga, 1999. – 127 p. – ISBN 9984-661-06-7).

The Fifth Baltic Stratigraphical Conference "Basin Stratigraphy – Modern Methods and Problems" is being held in Vilnius and already attracted interest of nearly 100 scientists from 10 countries.

The Conference is devoted to mark the 200th birth anniversary of Ignotas Domeika (Ignacy Domeyko), the worldwide known geologist, researcher and public man.

A. Grigelis, T. Jankauskas, J. Lazauskienė, J. Satkūnas
On behalf of Organisers

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PALYNOLOGICAL CHARACTERICTIC OF THE BOUNDARY DEPOSITS OF UPPER SILURIAN AND LOWER DEVONIAN OF BELARUS

Abukhovskaya Veronika

Belarussian Scientific Prospecting Geological Research Institute, 220114, Minsk, Staroborisovsky tr., 14

The boundary deposits of Upper Silurian–Lower Devonian are distributed in the South Western Belarus, Brest Depression and Volyn monoclinal. The most complete section of them has been discovered in borehole Tomashovka 11 k.

The Upper part of Silurian is presented by Kustin horizon of Prgidol Stage. Various acritarhs dominant among plant microfossils. Few criptospores are presented. The miospore assemblages include species: Scabratisporites minor (Naum.) McGregor, Retusotriletes semizonalis McGregor, Synorisporites papilensis McGregor, S. verrucatus Rich. et Lister, Retusotriletes warringtonii Rich. et Lister, Tholisporites chulus (Cramer) McGregor var. nanus Rich. et Lister, T. chulus (Cramer) var. chulus Rich. et Lister. One specimen of Emphanisporites neglectus Vigran is present. It is possible that this part of the section can be correlated to zone chulus-vermiculatus (McGregor, Camfield, 1976).

Lower Devonian in borehole Tomashovka 11k is represented by Lokkov Stage. It was divided into Borschov and Chortkov horizons (Pushkin, Kruchek, 1978). Emphanisporites microornatus Rich. et Lister, Chelinospora cassicula Rich. et Lister, Emphanisporites minor Allen appear in the miospore assemblages of Borschov horizon (depth 432–487.2 m).

Conodonts of zone I. woschmidti were defined in Borschov horizon (depth 454–460 m) (Moiseeva, Kruchek, 1969).

Chortkov horizon (depth 406–432 m) has been characterized by more numerous and various miospore assemblages. Besides before appearing species Dictyotriletes paululus Tschibr., Retusotriletes scabratus Turnau, R. frivolis Tschibr., Apiculiretusispora plicata (Allen) Streel, Stenozonotriletes incessus Allen are present. The number and various of miospores of genera Amditisporites and Emphanisporites increase.

By miospores Borschov and Chortkov horizons can be correlated with the deposits of zone Emphanisporites microornatus-Streelispora newportensis, established in the deposits of Lower Gedinnian (Lohkov) (Richardson, McGregor, 1986).

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THE EXOSKELETON OF *THYESTES VERRUCOSUS* (OSTEOSTRACI, AGNATHA) FROM THE SILURIAN OF SAAREMAA ISLAND: A MODE OF OSSIFICATION

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By the present day, the microstructure of the exoskeleton of *Thyestes verrucosus* Eichwald, 1854 from the Silurian of Saaremaa Island (Estonia) has been studied relatively well among other osteostracans (Denison, 1951; Afanassieva, 1985, 1991). However, these data have not been used yet for explanation of the processes of ontogenetic development of dermal ossification in that group of agnathans.

The surface of the shield of *Thyestes verrucosus* is covered with numerous tubercles of different size: large tubercles with curved tips arranged in longitudinal rows, medium-sized tubercles with straight tips and small ribbed tubercles. A large number of pore fields are located on the surface of the shield and on the slopes of large and medium-sized tubercles. As a rule, no trace of polygonal pattern is observed. The superficial layer is present only in the apical part of the large and medium-sized tubercles. We studied the cephalothoracic shield of *Thyestes verrucosus*, in which, as supposed, the processes of dermal ossification have not been completed. The material comes from the Viita or the Vesiku Beds of the Roosiküla Regional Stage. The fragments of the exoskeleton, taken from the interzonal part of the dorsal shield, were investigated under the SEM. In posterolateral parts of the dorsal side of that specimen's shield there were found the radiating canals opened on the surface of the exoskeleton (Fig. 1). It has been determined that pore fields on the slopes of large tubercles are lined straightly along radiating canals (Fig. 2). Distal parts of these canals, open from above, form a pattern, typical for osteostracans, and determine approximate borders of "tesserae" of various sizes. It's assumed that the large tubercles of longitudinal rows (along the ribs of rigidity of the dorsal shield) emerged the first. The formation of hard covers began with the laying of dentine tips of the tubercles and proceeded centripetally. Middle-sized tubercles with thin tips were formed between them. Every tubercle was laid in the center of individual "tessera". And, finally, small tubercles emerged the last in ontogenesis, which is proved by their location on the slopes of larger tubercles. The exoskeleton of *Thyestes verrucosus* developed relatively rapidly but slower then the one of species of *Tremataspis* genus. The existence of a system of units (tesserae) gradually increasing in size allowed the individual to grow during a longer period of time up to complete consolidation of the shield and also distributed the burden on the organism related to the rapid process of shield formation.

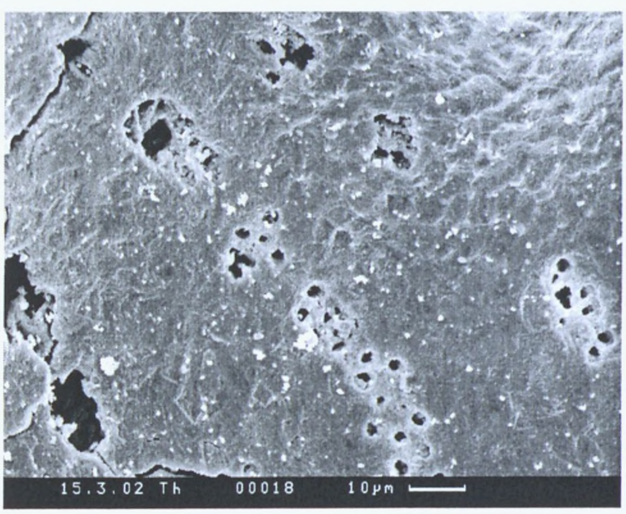

Fig. 1.

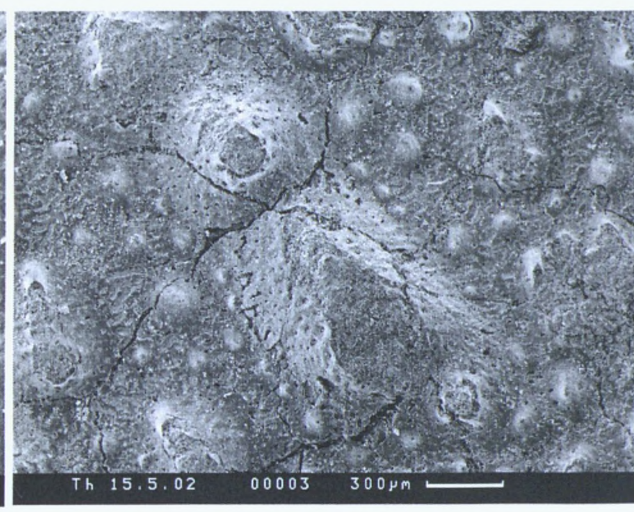

Fig. 2.

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THE "VOLKHOV COLLECTOR" IN WESTERN LATVIA – A SILICICLASTIC TURBIDITE BED IN THE ORDOVICIAN EPEIRIC CARBONATE BASIN

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The Baltoscandian Ordovician palaeobasin was a sedimentary starved epicontinental sea with extremely low sedimentation rate. In the Early and Middle Ordovician the Baltic area was slowly subsiding, and the sea was characterized by weak bathymetric and lithofacies differentiation described as ramp situation (Nestor and Einasto, 1997). Marginal areas of the basin (North Estonian and Lithuanian facies) were tectonically related to the southern slope of the Baltic Shield and northwestern slope of the Belarussian–Mazurian Anteclise, the central part of the basin (Latvia, western Lithuania) included the area of the Baltic Syneclise.

The siliciclastic sedimentation, i.e. formation of organic rich mud, quartz silt and sand prevailed in the basin during Early Ordovician. Cold- or temperate water carbonates, skeletal packstones and wackestones, covered the area in the Middle Ordovician. This change of sedimentation can be seen as relatively sharp transition from sandstone to limestone in the Billingen Stage in the area of upper/middle ramp (northern Estonia), and as a gradual transition from clay to argillaceous limestone in the Billingen and Volkhov stages in the lower ramp (western Latvia).

Due to the relatively high sea level and tectonically stable situation during the Ordovician, the erosion rate in surrounding land areas (Baltic Shield) was low, resulting in a very low siliciclastic input into the basin. The Middle and Upper Ordovician carbonate sequence of Baltoscandia is particularly poor in the sand-sized or coarser siliciclastic material. Therefore, the rare quartz sand accumulations in the carbonate section formed during more than 50 Ma are conspicuous markers of unusual sedimentation events. These events might be initiated, for example, by rapid eustatic sea level changes or by meteorite impact explosions. There are four distinctive stratigraphic levels with quartz sand enrichment in the Middle and Upper Ordovician carbonate sequence of the East Baltic area. One of these is Arenig (Volkhov–Kunda) interval with sandy limestone of the Pakri Formation (Kunda Stage) in northwestern Estonia and a conspicuous sandstone-siltstone layer in western Latvia known as the Volkhov Collector

The preliminary results on the study of the Volkhov Collector in Latvia are based on data from two core sections. Vergale-50 drillcore from western Latvia (Figure) was sampled in the interval of the Kriukai Formation. Samples from the bed of the collector sandstone/siltstone and surrounding limestone were dissolved in dilute hydrochloric acid and grain-size analysis of the insoluble residue was performed. Thin sections from the collector sandstone bed of Vergale-49 and Vergale-50 drillcores were analysed. Ostracodes were separated and studied from the 19 samples of the Kriukai Formation from Vergale-50 core to establish the stratigraphical position of the Volkhov Collector.

The Volkhov Collector in the Vergale-50 section comprise a layer of calcareous sandstone and siltstone with thickness of 25 cm (Figure). It occurs in a grey limestone bed within mainly red limestone succession. The collector layer can be clearly subdivided into four beds, which resemble well the Bouma divisions of the ideal turbidite succession (Bouma, 1962):

Bed A – massive sandstone, lowermost 15 cm (5 cm preserved in the core);

Bed B – planar laminated fine sandstone, 2–3 cm;

Bed C – cross-laminated fine sandstone with small climbing ripples, 1 cm;

Bed D – laminated siltstone, 5–6 cm.

According to the thin section analysis, the bed A is composed mainly of quartz grains, 0.1–0.3 mm in size, rounded or slightly angular. Other main components are skeletal fragments (mainly trilobites), feldspar and glauconite grains. Micritic material is missing, the sandstone is cemented mainly by calcite. The insoluble residue (mainly sand and silt) content decreases in collector bed upward from 77% in bed A to 58% in bed D. Surrounding limestone contains 21–38% of insoluble residue, mainly clay and fine silt.

The data on composition of the Volkhov Collector in the studied sections agree well with the data by Yakovleva (1977). According to her description, the Volkhov Collector is distributed in the westernmost

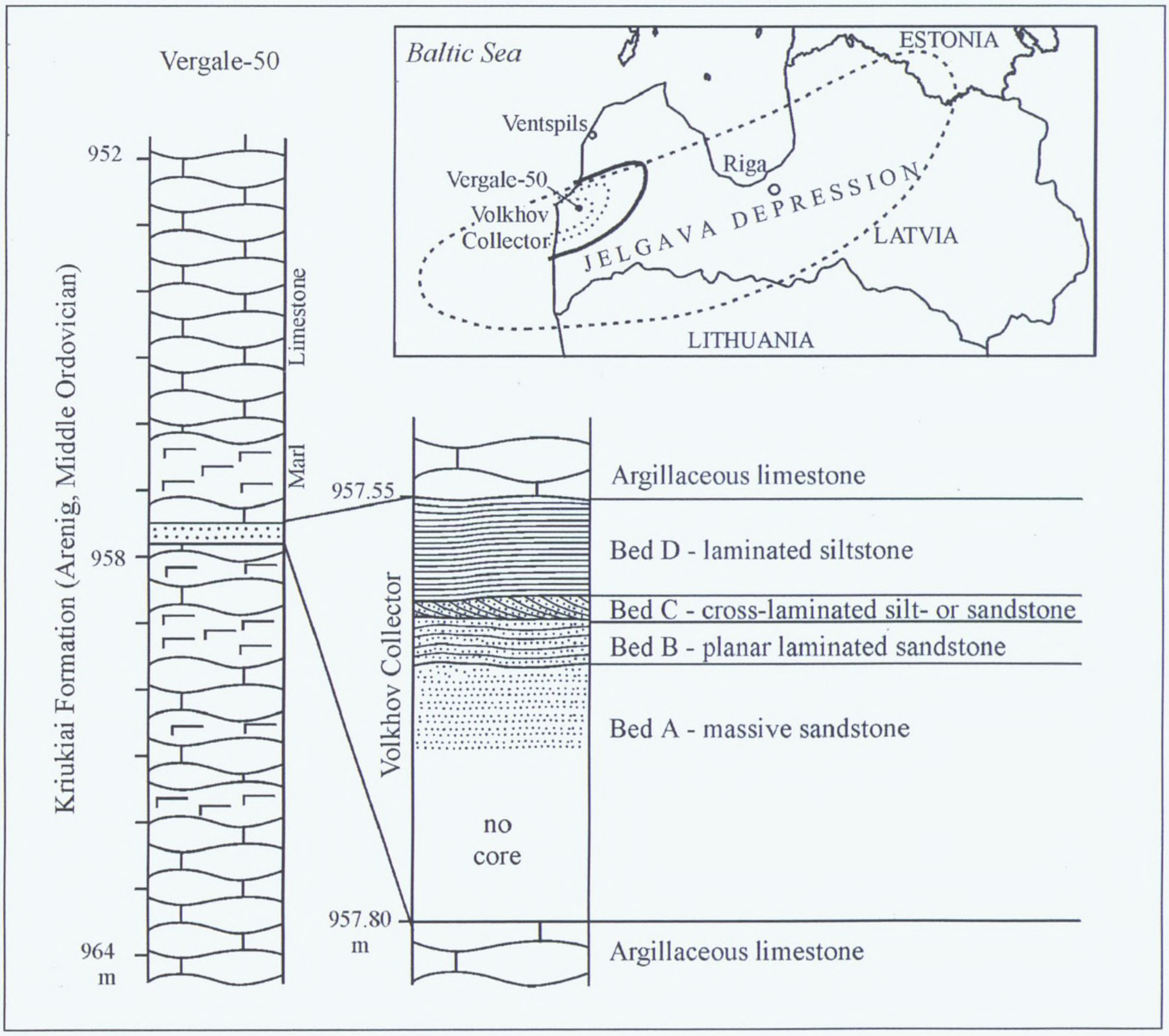

Figure. The Volkhov Collector in the Vergale-50 core section. The lithological subdivisions of the collector layer (A–D) resemble well the Bouma divisions of a turbidite succession (Bouma, 1962). The Volkhov Collector layer is distributed in northwestern flank of the Jelgava Depression in western Latvia and three lithofacies.

Latvia as a belt of 100 km long and 50 km wide, and it probably continues westward to the Baltic Sea area (Laškov and Yakovleva, 1977). The collector bed has the thickness of 0.1–0.8 m, in one core (Pavilosta-51) even 1.5 m. It is thicker and more sandy in the western part and in the centre of its distribution area, getting thinner and more silty towards the margins of the area (Yakovleva, 1977). Margulis and Shedrina (1996) mentioned washout signs at the top and bottom of the Volkhov Collector bed in the cores of the Kuldiga area. The sandstone/siltstone bed is, as a rule, oil-saturated.

The stratigraphical position of the Volkhov Collector, upper part of the Kriukai Formation, refers to the Volkhov age of the bed (Yakovleva, 1977; Ulst et al., 1982). Ostracodes, yielded from the limestone under- and overlying the collector bed in the Vergale-50 section – *Conchoprimitia gammae, Protallinnella grewingkii, Aulacopsis simplex* and *Unisulcopleura punctosulcata* are found abundantly in the Upper Volkhov and Kunda stages in Latvian and in southern Estonian sections. According to the preliminary estimates, the Volkhov Collector may be of late Volkhov or Kunda age.

The data on the textures and composition allow to conclude that the Volkhov Collector is a siliciclastic sandy/silty turbidite bed deposited on the carbonate lower ramp. The bed was formed on the palaeoslope of the Jelgava Depression, probably by a single turbidite flow. The sand and silt material was derived from the upper ramp, from western Estonia and/or adjacent areas of the Baltic Sea. There is an area between the Swedish island Gotska Sandön and the Hiiumaa Island in NW Estonia where the Volkhov Stage is missing and the Kunda Stage (middle substage) is in contact with the Cambrian or Lower Ordovician sandstones (Thorslund 1958; Meidla, 1997). If our interpretation is correct, the bathymetry of the Middle Ordovician carbonate ramp was not so weakly differentiated as it is proposed (Nestor and Einasto, 1997). It means that slopes on the flanks of the Jelgava depression were distinct already in Arenig time.

Another question is why there was only one turbidite flow episode in the inner area of the Baltoscandian epicontinental sea during the whole Ordovician. The only other turbidite-like deposit known from the area is related to the infilling of the Upper Ordovician Kärdla impact crater in NW Estonia (Puura and Suuroja, 1992). The siliciclastic turbidites might not form later because the source area of sand was covered by carbonates, and carbonate turbidites may be difficult to recognize in the packstone/wackestone sequence. If the turbidite event in Arenig was unique for the basin, it should have had a specific reason. The turbidite bed may be correlated with the Täljsten unit (lower Kunda Substage) in Sweden and southern Estonia. This grey limestone (grainstone) bed in generally red-coloured wackestones/packestones sequence is considered to be formed during a sea level fall episode (Dronov et al., 2001). Erosion and resedimentation of sandy material on the upper ramp near the Volkhov/Kunda boundary during the sea level drop might be responsible for the turbidite flow and formation of sandy limestone deposits of the Pakri Formation (Kunda Stage) in NW Estonia. There is evidence of an earthquake or explosion in the Osmussaar Island and in surrounding areas of NW Estonia during the Kunda age, expressed as strong brecciation of limestones (Puura and Tuuling 1988). Whether the turbidite bed in western Latvia is related to these processes should be shown by detailed stratigraphical correlation. Finally, the meteorite impact explosion as initiator of the turbidite flow can not be excluded, although no impact features are known from the turbidite sandstone so far.

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HISTORY OF EARLY PALAEOZOIC PASSIVE MARGIN OF NORTHEASTERN BALTICA

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The Lower Palaeozoic sequences (more than 4500 m thickness) exposed along the western slope of the Polar, Subpolar, Northern Urals and Pre-Urals Foredeep belongs to the marginal succession of the northeastern Baltica palaeocontinent (= Timan–Northern Ural region = Pechora Plate). The formation of a passive margin on the northeastern Baltica paleocontinent following the opening of the Ural paleoocean in the Late Cambrian(?) – Early Ordovician was the major tectonic event which defined the morphology and facies plan of the sedimentary basin (Puchkov, 2000). The rather well preservation in the West Urals sections the Lower Palaeozoic various reefs and organic buildups provides a relatively rare oportunity to understand the dynamic processes, style, and bathymetry associated with reef growth in an extensive (about 1000 km long) passive platform margin. Distribution and geometry of reefs were provided by regional tectonic event of the Pechora Plate causing a ramp-to-shelf and shelf-to-basin transition for the first time. Scale of the buildups depended on the magnitude and duration of sea level fluctuations that were connected with differences in the movements of isolated blocks of the Pechora's plate basement margin. Laterally distribution of the slope buildups various in age and displied wide reef facies reflect a progradation of platform margin (Antoshkina, 1999). The platform margin progided slowly eastward throughout the Lower Palaeozoic until the passive continental margin having changed to a compressive margin during the Uralian collision orogeny.

Lowermost Palaeozoic rock units were deposited during a period of predominant siliciclastic sedimentation in the Early Ordovician. Formation of coarse-grained volcanogenic- terrigenous deposits in the Early Ordovician was replaced by alluvial-deltaic and marine carbonate sediments in the Middle Ordovician.

The Llanvirn–Lower Emsian rock units featured submarine relief more than 2.5 km thick (Fig. 1). The major factors controlling formation and accumulation of carbonate deposits are tectonic situation, relative sea level fluctuation, climatic stability and biotic balance. The change of these factors in space and time was determinative for features of carbonate sedimentation and distribution of fauna. Carbonate deposits including various organic buildups originated on a shallow-marine shelf, at the shelf margin, and in relatively deep waters of a slope and ramp environments.

In the first half of the Middle Ordovician carbonate sedimentation was quite limited, being located in an area of a pericratonic deep. In the process of progressive subsidence of the platform margin a marine basin increased at the expense of Early–Middle Ordovician land. Marginal shoals were drowned and rather deep-water conditions were formed here. It caused sedimentation of clayish-carbonate muds containing a few bioclastic material. Since the middle Caradocian the marine conditions have become consecutively shallower (Subpolar..., 2000). Beginning from this time carbonate sedimentation occurred on the passive margin where shelf and continental slope zones are distinguished. Source areas were located on the west of the considered region where the Timan Ridge is situated now. The same environments extended to the Early Silurian.

In the late Ashgill the former ramp as a result of growth shoals belt of marginal uplift transformed in the epicontinental platform. A global eustatic fall of sea level and generation of barrier at the shelf margin resulted in formation of an extensive intrashelf lagoon. In the pericratonic area accumulated carbonate-sulfate thick. At the edge of an incipient late Ashgill evaporitic shelf margin grew the first Palaeozoic reefs (up to 400 m thick).

The development of late Ashgillian reefs (late Malaya Tavrota time) was strongly disturbed by syndepositional erosion. A sea level drop during the middle part of the late Ashgill is reflected in a various degree of erosion of the upper part of the reefs. The open-marine shelf sediments overlying the reef top indicate an abrupt sea level rise at the end of the late Ashgill, causing the termination of reef growth. The large sea level fall is marked at the top of Ashgillian sequences. Near the end of the Ordovician the carbonate platform was exposed during the glacio-eustatic sea level lowstand. Accordingly, it is most probable that the uppermost Ashgill (Hirnantian) and lower Ruddanian are missing in the westuralian successions. Unfortunately, poor palaeontological data from these strata do not allow identification of the exact position of Ordovician–Silurian boundary and duration of the gap suggested by the distribution of fauna.

The most significant of sea level changes are dated for the late Aeronian and lower Telychian. The gradational shallowing facies from deep shelf (late Ruddanian–early Aeronian) to intertidal (late Aeronian–earliest Telychian?) setting reflect gradual falling of sea level. During this lowstand phase (Fillip"el' time) patch-reefs (up to 60 m thick) fringed of the inner margin-shelf lagoons with high salinity. An isolated knoll-reef (up to 160 m thick) at the outer shelf margin is recognized. The Llandoverian glacio-eustatic sea level highstand dated for the Lower–Middle Telychian (Marshrytny time). The transgression developed rather rapid and was exhibited a sharp change of intertidal and supratidal limestones and dolostones predominate in most of sequences (or reefal rock units) by in fauna-reach deep-subtidal and deep-shelf limestones in the shelf margin.

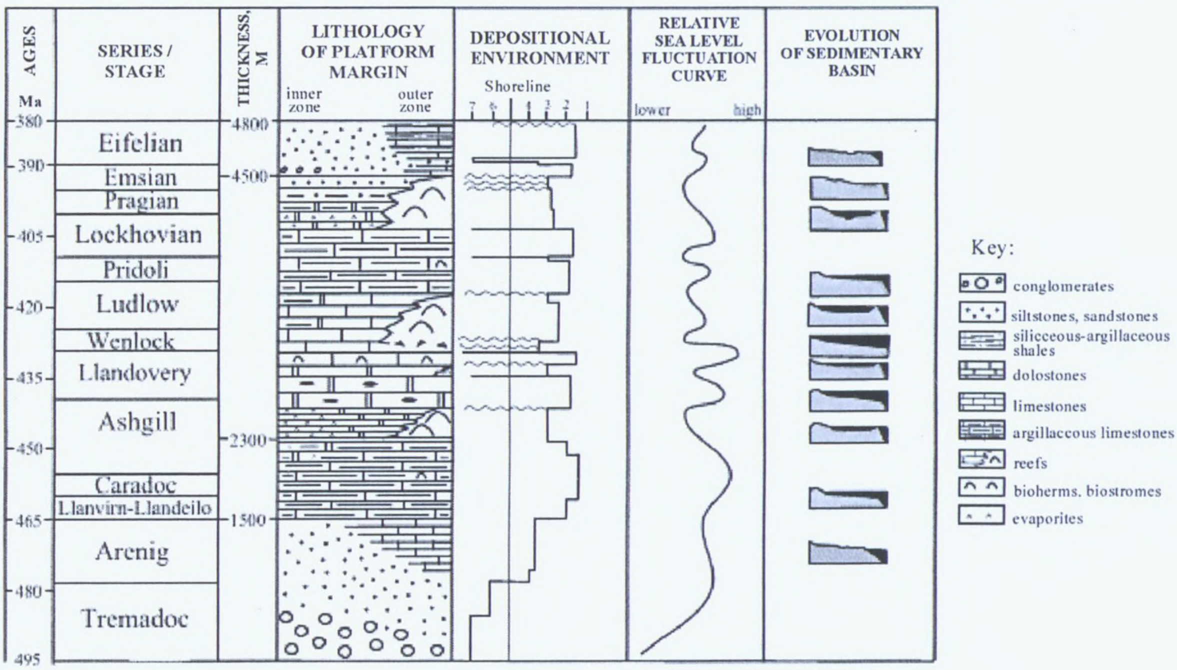
The rapid sea level fall near of the end of the Llandovery (early Ust' Durnayu time) was accompanied by a spread of lagoonal-intertidal facies. It is possible that a formation of basal reefal breccias of the Wenlock–Ludlow reef complex in the Subpolar Urals is correlated with this lowstand phase. They are represented by submarine canyon current deposits and reflect collapse of the Telychian platform margin. The Wenlock time was characterized by shallow-water environments and low diverse fauna. Shelf deposits were often eroded therefore lime gravelstones, silt-and sandstones, and silt-sand siliciclastic material in laminated shelf margin rock units were common. In Wenlockian reefal deposits a lithoclastic material was abundant. The Wenlock–Ludlow reefs (up to 700 m thick) was built at the edge of an incipient shelf margin after a drowned platform during Telychian time. The Ludlow reefs rimmed a carbonate shelf that accumulated back-reef lagoon deposits containing a rather monotonous fauna which reflect shallow-water, basically low turbulent, and locally salinity depositional conditions. Tidal flat deposits widespread westward in the inner margin shelf (Tshernyshev and Tchernov ridges). Reef growth was eventually terminated because an abrupt sea level fall at the early-late Ludfordian boundary was replaced by a relatively abrupt sea level rise.

In the Late Silurian–Early Devonian there began differentiation between stable and tectonically active structural zones along large regional faults into the Pechora Plate. It was accompanied by a series of grabens forming the Pechora–Kolva riftogenic zone (Malyshev, 2000). The beginning of the Ludlow characterize a stage of the Late Silurian–Lower Devonian degradation of sea sedimentation which included the transgression phases in the middle part of the Pridolian and early Lochkovian times. The significant change in the development of facies and fauna caused a sharp global sea level rise in the late Ludfordian. Terrigenous-carbonate successions formed skeletal detritus-rich sediments with various benthic fauna including small biostromes. They were strongly controlled by rather uniformly subsiding continental margin and following widespread transgression. These subtidal facies formed a laterally continuous tracts from shelf to relatively deep-water environments up to the middle Lochkovian time.

Since the late Lochkovian the sea basin has become extremely shallow. It is caused by a sharp fall of sea level and growth reef barriers at the shelf margin. The strong shallowing on a background of the aridity of climate has resulted in change of carbonate in the pericratonic area to galogenous-sulfate sedimentation. The late Lochkovian (Sotchemkyrta time) reefs (up to 170 m thick) originated at the new shaped evaporitic shelf margin after a long--term drowned platform margin during the late Ludfordian–Middle Lochkovian age. The largest barrier reefs (up to 1200 m thick) at the Pragian shelf margin being constructed by the most various reef assemblages covered the stromatolite-rich late Lochkovian reefs. The Pragian reefs growth was interrupted by an abrupt see level fall in the end of the Pragian. In the middle of the Emsian the Lower Devonian reefs formation was terminated finally. The Pragian back-reef lacustrine-marsh plain sediments were replaced by alluvial plain sediments in the early Emsian. The reefs originated in the platform margin before an abrupt differentiation of the Pechora Plate in result of the Pechora–Kolva palaeorift. The Lower Devonian reefal carbonates overlay by a, mostly terrigenous, younger Lower Devonian–Middle Devonian sequences. Pre-Middle Devonian time was preceded by a regional hiatus, brought about by a large uplift of the greatest part of the Pechora Plate.

So, in history of the northeastern Baltica palaeocontinent during the Lower Palaeozoic three large phases are distinguished. Each of them is characterized by peculiarities in the development of the basins and features of accumulated strata. The first phase represents the Upper Cambrian(?)–Lower Ordovician thick polygenic sequence with the volcanogenic-terrigenous accumulation of rifting troughs at the passive continental margin. The second phase characterises the Middle Ordovician–Lower Asgillian carbonate deposits as a typical ramp sedimentary series of a wedge-like shape body. It composed polygenic deposits of the ramp various-ment can correlate with a phase of the spreading Uralian Palaeoocean. The third late Ashgill–Early Emsian phase has began as a result of transformation of the carbonate ramp in the carbonate epicontinental platform in the late Ashgill. The succession of this phase reflected the relationships of ramps and platforms became complicated, but also the spectrum of conditions of their formation extends. It is possible, that a complex depositional process was connected with forming of the subduction system around the northern and eastern margins of Baltica (in mordern co-ordinates) and the collision of Baltica with Laurentia. The close relation between changes of global ocean level, regional and global tectonic events and formation of a sedimentary series is outlined. The last Lower Devonian strata corresponding to the late Emsian carbonate-terrigenous sequence represent the lowstand ramp sedimentary body.

The only direct relations between the general patterns of sedimentation and the change of sea-levels are that the transgressions and regressions were more frequent during times when carbonate production was low, and that these times, the transgressions did not extend so far into epicontinental shelfes as when carbonate production was higher. Combination of eustasy and tectonic subsidence produced a relative change of sea-level. The relative change of sea level creates the available space for accommodation of the sediments. The thickness of sediments is primarily controlled by tectonic subsidence. The depositional stratal patterns and distribution of facies are caused by the rate of relative change of sea-level. This expresses itself by the change of the relative sea level curve, which is mainly caused by eustasy (Figure).



<table>
<tr><td>Sediment</td><td colspan="7">Depositional environment</td></tr>
<tr><td>Clastic</td><td colspan="2"></td><td colspan="3">Deltaic</td><td>Lacustrine</td><td>Alluvial plain</td></tr>
<tr><td rowspan="3">Carbonate</td><td colspan="2"></td><td colspan="2">Lagoon</td><td colspan="3"></td></tr>
<tr><td>Slope</td><td>Shelf</td><td>Subtidal</td><td>Tidal flat</td><td>Supratidal</td><td>Lacustrine</td><td></td></tr>
<tr><td colspan="2">Ramp</td><td colspan="2"></td><td colspan="3"></td></tr>
<tr><td></td><td>1</td><td>2</td><td>3</td><td>4</td><td>5</td><td>6</td><td>7</td></tr>
</table>

Shoreline

Figure. Correlation of lithology with environmental and sea level changes reflecting evolution of the Lower Palaeozoic passive margin of the northeastern East European craton.

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SOME COMMENTS ON THE SILURIAN STRATIGRAPHY OF THE WESTERN URALS: STEPS TOWARDS GLOBAL CORRELATIONS

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The Silurian strata are exposed in a folded-thrusted belt stretching more than 2000 km from south to north along the western slope of the Urals. Rocks in this belt are considerably deformed and their real relations in the sections are sometimes complicate to find out. The thickness of the Silurian strata in the Western Urals varies greatly from region to region. In the Mikhajlovsk–Bel'sk structural-facial zone (dominated by various carbonate facies of shelf origin) the thickness of Silurian deposits changes from 430–630 m in the Southern and Middle Urals up to 830–1600 m in the Subpolar Urals. In the Sakmara–Lemva structural-facial zone (mainly black graptolitic and argillaceous cherty shales of deep-basin origin) the thickness of strata of 395–485 m in the southern Central Urals decreases to185–310 m in the Polar Urals.

The most recent officially accepted ideas about the Silurian stratigraphy in the Western Urals were published about ten years ago (Antsygin et al., 1993). Later studies revealed that there still exist many problems connected with the datings of regional stratigraphic units and that the scheme needs some revision (Antoshkina, 2000; Beznosova, 2000). It is evident that at least some of the levels identified in the Western Urals' sequence as the boundaries of the Silurian series in reality do not correspond to these levels in the International Standard.

The results of modern studies on event, sequence and isotopic stratigraphy in many regions all over the world have convincingly demostrated the existence of markers of global eustatic events. Recognition of these events in a particular section allows us to correlate it more precisely, and with higher reliability, with the sections in other regions. Restudy of Silurian stratigraphy in the northern part of the Western Urals, in the Timan–Northern Ural region, is in progress and some revised datings of regional stratigraphic units have been proposed (Antoshkina and Beznosova, 1999; Antoshkina et al., 2000; Männik et al., 2000; Melnikov and Zhemchugova, 2000).

In Palaeozoic, the Western Ural region formed the eastern margin of the Baltica Palaeocontinent. During the Silurian, an extensive carbonate platform formed the northeastern continental shelf of Baltica. The sedimentation in this region took place in highly variable (lagoon, reef and backreef, restricted and open shelf) conditions. The sediments formed are, as a rule, highly fossiliferous. Brachiopods, tabulate corals, stromatoporoids and ostracodes are the most common fossils. The best sections exposing strata from the Middle Orodvician up to the Upper Devonian are located in the Kozhym River region, Subpolar Urals. Recent studies have revealed that considerable gaps, often connected with the series boundaries, occur at several levels in the Silurian sequence in the Subpolar Urals.

The position of the Ordovician–Silurian boundary, proposed in the stratigraphic scheme of the Western Urals (Antsygin et al., 1993), lies in an interval of lithologically homogeneous secondary dolostones. In the majority of the studied sections, between the strata with Ordovician fauna below and Silurian fauna above lies an interval without any identifiable fossil, so-called "barren unit" (Beznosova, 2000). The lowermost Silurian fauna appearing in the sequences of the Timan–Northern Ural region above this interval includes brachiopods *Virgiana* sp. and *Borealis*? sp., and conodonts *Ozarkodina oldhamensis*, *Distomodus* cf. *kentukyensis*, *Oulodus*? *panuarensis* and *Aspelundia expansa*. In the other parts of the world, *O.*? *panuarensis* and *A. expansa* are known to appear in the uppermost Rhuddanian–lowermost Aeronian. Accordingly, it is most probable that the strata corresponding to the upper Ashgill (Hirnantian) and lower Rhuddanian are missing in the Kozhym River sequence.

The precise position of the Aeronian–Telychian boundary in the sequence is not yet known but, most probably, it lies in the strata corresponding to the Filipp"el' Regional Stage. This stage is, as a rule, represented by sediments formed in very shallow-water conditions during a major sea level fall. Stromatolites and interbeds of lagoonal dolostones are common. It is possible that in many regions (including Baltic) these strata correspond to a gap (Antoshkina et al., 2000).

The Llandovery–Wenlock boundary, as defined in its type section at Leasows (Welsh Borderland) lies in the middle part of the Ust' Durnayu Regional Stage in the Subpolar Urals (Männik and Martma, 2000; Männik et al., 2000; Melnikov and Zhemchugova, 2000). The $\delta^{13}C$ studies indicate that a considerable gap exists at this boundary (Männik and Martma, 2000). To the highly unstable sedimentation during the Ust'Durnayu time indicate also the occurrence of frequent interbeds of calcareous gravelstones, siltstones and sandstones, erosional surfaces, and rapid changes between facies types in relatively narrow interval in the sequence. The interbeds rich in silt and fine sand can be often followed over vast areas.

Siltstones and sandstones together with dolomitized stromatolitic limestones are most characteristic of the uppermost Wenlock strata. These facts show that quite often in Wenlock a syndepositional erosion occurred in the region. In other areas (in the Middle Urals) similar lithological features have been recorded also in the lowermost Ludlow strata (Antsygin et al., 1993).

Recently, it also appeared that the Ludlow–Přidoli boundary in the Timan–Northern Ural region lies in the Belush'ya Regional Stage but not below it as was considered earlier (Antsygin et al., 1993). The boundary, most probably, should be looked for in an interval above the level of disappearance of *Howellella pseudogibbosa* (Beznosova, 2000), in the lower part of the *Collarathyris trapezoideus* brachiopod Biozone sensu Modzalevskaya (Modzalevskaya and Wenzel, 1999).

The Silurian–Devonian boundary is characterized by sharp changes in facies and in faunal assemblages evidently connected with a rapid fall in sea level (Antoshkina, 2000). From the strata correlated with the uppermost Přidoli rare specimens of brachiopods *Collarothyris canaliculata* and *Grebenella parvula*, and ostracodes *Signetopsis arborea*, *Tollitia nota*, etc. have been identified. From the overlying mudstones of the lowermost Ovinparma Regional Stage (= Lochkov) brachiopods *Howellella angustiplicata* and *Protathyris praecursor*, ostracodes *Diszygopleura* sp. and *Knoxiella*? sp, and conodont *Icriodus woschmidti* were found.

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GEOLOGICAL POSITION OF THE AUGUSTOVIAN INTERGLACIAL DEPOSITS

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The Augustovian Interglacial sediments in stratotype profile at Szczebra represent a boreal floristic succession with *Pinus – Betula – Larix* and *Azolla filiculoides* during the first (I) temperate substage and *Pinus – Azolla – Picea (Azolla – Salvinia)* with thermophilous taxons *Carya, Juglans, Celtis, Eucommia, Salvinia, Trapa* during the second (II) temperate (climatic optimum) substage. Three cold substages are generally characterized by boreal and subarctic vegetation (Janczyk–Kopikowa, 1996; Ber *et al.*, 1998).

Also in the recently elaborated pollen diagram from Kalejty (Winter, 2001) two warm stages and three cold stages have been distinguished. As explained in the pollen diagram by author (Winter, 2001), at the beginning of the older warm stage (Kal 2–Kal 4) pollen spectra indicate *Pinus* pollen to be dominant with admixture of *Picea, Betula, Quercus* and high frequencies of *Cyperaceae*. Gradually boreal forest is enriched by *Alnus, Ulmus, Tilia* and *Corylus*. The assemblage indicates the presence of mixed coniferous – deciduous forest. The occurrence of deciduous forest with high frequencies of *Quercus, Carpinus* as well as considerable share of *Ulmus* and *Corylus* is characteristic for the younger warm period (Kal 6–Kal 8). The presence of *Juglans, Carya, Ilex, Viscum, Hedera* and *Azolla* indicates temperate climatic condition. At the upper part of this stage pollen *Tsuga* occurs.

The cold stage separating two warm stages is marked by very high values of *Betula* and herb plants: *Cyperaceae, Gramineae and Artemisia*, also by *Betula nana* type pollen.

Pollen succession from the Kalejty section is correlated with the Augustovian pollen successions from the Szczebra section. Nevertheless there are differences among both succesion. The Augustovian pollen succession from Szczebra is discontinuous. The first stage with abundance of *Quercus, Ulmus, Tilia* and *Corylus* pollen has not been recorded in the Szczebra pollen diagram. Other lake sediment profiles such as Czarnucha, Żarnowo, Komorniki and Sucha Wieś are under palynological studies (Janczyk–Kopikowa; Winter).

In the lake sediments of the Augustovian Interglacial stage at Szczebra 11 bivalve species, 7 gastropod species, fragments of other animals (Coleoptera, Pisces, Aves) and 14 ostracod species (Skompski, Ber, 1999) were found. On the base of some molluscan species (*Bithynia leachi, Pisidium supinum*) we can interpret the moderate climatic conditions. Another molluscan species (*Valvata naticina, Pisidium supinum, Sphaerium solidum*) indicale that in this lake basin moving water existed $\delta^{18}O$ and $\delta^{13}C$ values were examined at Kalejty and Szczebra profiles from Quaternary lake sediments of the Augustovian Interglacial (Jędrysek, 1997 – unpublished; Nitychoruk *et al.*, 2000). Isotope examination indicated the presence of a considerable redeposition of older sediments during lower lake levels mostly during the cool intervals.

The interglacial series at Szczebra and Kalejty and in other profiles are underlain by thick layer of the lodgement till and covered also by thick lodgement till. The lowermost (older) till is characterized at Szczebra profile by the following petrographical coefficients: K/W = 0.86; O/K = 1.23; A/B = 1.12 and Dp/Wp = 0.46 where: O – total pebbles of sedimentary rocks; K – total of crystalline rocks and quartz; W – total of carbonate rocks; A – total of rocks non-resistant to destruction; B – total of resistant rocks; W_p – limestones of northern origin and D_p – dolomites of northern origin. Typical local rocks of this till are Palaeocene marls and their content is very high (to 32.2%). For upper (younger) till horizon the values of petrographical coefficients are: K/W = 0.61; O/K = 1.70 i A/B = 1.58. The dominant local rocks are Palaeocene and Cretaceous marls and limestones (up to 5.8%). The Augustovian Interglacial sediments according to presented data, are underlain by a till of the Narevian (Menapian) Glaciation and are overlain by the till of the Nidanian (Glacial A – Cromer Complex) Glaciation. The lithostratigraphy of the tills was established on the basis of petrographic and lithologic studies (Czerwonka, Krzyszkowski, 1995 – unpublished) and those results are correlated to results of the palynological studies.

The age estimates based on palaeomagnetism indicate, that Augustovian Interglacial lake sediments at Kalejty borehole developed during the time interval between the Bruhnes normal polarity and the Matuyama reversal polarity. Therefore these sediments correlate to the oxygen stage 18/19. This correlation corresponding with Cromer I (Waardenburg) Interglacial (Turner, 1996) position.

However, the Augustovian Interglacial sediments according to lithologic and petrographic study

(Czerwonka, Krzyszkowski, 1995 – unpublished) as well as by geologic setting are underlain by till of the Narevian (Menapian) Glaciation and are overlain by the till of Nidanian (Glacial A – Cromer Complex) Glaciation. Its coexistance with Bavelian (Leerdam) Interglacial position (West, 1996; Zagwijn, 1992, 1996; Zagwijn & De Jong, 1984). As it is visible, the setting and age of the Augustovian Interglacial has not been definitely determined and correlation with interglacial sites in neighbouring areas can be now provisional only (Ber *et al.*, 1998; Ber, 2000; Marks, 2000). Other profiles are presently under palaeomagnetical study.

LATE GIVETIAN ACANTHODIAN FROM THE PECHORY QUARRY (PSKOV DISTRICT, RUSSIA)

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Acanthodians are relatively rare and not very diverse in the Gauja Regional Stage deposits (Late Givetian). They are represented by isolated remains of *Acanthodes?*, cf. *Cheiracanthus*, *Devononchus*, *Haplacanthus*, *Homacanthus*, *Diplacanthus*, *Rhadinacanthus*, *Nostolepis*, *Nodacosta* (Valiukevičius, 2000). However, some forms are very peculiar and unique in the preservation. A new genus and species of mesacanthids – *Lodeacanthus gaujicus* Upeniece – has been described based on the articulated specimens from the Lode Quarry (Upeniece, 1996). This species was just known from one locality in Latvia hitherto. The isolated minute scales with unornamented crown and a fragment of fin spine were found in the clays of Pechory Quarry during the field trip in 2001. These remains were preliminary determinated as *"Acanthodes"* sp. But they differ from the taxa described by Valiukevičius as *Acanthodes?* (1985). It was necessary to compare them with *Lodeacanthus gaujicus* for the definition of their real taxonomic position. A study of scale structure of *Lodeacanthus gaujicus* was proposed for that aims.

Lodeacanthus gaujicus. The squamation is almost complete developed in the specimens which are more than 2.5 cm in the length. It occurs in the whole body, except the narrow line between the pectoral fins, the gill region and the lateral walls of fins. Several zones of squamation can be traced on the body of adult individuals (Upeniece & Beznosov, 2002, *in press*). Most part of the lateral side (zone 3) is covered by the scales, similar in the shape and size.

The size (length of the base) of trunk scales ranges from 0.08 up to 0.17 mm in adult individuals (4 cm long). The crown has commonly round or oval shapes with elongated and sharp posterior process. The surface of the crown is smooth and flat. The poor preservation of the scales does not allow to determine the presence of microscopic pores. The neck is well developed. The base is convex, the length is about 2/3 of crown length. The base width usually overpasses the length in the scales of lateral side except the caudal region.

The histological structure corresponds to the *Acanthodes* type. The scale base is composed of the acellular bone. The presence of vascular canals there has not been recognized. The crown consists of the mesodentine and is covered with relatively thin enameloid layer. The system of dentine tubules is well developed in the crown. The circular canals of the basal part of crown are located accordingly to the growth zones of scale. They became thicker along anterior edge and have several short branches. The ascending canals are more developed in the posterior part of crown. They are represented by a few relatively straight and thick main tubes with the simple branches. The latter often penetrate into the adjacent growth zones. The radial and compicated primordial dentine canals are absent.

The fin spines are slender and ornamentated on each sides with a deep anterior groove added with the several narrow grooves. In the adult individuals the length of spines in pectoral, dorsal and anal fins is up to 5.8 mm, in the pelvic fin – up to 3.6 mm, in the intermediate fin – up to 2.4 mm (Upeniece, 1996).

"Acanthodes" sp. Most of the scales are relatively larger than the trunk scales in adult individuals of *Lodeacanthus gaujicus*. The length of base ranges from 0.10 to 0.18 mm. The crown is rombic in the shape with the microscopic pores on the surface. These scales resemble the scales of *Lodeacanthus gaujicus* in another morphological features. Their histological structure is similar but there are some differences:

1) more numerous growth zones; 2) relatively thicker enameloid layer; 3) ascending dental canals of the crown are more numerous and stronger branched. Probably these differences such as the presence of narrow vascular canals in the base depend on the preservation of those scales because they occur as the isolated state. A fragment of fin spine, 1 mm long was also found in the assemblage with those scales. The spine transverse section is similar to that of *Lodeacanthus gaujicus*.

Thus, acanthodian remains were found in the Pechory Quarry much resemble the scales and fin spines of *Lodeacanthus gaujicus*. Possibly the differences between them in the size and shape of the scales depend on the onthogenetic variability. In that case the larger body length of *Lodeacanthus gaujicus* known by the complete specimens is not a limit for that species.

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ORDOVICIAN–SILURIAN BOUNDARY IN THE SUBPOLAR URALS, RUSSIA

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The position of the Ordovician–Silurian boundary in the Timan–Northern Ural region has been under discussion for a long time. During the second half of the last century several possible boundary levels were proposed and subsequently discarded as later studies indicated that all of them still lied well in the Ordovician. The special field-meeting devoted to this problem was held in the Kozhym River region in 1987, in the course of which several sections supposed to expose the boundary were examined (Tsyganko and Chermnykh, 1987). Although officially the Ordovician–Silurian boundary in the Subpolar Urals (as proposed in the stratigraphic scheme of the Timan–Pechora region, Kaljo, 1987) has been considered to correspond to the contact between the Malaya Tavrota and Yaptikshor formations, no general agreement about its location in the sequence was achieved during the field-meeting. Shortly after the meeting it appeared that the Yaptikshor Formation still contained the Late Ordovician *Proconchidium–Holorhynchus* brachiopod and *Aphelognathus–Belodina* conodont assemblages. Hence, in the new stratigraphical scheme for the Urals (Antsygin et al., 1993) the Ordovician–Silurian boundary was already drawn between the Yaptikshor and Dzhagal formations. During the 1987 field-meeting, mapping geologists from the Geological Producing Enteprise "Polyarnouralgeologiya" (Vorkuta) demonstrated a section (Kozhym-108, not studied during the meeting) exposing a continuous succession of strata across the problematic interval.

The Kozhym-108 section is located in the region of the mouth of the Yarenej-shor creek, on the right bank of the Kozhym River (Beznosova, 2000). The strata from the lower part of the Upper Ordovician Malaya Tavrota Formation below up to (at least) the Telychian Marshrutnyj Formation above crop out. The lower part of the section, up to the lower Dzhagal Formation, is exposed almost completely.

In the Kozhym-108 section the lower boundary of the uppermost Ordovician unit in the region, the Yaptikshor Formation, corresponds to a sharp contact between light-grey thin platy dolostones below (Malaya Tavrota Formation) and grey massive organogenous dolostones with *Proconchidium muensteri* coquinas in its base above (Yaptikshor Formation). The Yaptikshor Formation is represented mainly by massive dolostones. It is evident that originally the rock contained in abundance sorted and unsorted bioclastic material, and macrofauna (stromatoporoids, tabulate corals, rugose corals, brachiopods, gastropods, etc.), but due to very heavy dolomitization most of the fossils are strongly recrystallized and unidentifiable. Few better preserved specimens have been identified only at some levels (Figure), indicating that at least up to the lowermost part of bed 20 the rocks are of Late Ordovician age. As no fossils were identified from the 55–60 m thick strata above this level (main part of bed 20, bed 21, and the lower part of bed 22; Figure), the Ordovician–Silurian boundary was tentatively drawn above the level of the last occurrence of *P. muensteri* (Beznosova, 2000; Figure).

At the boundary between beds 20 and 21 the massive dolostones characteristic of the Yaptikshor Formation are replaced by thin- to medium-bedded dolostones that have been strongly fragmented due to the tectonic processes during the formation of the Ural mountains and evidently represent already the Dzhagal Formation. Only few unidentifiable specimens of Pentameridae, and of a tabulate coral identified as *Mesofavosites* sp., were found in the studied interval of the Dzhagal Formation (about 35–40 m above its base). Relatively well preserved Silurian pentamerides, identified as *Pentamerus* sp. (aff. *oblongus*) (Beznosova, 2000), appear about 230 m above the uppermost level with *P. muensteri*. Morphologically, these specimens are closest to the forms occurring in the uppermost part of the Nurmekund Formation (Aeronian) in Central Estonia (Nestor et al. in press).

The rarity of fossils, and their almost complete absence in an up to 60 m thick interval (Figure) necessitated some additional ways beside the traditional biostratigrahpic method to identify the level of the Ordovician–Silurian boundary in the studied section. The $\delta^{13}C$ studies in combination with biostratigraphy have proved highly promising for detailed stratigraphy. As brachiopod shells are preserved only at some levels in the

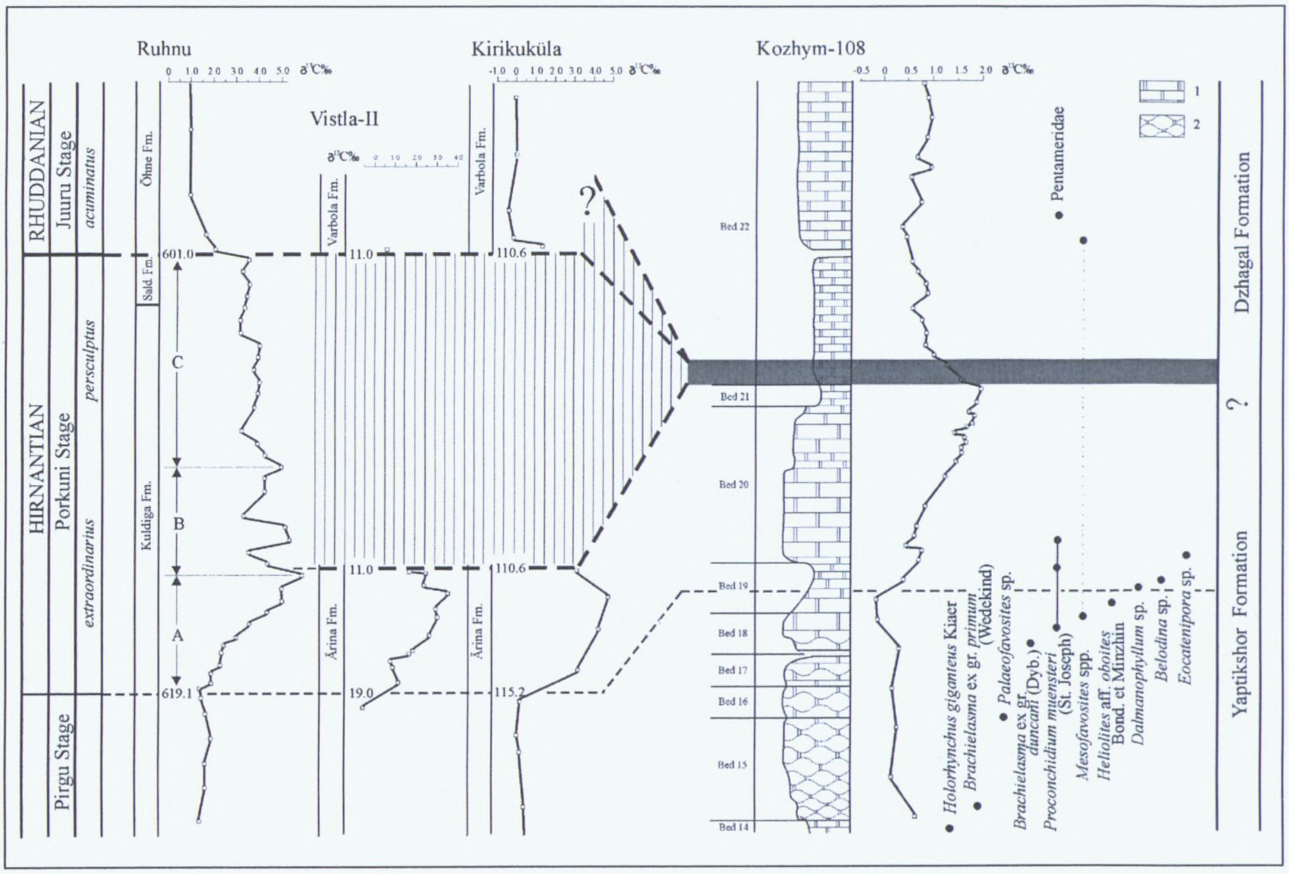

Figure. The Ordovician–Silurian boundary beds in three Estonian sections and the Kozhym-108 section correlated by isotope data. The $\delta^{13}C$ curves for the Ruhnu and Vistla-II cores are taken from Kaljo and Martma (2000) and Kaljo et al., (2001). A, B and C mark the intervals of the curve discussed in Kaljo et al., (2001) and in the text. Dashed lines are correlation lines. Lithology: 1, bedded dolostone; 2, nodular dolostone. Vertical ruling shows gaps; grey belt in the Kozhym-198 section indicates the most probable position of the Ordovician–Silurian boundary. Sald., Saldus Formation.

Kozhym-108 section, bulk rock samples (in total 49) were studied. The method used has been discussed earlier (e.g. Kaljo et al., 1997). The $\delta^{13}C$ sequence measured in the Kozhym-108 section was compared with the data from Estonia. The $\delta^{13}C$ curve across the Ordovician–Silurian boundary from Estonia is correlated with those from elsewhere (including Dobb's Linn, Scotland – Kaljo et al., 2001) and is, in that way, precisely linked to the graptolite biozonation. This gives us a good opportunity to use the $\delta^{13}C$ data for dating rocks close to the Ordovician–Silurian boundary.

The most complete latest Ordovician sequences in Estonia have been studied in the Kardla and Ruhnu drill cores (Kaljo et al., 2001). In the Ruhnu core, the $\delta^{13}C$ curve was subdivided, based on its general morphology, into three main parts (Kaljo et al., 2001; fig. 4). The lowermost part ("A" in Kaljo et al., 2001; Figure) is characterized by steady increase in $\delta^{13}C$ values up to the first peak. This strongly resembles a similar pattern in North Estonian sections (e.g. Vistla-II and Kirikuküla; Figure). The next part (B) is termed as an "extended plateau" in Kaljo et al. (2001) and is characterized by variable high values of $\delta^{13}C$. The uppermost part of the section (C) displays less variable and continuously decreasing $\delta^{13}C$ values. At the beginning of the Silurian the $\delta^{13}C$ values decrease rapidly. In North Estonian sections only the strata corresponding to part A of the $\delta^{13}C$ curve in the Ruhnu core are preserved. The higher strata (i.e. the intervals corresponding to parts B and C in the Ruhnu core) correspond to a gap (recognized also based on the biostratigraphical data) (Kaljo et al., 2001).

Comparison of the $\delta^{13}C$ curve from the Kozhym-108 section with data from Estonia shows that this curve maches almost precisely with those recognized in North Estonian sections (Figure; Kaljo et al., 2001, fig. 3). Based on the general configuration of the $\delta^{13}C$ curves from these regions, several conclusions can be made.

In the lower part of the Yaptikshor Formation (from bed 15 up to the lower half of bed 19), $\delta^{13}C$ values are low and display a gradually decreasing trend. Similarily, in Estonia low $\delta^{13}C$ values have been reported from the middle Ashgill Pirgu Stage (Kaljo et al., 2001; Figure). Also, the occurrence of *Holorhynchus giganteus* in the lower Yaptikshor Formation suggests the correlation of this interval with the Pirgu Stage in Estonia (and with the pre-Hirnantian strata elsewhere) (Figure; Beznosova, 2000). In the Baltic, *H. giganteus* has been found in the upper part of the Pirgu Stage. Also, it has been proved that the strata with *H. giganteus* elsewhere are older than the Hirnantian (Brenchley et al., 1997).

In Estonia, considerable increase in $\delta^{13}C$ values begins in the lowermost strata of the Porkuni Stage (Kaljo et al., 2001). In the Kozhym-108 section this event occurs in the middle of bed 19 (Figure). General configuration of curves in both regions is very similar (Figure). In North Estonia, the $\delta^{13}C$ values reach a maximum in the upper part of the Ärina Formation of the Porkuni Stage (Kaljo et al., 2001), in the Kozhym-108 section in the uppermost part of bed 21 (Figure). Upwards in the section, in both regions the $\delta^{13}C$ values drop rapidly to almost the level they had before the positive excursion started. In North Estonia, this seemingly rapid decrease in the $\delta^{13}C$ values is caused by a major gap in the sequence corresponding to the *persculptus* graptolite Biozone (Kaljo et al., 2001). In South Estonia, in the Ruhnu core where the section is more complete, after a minor drop in the $\delta^{13}C$ values above the first peak in the curve, the values increase again and then start to decrease gradually.

Considering the general configuration of the $\delta^{13}C$ curve from the Kozhym-108 section, it is evident that also here the strata corresponding to the intervals B and C in the Ruhnu core are missing like in the sections in North Estonia (Figure). It means that, based on the correlations between Estonia and Dobb's Linn (Kaljo et al., 2001), in the Kozhym-108 section most probaly: (1) the upper part of the Yaptikshor Formation (upper half of bed 19 and bed 20) and the lowermost Dzhagal Formation (bed 21, the lowermost part of bed 22?) correlate with the *extraordinarius* graptolite Biozone (and, accordingly, with the lower Hirnatian); (2) the upper Hirnantian strata correspond to a gap; (3) the Ordovician–Silurian boundary lies at or above the contact of beds 21 and 22.

To determine the age of the basal Silurian strata in the Kozhym-108 section, additional biostratigraphical studies are needed. The lowermost Silurian fossils in the section, *Pentamerus* sp. (aff. *oblongus*), were identified about 200 m above the proposed here system boundary. These pentamerids are most probably identical to those known from the Aeronian strata in Central Estonia (see above).

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DEVONIAN HETEROSTRACAN PTERASPIDOMORPHS (VERTEBRATA) FROM SEVERNAYA ZEMLYA (RUSSIA) – NEW DATA ON TESSERASPIDS

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Severnaya Zemlya (Kara–Tajmyr palaeocontinent) is a key-zone to understand the palaeogeographical relationships of the Old Red Sandstone Continent and Siberia in Mid–Palaeozoic times. Its Upper Silurian–Devonian sedimentary sequence bears a rich fauna of heterostracans which authorizes biostratigraphical correlations with mainly Spitsbergen (on the Barentsian palaeocontinent), but also with the Canadian Arctic (on the Old Red Sandstone Continent) and Central Taimyr–NW Siberia (Siberian palaeocontinent). This fauna is composed of various assemblages from the Ludlow (Upper Silurian) to the Frasnian (Upper Devonian), with the richest assemblages in the upper Lochkovian, and a gap in the Emsian (Lower Devonian) [1].

Among these assemblages, new corvaspid material has been recently described from the Lochkovian, Severnaya Zemlya and Pod'emnaya formations of October Revolution Island, as *Corveolepis elgae* Blieck and Karatajūtė-Talimaa, *Corveolepis*? sp. cf. *C.*? *graticulata* (Dineley), and Corvaspididae gen. et sp. indet. [2].

Here we focus on new material of tesseraspids from the uppermost Severnaya Zemlya Formation, where the tesseraspids have been collected in association with corvaspids (see above), articulated and disarticulated acanthodians, and rare but big specimens of osteostracans. The tesseraspid material is not abundant, and most often preserved as a «mish-mash» of broken fragments of head carapaces, except a few partly articulated specimens. Among the latter, we redescribe in detail the holotype of *Tesseraspis mosaica* Karatajūtė-Talimaa 1983 [3], whose head carapace is preserved as a flattened muff of adjacent platelets. All the other partly articulated specimens are of bigger size than the holotype, so that we are not sure whether or not they belong to the same species. So, we keep an open nomenclature for these specimens, that is, *Tesseraspis* sp.

This material is compared to the already published tesseraspid taxa, viz., *T. mutabilis* (Brotzen 1934) [4], *T. tessellata* Wills 1936 [5], *T. toombsi* Tarlo 1964 [6], *T. orvigi* Tarlo 1964 [sic], *T. denisoni* Tarlo 1964, and *T. talimaae* Tarlo 1965 [7]. All the published tesseraspid taxa are known after rare material. This material is always incomplete: no trunk and tail in connection with head carapaces are known. So, the intraspecific variability is unknown, and the distinction between species is mainly based on the detail of the «ornamentation» (or better superficial ultrastructure) of the platelets of the head carapaces, which is unsatisfactory.

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THE UPPER CAMBRIAN OF SEVERNAYA ZEMLYA: NEW PALAEONTOLOGICAL DATA AND THEIR UTILITY FOR PALAEOGEOGRAPHY

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Severnaya Zemlya, together with northern Taimyr, belongs to the North Kara plate (1), palaeogeographical position of which during the early Palaeozoic remains controversial: it may have been a part of separate microcontinent Arctida, or contiguous with Siberia, Baltica or Laurentia.

A detailed palaeontological study has been carried out in the southern part of October Revolution Island of Severnaya Zemlya. The new material originates from the upper part of the Upper Cambrian Kurchavinskaya strata exposed in the middle reaches of the Kruzhilikha River. Here, grey to dark grey shales with limestone concretions and interlayers of rare sandstones crop out, providing a rich record of inarticulate brachiopods, helcionelloid molluskcs (*Helcionella* sp.), ichnofossils of several patterns, including *Cruziana semiplicata*, trilobites (*Kujandaspis ketiensis*, *Maladiodella* aff. *abdita*, *Protopeltura holtedahli*, olenid indet) and brachiopods (*Billingsella* sp., *Finkelnburgia* (?) sp., huenellid gen indet, eoortids). This succession is overlain by grit sandstones of the Lower Ordovician Kruzhilikhinskaya Formation with an angular unconformity. Cambrian diagnostic acritarchs appear shortly below this unconformity, including *Acanthodiacrodium commune*, *A. partiale*, *A. perspicuum*, *A. unigerminum*, *Impluviculus muntiangularis*, *I.* aff. *vellosiusculus*, *Leiofusa stoumonensis*, *Lusatia dramatica* and *Verynachium dumontii*. According to this palaeontological record, the age of the clastics of the Kurchavinskaya Formation in the studied strata interval indicates a level between the *Olenus-Agnostus besus* to the *Peltura scarabaeoides* trilobite zones of Late Cambrian. These data confirm that there is a break in the late Cambrian stratigraphy of Severnaya Zemlya, represented by the early Ordovician unconformity.

In terms of biogeographic relationships, a comparison between these fossil groups is difficult to assess. For example, the acritarch association from Severnaya Zemlya strongly resembles in taxonomic composition the assemblage previously reported from Baltica. However, affinities to Siberia are not discarded, as no information is available on the late Cambrian acritarch distributions in Siberia. Fourteen trilobite taxa are identified from the Upper Cambrian strata of Severnaya Zemlya (2, 3, 4). Most of them show a broad geographic distribution; of the remaining forms the olenid trilobites *Parabolina* and *Parabolinites* occur in Baltica, Severnaya Zemlya and North America, *Protopeltura holtedahli* is common for Baltica and Severnaya Zemlya, while Siberia and Severnaya Zemlya share the occurrence of *Kujandaspis ketiensis* and *Maladiodella* aff. *abdita*. The species *Maladiodella abdita* shows a widespread peri-Gondwanan distribution as well, but no Laurentian occurrences of *Maladiodella* are recorded. We believe that the Upper Cambrian faunas from Severnaya Zemlya, in particular trilobites, show a stronger affinity to both Baltica and Siberia, even though faunas of these two regions mark their difference and endemism, than to the other areas. Based on the evidence presented, the best palaeolatitudional position of Severnaya Zemlya is estimated at approximate 30–40 ° southern latitude, and in the vicinity of Baltica and Siberia.

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SEQUENCE STRATIGRAPHY OF THE EASTERN PRE-TIMAN LOWER ORDOVICIAN DEPOSITS AND TENTATIVE CORRELATION WITH THE BALTIC REGION

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Lower Ordovician deposits represent the most ancient rocks of the Eastern Pre-Timan area, which is situated in the western part of the Pechora basin near the Timan ridge. They are not exposed at the surface and are penetrated only by deep well bores. The total thickness of the Lower Ordovician exceeds 1000 m. The area of Early Ordovician sedimentation was bounded by the ancestral Timan and Pechora–Ilych palaeo-uplift. In the Early Ordovician the Eastern Pre-Timan was occupied by a shallow sea that was constantly or periodically connected with an oceanic basin to the northwest of the Pechora plate (in recent geography).

The Lower Ordovician deposits are subdivided into the Sed'el and Nibel' formations. The type section of Sed'el formation includes two members; a lower member of polymict sandstones and an upper member of pure quartz sandstones. The Nibel' formation is traditionally divided into four members and is predominantly of red beds. From the base to the top it has a pink sandstone member, followed by a mudstone, sandstone-mudstone and finally a sandstone-siltstone member.

Investigations of the Lower Ordovician deposits of Northern America, Australia, Great Britain and Baltoscandia have revealed that breaks during the Cambrian/Tremadoc, Tremadoc/Arenig and Arenig/Llanvirn are connected with global falls of sea level, causing essential changes in structure of the organic world [1, 2].

Based on outcrop data, the Lower Ordovician succession of Baltoscandia is subdivided into three depositional sequences that represent third-order cycles of relative sea-level changes. From the base to the top they are as follows: Pakerort, Latorp and Volkhov [3]. In the Eastern Pre-Timan it is also possible to recognize three depositional sequences; one in the Sed'el formation and two in the Nibel' formation. Each of them represent upward-fining sequences with their boundaries marked by a basinward shift in facies.

Early Ordovician sedimentation features of the Baltic and the Eastern Pre-Timan regions are similar in many respects and were strongly eustatically controlled, as around the world. On the basis of our investigations, the Sed'el formation has been tentatively compared with the Pakerort sequence and the Nibel' formation with the Latorp and Volkhov sequences. The boundary between the Sed'el and Nibel' formations probably represent Tremadoc/Arenig boundary in this region.

In conclusion we should note, that this work represents the first attempt to create a scheme of interregional correlation for the Eastern Pre-Timan and Baltic regions Lower Ordovician deposits. This preliminary scheme will undoubtedly be corrected and supplemented by future studies.

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PLANT MACROFOSSILS AND STRATIGRAPHY OF THE QUATERNARY LAKE SEDIMENTS IN LATVIA

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Investigations of the macroremains in Latvia cover the whole section of the Quaternary deposits. This method is mainly used for investigations of the late glacial sediment sequence or interstadial and interglacial basin sediments, but the number of the investigated Holocene sites are very few.

The method has given a possibility to determine interruptions in the sedimentation of deposits, to find out redeposited sediment sequences, which contain mixed floras, as well as to make climatostratigraphical and paleoecological conclusions.

The results of paleocarpological investigations frequently have been expanded on by palynological data, but in some cases these data are diverse and complicate conclusions. That can partly be explained by more regional reflection of vegetation in pollen spectra, while paleocarpological data indicate composition of more local plant association.

The first extensive studies of plant macroremains are connected with investigations of the Late Glacial "*Dryas* floras". The *Dryas* flora of the Baltic Ice Lake sediments from Jelgava vicinity is the most famous in Latvia, which was investigated by E. Tolls, K. Kupffer, P. Galenieks, and V. Stelle. Characteristic flora with *Dryas octopetala* is distributed in lake sediments (Raunis, Lidumnieki, Tadaiki-Krikmani, etc.), which were determined as the Late Glacial interstadials. The *Dryas* flora has been also found in several sediment sequences of the Middle Pleistocene (Jaunskieri, Deseles Lejnieki, Vorzava), where it characterizes the vegetation of the Letiza (Elsterian) Late Glacial. That flora has been found in the deposits underlying the Pulvernieki (Holsteinian) interglacial lake sediments. It is characteristic that the macroremains of the late glacial basin sediments do not contain redeposited plant remains from the older deposits. Unfortunately, they are not very different. Sometimes only their location in the geological section or datings with geochronological methods allow to make stratigraphical decisions.

Mixed type periglacial floras, where that *Dryas* type macroremains are found together with redeposited termophilous plant remains from the preceding (Pulvernieki or Felicianova) interglacials are found in the Early Glacial deposits of the Latvija (Weichselian) and Kurzeme (Saalian) glaciations particularly in some intervals of the Rogali and Jurkalne layers.

The data obtained from the sections of the Rucava vicinity have been discussed by V. Seglinsh (1988), but interpretation of them still cause debates. The macroremains of the original mixed *Dryas* flora have been found in the depth interval 14.1–11.1 m of the borehole Rucava-186, which correspond to lower interval in pollen diagram subdivided by O. Kondratiene. Pollen spectra from this interval have been compared with those of lower interval from borehole Laukzeme-188 section and pollen diagram of Laukzemis-1 (Seglinsh, 1988).

The macroremains from depth interval 14.1–11.1 m of Rucava-186 borehole and 33.6-33.2 m of Laukzeme-188 are found in deposit layers with similar lithological characteristics: fine sand with micas and small lenses of washed plant remain detritus (Table 1). During investigations of sediment samples under the microscope was stated that sand contains also small fine silt lenses. It must be noted that plant remains were very mineralised. The pieces of amber have been found together with plant remains.

The following features are characteristic for the described macroremains complex:

1. The plant macroremains undoubtfully are allochtonous, which are proved by characteristic lamination textures indicated by:
   a) large amount of rounded wood pieces, which compose largest part of fine detritus;
   b) largest seeds are damaged, their surface are secondary corroded (particularly *Potamogeton* and *Myriophyllum* seeds).

Table. Macroscopic plant remains from Rucava (borehole 186) and Laukzeme (borehole 188) sites.

Name of taxon	Type of remains	Rucava 186			Laukzeme 188		
		sand, m			sand, m		
		13.7–14.1	12.8–12.1	11.1–11.95	33.4–33.6	33.4–33.2	33.2–33.0
		Number of specimens					
1	2	3	4	5	6	7	8
Trees and schrubs							
Picea sp.	fr. of needle	5					
Larix sp.	fr. of needle	1					
Salix polaris Whlb.	leaf	7					
Salix sp.	leaf	4					
Betula alba L.	nutlet	20					
B. nana L.	nutlet	200	7				
	fruit shells	2					
	leaf	7					
Betula sp.	nutlet		3		7	23	
	fruit shells		1				
Rubus idaeus L.	fr. of seed	1					
Kseromesophyte							
Caryophullaceae gen.	seed	1					
Potentilla anserina L.	nutlet	3					
P. cf. *norvegica* L.	nutlet		1				
Potentilla sp.	nutlet	16			1		
Dryas octopetala L.	fr. of leaf	3					

1	2	3	4	5	6	7	8
Telmathophyte and swamps plants							
Selaginella selaginoides (L.) Link.	macrospore	165	86	5	43	69	27
S. cf. *helvetica* (L.) Spring.	macrospore	13	2				
S. tetraedra Wieliczk.	macrospore	7					
Urtica dioica L.	nutlet	13	6	1			
Polygonum sp.	nutlet	1					
Ranunculus sceleratus L.	nutlet	16	10		1	2	
R. cf. *sceleratoides* Nikit. et Dorof.	nutlet	10					
Ranunculus sp.	nutlet	1	6				
Viola sp.	seed	17	1		1		
Ajuga sp.	seed		1				
Mentha sp.	nutlet						1
Amphiphyte							
Typha sp.	tegmen	1	1				
Alisma plantago - aquatica L.	seed	1	4			2	
Eleocharis palustris (L.) Roem. et Schult.	nutlet	2					
Scirpus sp.	nutlet		1			4	
Carex sp.	nutlet	160	5		3	15	3
Juncus sp.	seed	2					
Rumex maritimus L.	nutlet	18				5	
Rorippa palustris (L.) Bess.	seed	1					
Lycopus europaaeus L.	nutlet	1			1	1	
Aquatic plants							
Characeae gen.	oogonium	8	1		2		1
Isoetes lacustris L.	macrospore	1	2			1	
Salvinia natans (L.) All.	macrospore	39	19	1	18	20	10
Azolla interglacialis Nikit.	macrospore	200	150	7	23	62	13
Potamogeton filiformis Pers.	stones	1	1				
P. cf. *vaginatus* Turcz.	stones	1					
P. cf. *pusillus* L.	stones	2					
P. cf. *cirpus* L.	stones						
P. cf. *natans* L.	stones						
Potamogeton sp.	stones						
Zannichellia palustris L.	seed	60	5		3		
Zannichellia sp.	seed					1	
Caulinia minor (All.) Coss. et Germ.	seed	3					
C. cf. *goretskyi* (Dorof.) Dorof.	seed	1					
Najas marina L.	seed	1					
Sagittaria sagittifolia L.	seed	6	6				
Lemna trisulca L.	seed	1			1		
Batrachium sp.	nutlet	10	29	1			
Elatine cf. hydropiper L.	seed		2				
Myriophyllum cf. *spicatum* L.	stones		3				
Myriophyllum sp.	stones	1	16			1	
Hippuris vulgaris L.	stones	5					
Empetrum nigrum L.	stones		1				
Menyanthes trifoliata L.	seed	2					
Fragments of Mesozoic or Paleozoic macrospore	fragments of macrospore	12	3				

2. The macroremains of water and mire plants (*Potamogeton, Myriophyllum, Hippuris, Carex, Ranunculus sceleratus,* etc.) are widely distributed. However, they have not relevant sense for determination of age.

3. The megaspores of water fern *Azolla interglacialica* have been found in large quantities, particularly in borehole Rucava-186. The megaspores are very good preserved and have been found as "colonies" and clinched together in umbel or make some lenses together with herb or moss pedicels, which are interchanged with grey silt laminas. A significant part of megaspores has been flatted parallely to delaminating. Thereby, the remains of this plant are allochtonous in silt. Pollen analysis of borehole Rucava-186 has been made by O. Kondratiene and she considers *Azolla interglacialica "in situ"* (Seglinsh, 1988). Such large concentration (several hundreds per 1 dm^3 of sediments) of Azolla interglacialica megaspores usually can be found in the Middle or older deposits. They have not been found in the Holsteinian interglacial lake sediments elsewhere in Latvia, except for the Ulmale area, where separate megaspores have been found in marine sediments.

4. *Azolla interglacialica* megaspores have been found in sand together with few megaspores of *Selaginella helvetica* and *Caulinia* cf. *goretskyi* seeds, which are characteristic for Holsteinian interglacial. Nutlets of *Carex paucifloroides*, *Scirpus atroviroides* and *Ranunculus sceleratoides* have been found as well, which are characteristic for Sklova type flora (Velichkevich, 1982). Seeds of *Elatine* cf. *hidropiperus* have been first found in Latvia.

5. The tundra plant remain complex represented by large number of *Selaginella selaginoides* megaspores, *Betula nana* nutlets and even well preserved leaves, *Salix herbacea* leaves, separate fragments of *Dryas* leaves, has been found together with plant remain complex of the interglacial.

Mixed flora composition allows to suppose, that sediments can be accumulated during some stage of early glaciation, when in the surroundings of basin the tundra vegetation was already distributed. The allochtonous interglacial flora in general is characteristic for Middle Pleistocene, while large amounts of *Azolla interglacialica* megaspores and their good preservation level, as well as presence of some other flora representatives characteristic of Early Middle Pleistocene allow to suggest that also other deposits there can be redeposited, older than the Holsteinian Interglacial.

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THE GEOGRAPHY AND MARGINS OF BALTICA IN THE LOWER PALAEOZOIC

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The margins of Baltica today are entirely the products of post-Lower Palaeozoic tectonics. To the west they are bounded by the Scandian Caledonides of Sweden and Norway, to the east by a series of tectonised sutures in the Ural Mountains, to the north by the Barents Sea and Arctic Ocean and to the south largely by the Trans-European Suture Zone (TESZ).

The south-east corner lies in the Caspian Sea. However, the TESZ, although generally coinciding with today's Baltica margin, runs north of the Malopolska and Lysogory blocks, both of which are partly exposed in the Holy Cross Mountains of Poland. There is no doubt from both the Cambrian and Ordovician faunas of both those blocks that they formed a marginal part of the Baltica terrane itself during the Lower Palaeozoic: for example, the early Ordovician brachiopod Lycophoria (which is the only genus within its family) occurs in rock-forming abundance in Estonia, the Russian part of Baltica and Sweden, as well as in the Holy Cross Mountains, and is known from no other terrane. To the north, it was previously considered that late Ordovician brachiopods from the central part of the Taimyr Peninsula of Siberia indicated that perhaps it was also part of Baltica at that time,but a more thorough review has not confirmed this, and it is now certain that central and south Taimyr formed part of the Siberian plate during the Lower Palaeozoic. However, northern Taimyr, together with the islands comprising Severnaya Zemlya, are now known to have formed the Kara terrane, from which late Cambrian trilobites indicate links with both Baltica and Siberia: but the brachiopods from the same beds are largely endemic, which confirms the palaeomagnetic evidence that Kara was independent.

Palaeomagnetic evidence has also indicated that Baltica rotated by more than 100 degrees between mid-Cambrian and end-Ordovician times, which means that today's TESZ margin initially faced towards Laurentia. Baltica underwent a soft docking with Avalonia at about end-Ordovician (443 Ma) time, and with Laurentia in a substantial collision which formed the Caledonides from early Silurian (435 Ma) to early Devonian (405 Ma) time, with maximum activity at 418 to 412 Ma. During the Lower Palaeozoic, Baltica drifted from a high subpolar to temperate palaeolatitude in the late Cambrian and early Ordovician, through lower temperate latitudes in the later Ordovician, into subtropical and tropical palaeolatitudes in the Silurian. This migration is reflected in both sediments and faunas. The early Ordovician sediments consisted of clastics and cool-water limestones, but warm-water limestones are rare before the late Ordovician. The most substantial reefs are in the mid–Silurian of Gotland, Sweden. The faunas, particularly the shallow-water benthos, underwent a great increase in diversity as time progressed. In the early Ordovician they formed a distinctive Baltic Province since the surrounding oceans were so wide, but progressive faunal interchange occurred with time, firstly with Avalonia as the two terranes became closer and then later with Laurentia. After the Caledonian Orogeny, when Baltica formed part of the superterrane of Laurussia, there was much elevation of land, and this is reflected by the sediments and freshwater faunas of the Old Red Sandstone continent over much of the area.

EVALUATION CRITERIA OF THE STRATIGRAPHICAL SIGNIFICANCE OF THE QUATERNARY SEDIMENT SECTIONS

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The stratigraphical significance is the one of the most essential evaluation aspects of the Quaternary sediments sections during either the investigation or analysis of the research materials from earlier studies. The primary criterion of the stratigraphical significance for the Pleistocene deposits sections is the determination of the presence or absence of the interglacial or interstadial sediments sequence. Development of a scheme with a wide range of classified criteria is necessary for the further evaluation of the sediment sequences and correlation of data in a large area. Documentation of classified schemes for the Quaternary sequences haven't yet been found in the literature. The following attempt can be useful for the evaluation and correlation of the data for large number of the stratigraphical investigations.

The range starts with the above mentioned primary criterion, which is not important in cases when analysis and estimation of the sections has already been concluded and described in the publications or in materials of Geological Funds (archive of National Geological Survey).

The scheme of the evaluation criteria for stratigraphical significance contains 6 stages.

1. Evaluation of the characteristics of the geological structure.

According to this evaluation criteria all sections of the Quaternary sediments could be subdivided in the following groups:

1.1. The sections without any stratigraphical significance.

The sediments sequences do not contain any interval with organic matter even partly relevant for the stratigraphical investigations, neither any information which could be useful to forecast stratigraphical significance by indirect characteristics.

1.2. The sections with limited and problematic stratigraphical significance.

These sequences contain fragmentary sediments and deposits, which genesis is unclear, sometimes with tiny inclusions or interlaminas of organic material.

1.3. The sections with perspective stratigraphical significance.

The interglacial or interstadial type sediment sequence or sediments rich in remains of organic matter and fossils are represented, which beyond doubt prove stratigraphical significance of section.

2. Evaluation of the bedding conditions of the interglacial sediments.

Such evaluation would be certain at outcrops or sites where numerous test drillings are within close proximity thus characterising these conditions. Such sections can be subdivided into four groups according to the following criteria:

2.1. The sections of limited and partly problematic stratigraphical significance.

The interglacial and interstadial sediments are not located in the original bedding, where substantial relocation and fragmentation have been found as erratics, incorporations and inclusions.

2.2. The sections with conditional importance while additional information will be gathered.

These sections contain not enough indisputable evidence concerning sediments bedding conditions and stratigraphical significance.

2.3. The sections useful for further evaluation of their stratigraphical significance.

Small deformations of interglacial and interstadial sediment have been stated in the sediments sequence, however, they are located in the original place of their formation.

2.4. The sections excellent for further evaluation of their stratigraphical significance.

The sediments sequences investigated are rich in organic matter, fossils and lay without any disturbance in respect to the original bedding.

3. Evaluation of the sediments genesis and composition characteristics significant for stratigraphical investigations.

The evaluation of the genesis and composition characteristics of the sediments before palaeontological investigations is the basic criteria for selecting the method of investigations, the type of sampling procedure and subsampling interval. The characteristics of the sediment composition are very important, when obtained palaeontological data have been analysed and proportions of remains both redeposited and synchronous for deposition have been estimated. That information is important also in further evaluation stages especially related to the specifics of the investigation. For this stage two section groups can be subdivided:

3.1. The sections without enough sufficient information for obtaining the certain stratigraphical conclusions.

The paragenetic series of slopes, proluvial, alluvial or terrigenic deposits of the basin shore zone form the stratigraphically most significant part of sediment sequence. Such deposits, especially, if they are comparatively coarse, could be subsampled with larger intervals and sediments may contain a significant number of redeposited fossils.

3.2. The provisory most optimal sections for evaluation of stratigraphical significance.

These sections contain mire or lake organic type sediments, especially peat and gyttja.

4. Evaluation of the completeness degree of the stratigraphically significant layers.

If lithologically expressed interruptions are to be stated in the sections of the layers or strata or the deposits are thin, then it is a base for assumption, that sediments of these sections do not reflect the entire respective time span. The palynological investigations are the most convincing criteria for all sections, which certify, if there are representation of all pollen zones within respective time in the exact sediment sequence. Three groups of section could be subdivided according to these criteria.

4.1. The sections of little stratigraphic importance.

These sections contain interglacial sediments of fragmentary character, where the most stratigraphically significant sediment part reflecting the climatic optimum is missing.

4.2. The sections of the limited stratigraphical significance, but, nevertheless, at times could be used for wider stratigraphical correlation.

The deposits of the respective interglacial time are represented incompletely, but at least partly reflect time of climate optimum, allowing conformity define their belonging to the certain interglacial.

4.3. Stratigraphically significant type sections.

The sections contain sediments sequences, reflecting full and sometimes even part of the previous and the following climatic cycle.

5. Evaluation of the investigation degree in details and complex.

This evaluation stage comprises evaluation of the frequency of analysed samples, the complexity of the pollen spectra components, the application of other palaeontological methods, necessary for investigations of that sediment type. Most important is to evaluate how the data obtained by other methods can substantially deepen the validity of the stratigraphical and palaeogeographical conclusions. In the investigations of the Baltic Sea sediments, for example, the most effective are combination of data from palynological and diatom studies and less important information could be obtained from the investigation of the other fossils. The sections according mentioned criteria could be subdivided into three groups.

5.1. The sections of the limited stratigraphical significance.

The obtained palynological data do not enough clearly reflect peculiarity of the separate pollen zone or their sequence, at least in the some interval of section.

5.2. The stratigraphically significant, but insufficiently investigated sections.

Palynologically clearly characterised sections, but other stratigraphical methods important for stratigraphical conclusions are not used.

5.3. Stratigraphically very significant or type sections.

The sections are investigated sufficiently in details.

<u>6. Evaluation of the obtained results from biostratigraphical investigations, solidity and unambiguity of their interpretation.</u>

The sections according this criterion can be subdivided into two groups.

6.1. The sections with limited or problematic stratigraphical significance.

The sediments sequences investigated in detail have pollen spectra with very specific characteristics, probably due to local or other conditions, which substantially complicate efforts to make certain justification of the stratigraphical conclusions.

6.2. The stratigraphically most significant sections.

The data of the sediments sequence investigated clearly reflect development of vegetation characteristic for exact climatic cycle, as well as, having characteristic regional features for those investigated area during determined cycle.

Evaluation practice of the stratigraphical significance of the Quaternary sections is based on the experience formed during a long period and is reflected by the range of the described evaluation criteria. Some researchers have practice only in the one particular aspect of the above mentioned criteria. However, sometimes this can be insufficient for evaluation.

ORDOVICIAN SEA-LEVEL CURVE: BALTOSCANDIAN VIEW

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The Ordovician epicontinental basin of Baltoscandia is characterised by an extremely low average rate of sediment accumulation, rarely exceeding 1–3 mm per 1000 years, very flat bottom topography, and a long-term tectonic stability. These unique conditions lead to the situation where the stratal architecture and facies distribution within the depositional sequences depends predominantly on eustasy Based on outcrop data the Ordovician succession of Baltoscandia has been subdivided into ten major depositional sequences. All these sequences represent third-order cycles of relative sea-level changes (in the sense of Vail et al., 1977), and have an average duration of between 1.5–3.0 and 8–9 m.y. For ease of reference and identification, individual names have been given to all the depositional sequences. From the base to the top they are as follows: (1) Pakerort; (2) Latorp; (3) Volkhov; (4) Kunda; (5) Tallinn; (6) Kegel; (7) Wesenberg; (8) Fjäka; (9) Jonstorp; and (10) Tommarp sequence (Dronov, Holmer, 1999).

The most prominent unconformities with extensive erosion of the underlying beds coincide with the base and at the top of the Ordovician succession as well as at the base of the Latorp and Wesenberg sequences. The strong erosion and development of regional unconformities can be regarded as evidences for a sea-level drops of a prominent magnitude comparable to the modern glacial regressions (about 100 m). The Latorp, Volkhov and Kunda sequences demonstrate a deepening of the basin starting after regression at the base of the Latorp sequence. The Volkhovian deposits are the most widespread and the total area of marine red beds in the Volkhovian exceeds the area they cover in the Latorpian and Kundan (Männil, 1966). The lower boundary of the Volkhov sequence is interpreted as a 2nd – type sequence boundary (Dronov, 2000) with a long period of still stand and non-deposition. The magnitude of the sea-level lowering probably did not exceed 10–20 m. The overlying Kunda sequence is very similar to the Volkhov sequence in its lithology. The magnitude of sea-level drop at the Volkhov/Kunda boundary was larger than that at the Latorp/Volkhov boundary (30–40 m).

There are no evidence of prominent erosion at the base of the Tallinn sequence and it is represented by more shallow water deposits as compared with the underlying Kunda and Volkhov sequences. The shallowing of the basin was not a result of forced regression but rather a consequence of an increasing sediment input. In the Tallinn sequence, the marine red beds in the central parts of the basin were replaced by a grey-coloured deposits. The organic-rich kukersite-bearing strata demonstrate progradational stacking patterns and form highstand systems tract of the sequence.

The Kegel sequence is comparable in lithology with the underlying Tallinn sequence. The unconformity at the base of the Kegel sequence is well developed only in north-eastern Estonia and north-western Russia, where a shallow-water kukersite-bearing facies are well developed. The sea-level drop probably did not exceed 10 m. The Kegel sequence is remarkable for its transition from cool-water temperate to warm-water carbonate sedimentation and the rapid growth of reefs (Dronov, 2002).

The unconformity at the base of the Wesenberg sequence is one of the most remarkable in all the Ordovician of Baltoscandia. The regression seems to be comparable in magnitude to that of the Volkhov/Kunda boundary and can be estimated as much as 40–50 m.

The base of the next Fjäka sequence is interpreted as a 2nd type sequence boundary. The black shale sedimentation reached its maximum in the transgressive systems tract of the Fjäka sequence, which indicates a rapid deepening of the central parts of the basin.

There is no extensive erosion at the base of the overlying Jonstorp sequence but the following transgression is marked by the return of marine red beds. The scale of this transgression seems to be more prominent than that of the Fjäka.

The topmost Tommarp sequence is poorly exposed. Most of its sediments were eroded during the subsequent regression at the Ordovician/Silurian boundary.

The reconstructed sea-level curve for the Ordovician of Baltoscandia is different from that of Vail et al., (1977) and Ross and Ross (1992, 1995) (Fig. 1). The North American models assumes a prominent sea level drop at the base of the Middle Ordovician and a long-term lowstand during all the «Volkhovian» and Darrewilian (80–100 m lower than in the Lower and Upper Ordovician). In contrast, the data from Baltoscandia points rather to a moderate sea-level drop at the base of Volkhov without any prominent erosion, of comparable scale to the erosion events at the base and top of the Ordovician or at the lower boundaries of the Latorp and Wesenberg sequences. Moreover, the Volkhovian and Kundan highstands seems to be the most prominent transgressions in all of the Baltoscandian Ordovician, which means that the Middle Ordovician was not a lowstand but rather a highstand interval. It is clear that further investigations of the Ordovician eustasy are required in order to solve these matters.

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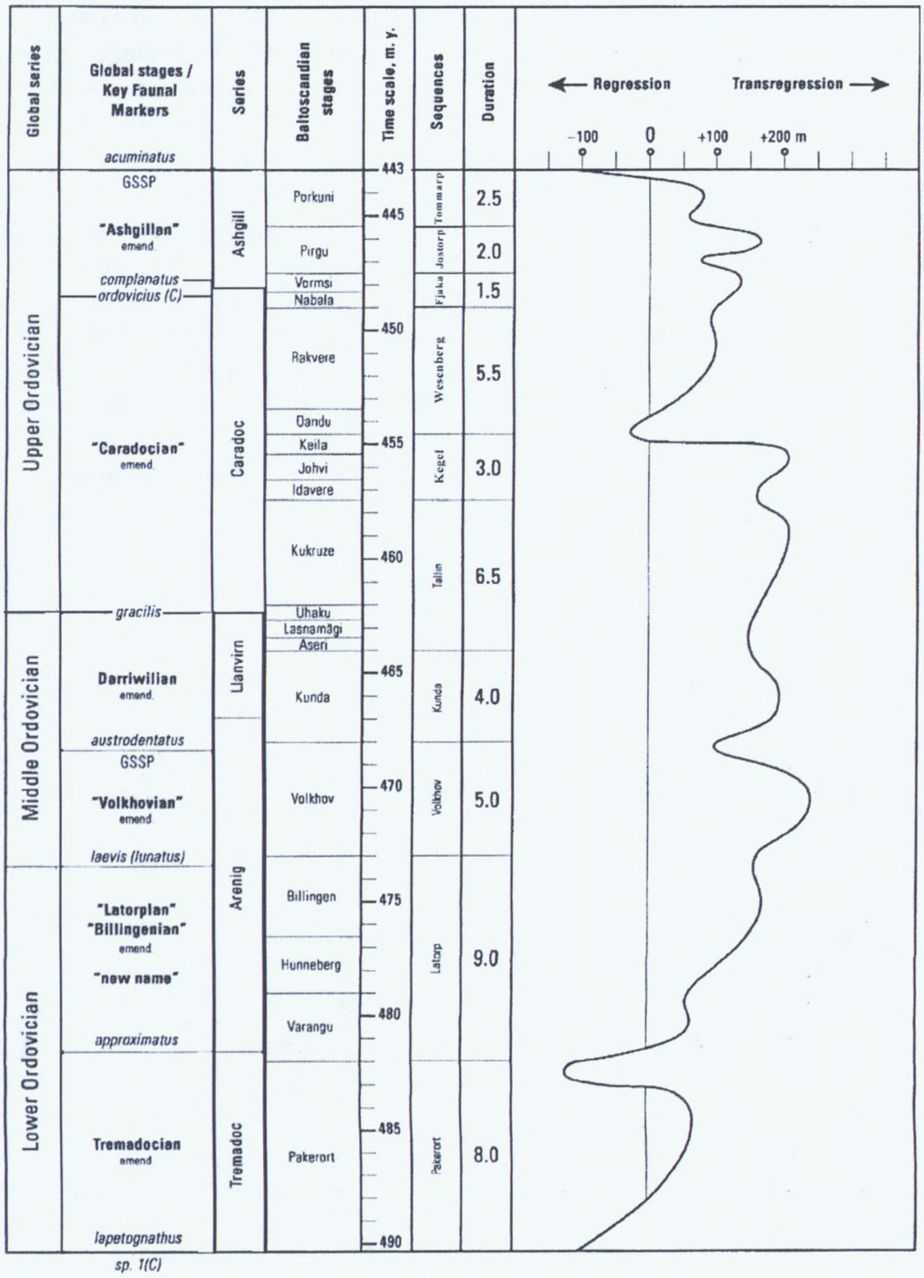

Fig. 1. Eustatic sea-level changes in the Ordovician based on Baltoscandian Basin.

CHANGES IN BENTHIC OSTRACODE ASSOCIATIONS ACROSS THE SARGAEVO–SEMILUKI BOUNDARY BEDS (FRASNIAN) IN THE MAIN DEVONIAN FIELD, RUSSIA

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In general terms, the Early–Middle Frasnian stratigraphical chart of the Main Devonian Field was worked out by D. V. Obruchev and R. F. Gecker in 1930–1932. Later it was somewhat detailed by R. F. Gecker, V. S. Sorokin and others. The Early–Middle Frasnian deposits were subdivided into a number of units based on the occurrence of the specific brachiopod and, partly, vertebrate associations (Snetnaya Gora, Pskov, Chudovo, Shelon', Svinord, Il'men, Buregi, and Snezha beds). The cyclicity of sedimentation and succession of the main taxonomic groups of fauna (mainly brachiopods) allowed to subdivide the Shelon' Beds into two units separated by a regional unconformity surface. In 1964 R. F. Gecker proposed to rename the Lower Shelon' and Upper Shelon' Beds into Dubnik and Porkhov Beds respectively. The lower boundary of the Porkhov Beds, corresponding to the lower boundary of the Semiluki Regional Stage, approximates to the proposed boundary between the Lower and Middle Frasnian.

Pioneer study of the Frasnian ostracodes in the Main Devonian Field conducted by M. A. Batalina was followed by E. M. Glebovskaja, V. S. Zaspelova, and V. G. Egorov. Their investigations resulted in a number of detailed monographs devoted mainly to the taxonomy of certain ostracode groups as well as to some biostratigraphical evaluation of ostracode associations (1, 2, 3, 4). According to the stratigraphical chart used at that time, the interval under consideration was referred to the undivided Shelon' Beds. The lack of knowledge of exact position of ostracode samples within the sections studied by these investigators does not permit to determine any peculiarities of the stratigraphical distribution of ostracodes in the Shelon' Beds. During the geological survey in 1998–1999 the key sections were restudied and an extensive ostracode fauna collected. The obtained material has allowed us to define more precisely stratigraphical range of ostracodes as well as to characterize ostracode associations in the Dubnik and Porkhov beds.

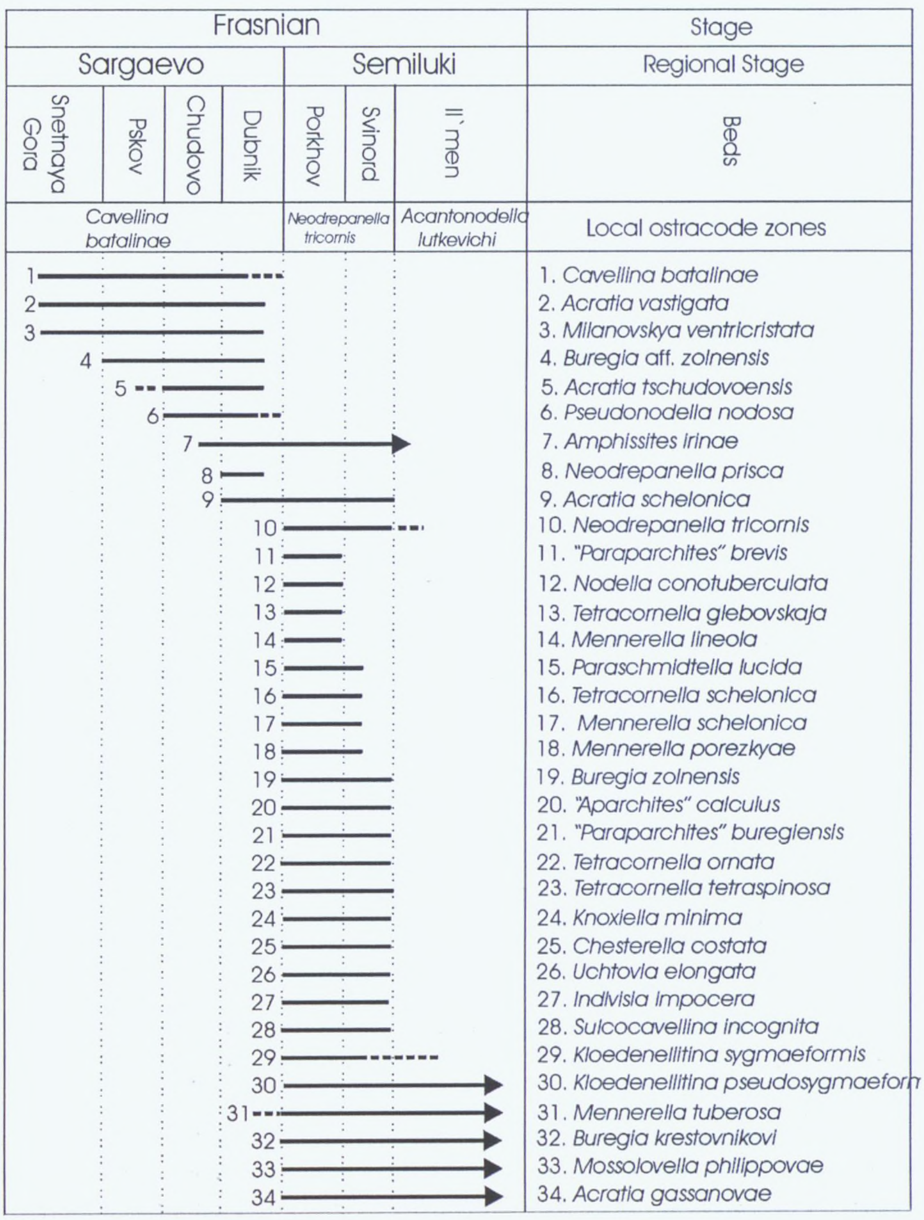

Figure. Stratigraphical distribution of the main ostracode species in the Dubnik and Porkhov boundary beds.

This level is characterized by facies and biotic changes as a result of an extensive transgression. Lithologically the sharpest changes are registered in the central part of the Main Devonian Field where carbonate deposits of the Porkhov Beds follow terrigenous, often gypsiferous, sediments of the Dubnik Beds. In the other parts of the Main Devonian Field facies changes are less remarkable and the boundary can be distinguished by the faunal changes. At this level an essential renewal of ostracode fauna is marked by the appearance of an abundant and diversified ostracode association in the Porkhov Beds (Figure). The main elements of this association are palaeocopids (especially neodrepanellids and nodellids) and kloedenellocopids. Podocopids are infrequent.

In the whole study area, Dubnik ostracode species diversity is low. Taxonomic composition of ostracodes coming from the Dubnik Beds is represented mostly by podocopids and platycopids. Most frequent genera are *Acratia* (dominant in terms of numbers of species) and *Cavellina* (dominant in terms of numbers of individuals). Palaeocopids and kloedenellocopids are rare.

Generally, remarkable changes in the benthic ostracode associations across the Sargaevo–Semiluki boundary beds similar to those within the Main Devonian Field can be also traced in the Central Devonian Field and, to a lesser degree, within the Pripyat' Trough and the South Timan.

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STRATIGRAPHY OF THE EARLY PALAEOZOIC AND NEOPROTEROZOIC STRATA IN NE–GERMANY

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The Early Palaeozoic and the Precambrian of NE–Germany and the neighbouring Baltic Sea is known from the subsurface by a few boreholes on the mainland, the island of Rügen and the easterly and north-easterly adjoining offshore region (fig. 1). The Caledonian Deformation Front (CDF) subdivides the area into two regional geological units. Caledonian deformed sediments occur south of the CDF. They belong to the external zone of a Caledonian tectogene, which stretches in WNW–ESE direction from Denmark, across N–Germany to Poland (Katzung, 2001). This unit is called the Rügen Caledonides (Katzung et al., 1993). The second unit north of the CDF is marked by Early Palaeozoic cover sediments of the East European Platform. This unit in the foreland of the Rügen Caledonides is situated in the Adlergrund, a nearby shallow ground of the Baltic Sea.

The Rügen Caledonides are divided into five lithostratigraphic formations (Beier et al., 2001b). Due to the Caledonian thrust tectonics all successions are disturbed. The formations are limited by faults and/or cut by erosion.

The Varnkevitz–Sandstein-Formation, the Arkona–Schwarzschiefer–Formation and the Nobbin–Grauwacken–Formation are of Ordovician age (Servais & Katzung, 1993; Maletz, 1998; Samuelsson, 1999). They were deposited in different tectonic settings in the forefield of the active margin of Avalonia, a terrane that took off Gondwana and moved north colliding with Baltica. The Varnkevitz–Sandstein–Formation consists of an interbedding of mature sandstones, siltstones and mudstones. The Arkona–Schwarzschiefer–Formation is characterised by hemipelagic mudstones with distal turbiditic intercalations deposited in deep water. The Nobbin–Grauwacken–Formation consists of massive greywackes and mudstones. According to their detrital composition, the greywackes represent the debris of a complex structured continental

Fig. 1. Location of the boreholes with Early Palaeozoic and Neoproterozoic rocks. CDF – Caledonian Deformation Front.

magmatic arc, with fragments of oceanic crust as well as shelf deposits, transported by turbidity currents into a deep basin (Giese et al., 1994).

The Schwarbe–Buntschiefer–Formation and the Lubmin–Sandstein–Formation consist of red and green coloured mudstones and sandstones. Until now, they were regarded as Early Ordovician rocks with a Gondwanan (Avalonian) provenance (Franke & Illers, 1994; Giese et al., 1994). However, recent isotope studies point to a deposition on the shelf of the palaeocontinent Baltica (Dallmeyer et al., 1999; Giese et al., 2001; Tschernoster, 2001). Sedimentary structures refer to a higher age of the fossil free rocks. Small pebbles in a clayey matrix and a varve-like interstratification of mudstones and siltstones occuring in parts of the Schwarbe–Buntschiefer–Formation hint to a glaciomarine deposition most probably during the late Neoproterozoic (Beier, 2001). These new results enable an interpretation of the crustal composition in NE-Germany. The presence of deformed Baltic sediments in the borehole Loissin 1 (fig. 1) as well as magnetotelluric data (Hoffmann et al., 1998) suggests that the East European Craton reaches probably to the Anklam fault. However, Zircon xenocrysts of 1,45 Ga in Permian magmatic rocks in the boreholes Friedland 1 and Penkun 1 (Breitkreuz & Kennedy, 1999), south of the area shown in fig. 1, might indicate that Baltic basement is present even further to the south.

The borehole G 14-1 is the only outcrop in the foreland of the Rügen Caledonides on German offshore area. It exposes an Early Palaeozoic cover sequence of the East European Platform which can be correlated, with minor differences, directly with the successions on the island of Bornholm (Denmark) and in Scania (Sweden). Therefore, the names of the lithostratigraphic formations were derived mainly from the traditional names of the Early Palaeozoic successions of southern Scandinavia (Beier et al., 2001a).

The succession of the G 14-1 borehole starts with the Adlergrund-conglomerate which has no equivalent in southern Scandinavia. It is overlain by about 250 m thick lower Cambrian sandstones of the Adlergrund–Sandstein–Formation and the Laesa–Sandstein–Formation. The middle Cambrian to Arenigian is represented by a condensed succession of about 35 m of black bituminous mudstones with fossiliferous limestone intercalations. Of interest is the presence of the Bjorkasholmen limestone between the Alum Shale and the Toyen Shale which is proved for Scania as well but not for Bornholm (Stouge, 2001; fig. 2). The Llanvirnian is missing almost completely. The approximately 50 m thick Upper Ordovician consists of mudstones of the "Dicellograptus"–Schiefer–Formation (Caradocian) and an interbedding of carbonaceous

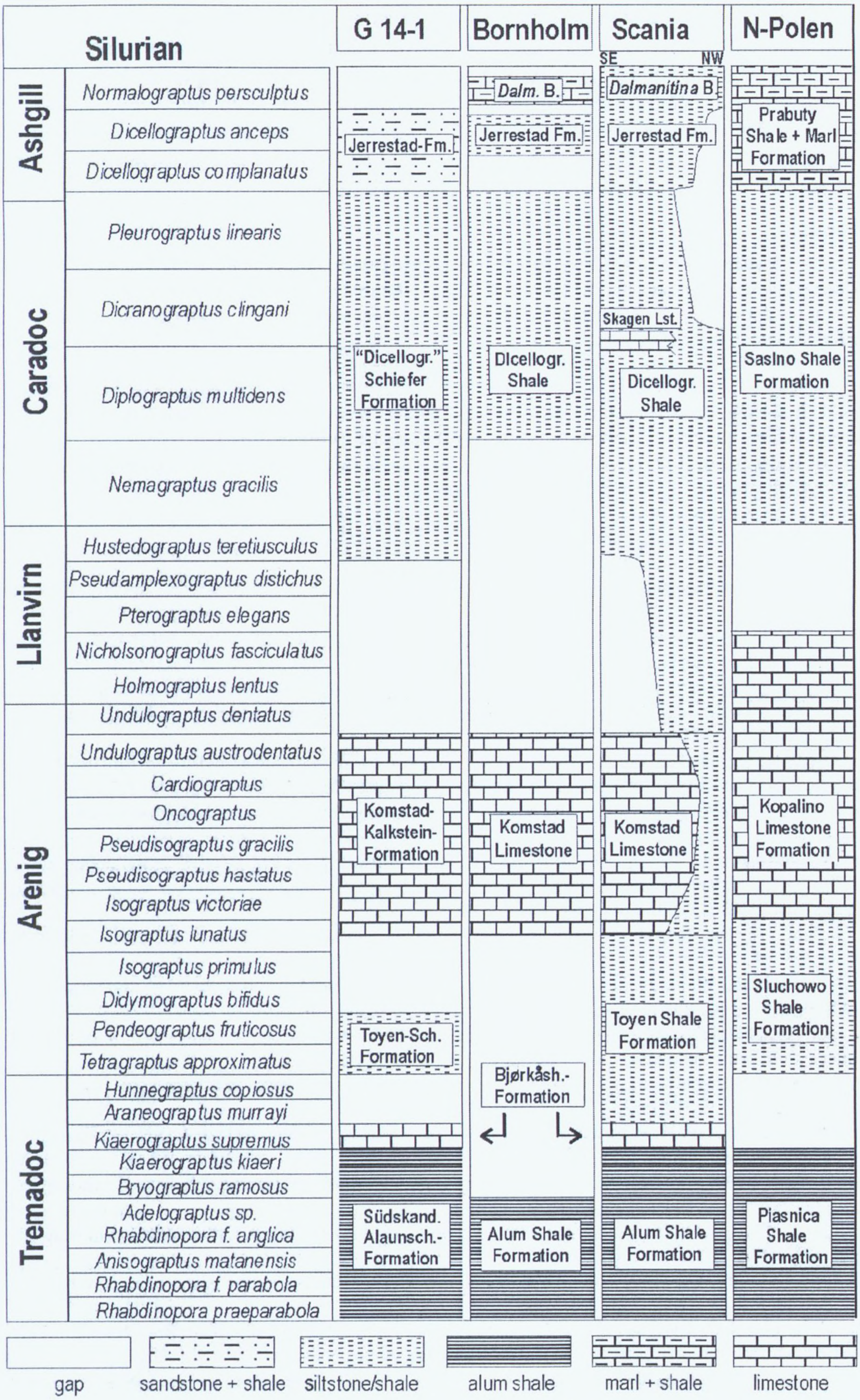

Fig. 2. Comparison of the lithostratigraphy of the Ordovician of the G 14-1 borehole, Bornholm, Scania and N–Poland. Stratigraphy after Beier et al., (2001a); Gravesen & Bjerreskov, (1984); Bergström et al., (1982); Lindholm, (1985); Modlinski & Szymanski, (1997).

sandstones with siltstones and mudstones of the Jerrestad–Formation (Ashgillian). The Silurian is represented by a succession of more than 300 m of mudstones, siltstones and rarely sandstones, the "Rastrites"–Schiefer–Formation (Llandoverian). According to the biostratigraphic investigations more than 240 m of these sediments belong to the *Monograptus gemmatus* subzone of the *Spirograptus guerichi* biozone. This contrasts to the area of Bornholm, where the whole *Spirograptus guerichi* biozone is only 12 m thick (Maletz, 1997). The Early Palaeozoic succession of the borehole G 14-1 ends in the middle Llandoverian with the "Rastrites"–Schiefer–Formation. Triassic rocks discordantly overlay the succession.

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THERMOLUMINESCENCE AGE OF MIDDLE AND LATE PLEISTOCENE DEPOSITS OF VILKIŠKĖS EXPOSURE IN EASTERN LITHUANIA

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A different conclusions were made on the age and origin of a Pleistocene deposits in Vilkiškės outcrop [1, 7, 8] and generally in the Neris River valley between Belarus border and Vilnius city. The present study focused in Vilkiškės exposure (latitude 25°22'35" N and longitude 54°50'16" E). The objective of our recent investigation is the dating by thermoluminescence method (TL) deposits of Vilkiškės section. The twenty samples of sandy deposits were dated by St. Fedorowicz in laboratory of Gdansk University [5]. TL dates are given in Fig. 1. The Vilkiškės section comprises some different chronostratigraphical complexes and sedimentational phases of two glacial-interglacial cycles. On base of TL datings of Middle and Late Pleistocene sediments, a fine grained sands of aquatic origin are attributed to the Snaigupėlė (Drenthian–Warthian) Interglacial of Middle Pleistocene in lowermost part of Vilkiškės exposure. TL dates of sediments in uppermost part of the section founded the base to determine the ages of Merkinė (Eemian) Interglacial and Nemunas (Weichselian, Vistulian, Valdaian) glacial of Late Pleistocene.

TL method. The annual radiation doses (Dr) are determined by taking the potassium, radium and thorium counts with an gamma spektrometer. A preliminary procedure preceded determination of the equivalent dose (ED) and involved the removal of the external shells of quartz grains (grain size 88–102 micrometer – samples no: 5, 6, 13 and grain size 40–60 micrometer – samples no: 1–4, 7–12, 14–20).

The purified fraction was then treated with 10% HCl for 60 minutes. ED was assessed using the reproduction method. The fraction under test was exposed to UV radiation for 24 hours, after which the residual TL level was measured. The sample was then exposed to radiation from a Co – 60 cobalt bomb of such an intensity that values of the TL induced by this dose would between the residual TL and the natural TL levels. The increase in TL was found to be linearly dependent on the size of the dose and ED was then obtained by extrapolation. The occurrence of TL saturation was noted. The accuracy of dating, taking into account laboratory factors, was estimated at around 15% of the assigned age values.

Snaigupėlė Interglacial. In the lower part (23.9–37.1 m from top) of the Vilkiškės outcrop a lacustrine fine grained sands with lamination and interlayers of silt are attributed by us to Snaigupėlė Interglacial. This part of section comprises two chronostratigraphical units: lowermost – 253.0±38.0 – 280.6±420 ka BP and uppermost – 182.3±28.0 – 192.0±28.8 ka BP. They belong to Lower and Upper Snaigupėlė Interglacial units correspondingly (Fig. 1). Two sublithocomplexes (VI^a and VI^c) were distinqueshed based on the results of structural, grain size and geochemical data [1]. Greenish yellow sand with silt laminas was accumulated in the lacustrine basin at the beginning of the Snaigupėlė Interglacial.

White and yellowish light grey, very fine grained, massive, horizontally and wavy laminated sand with calcareous tuff in separate laminas belongs to the main part of Snaigupėlė Interglacial. Sandy sediments contain a highest quantity of quartz.

The optically stimulated luminescence (OSL) dating of lacustrine sediments of lower unit were made in 2000 by Galina Hütt in the Institute of Geology Tallinn [3]. The analogical age (250,000±17,400 and > 250,000 years) to three samples was determined. Similar TL age (> 250 ka BP and 175±18 ka years BP) of the Snaigupėlė Interglacial lake sands at the Antaviliai outcrop near Vilkiškės site have been published by J. Satkūnas and G. Hütt [8].

In the studied Vilkiškės outcrop the Snaigupėlė Interglacial sediments are not overlain with Medininkai (Warthian) Glaciation till. This till was affected by denudation later. The Medininkai (Warthian) till has been noted at a other outcrops of Neris River between Vilnius city and Belarus border during earlier investigations by different researchers [1, 6] and us too.

Merkinė Interglacial. One sample of fine-grained sand with organics is dated TL method at 103.0±15.0 ka BP. It belongs to Merkinė (Eemian) Interglacial. Fine structure and horizontal texture suggest a lacustrine sedimentation.

Nemunas Glacial. The TL dating was done for twelwe samples of deposits of the Nemunas (Weichselian, Vistulian, Valdaian) silty sands, with frozen involutions, which formed about 66.6±10.0 – 54.8±8.2 ka BP. The second unit of the laminated lacustrine sand formed about 38.0±5.7 – 32.4±4.6 ka BP corresponds to the Middle Nemunas (Vistulian, Weichselian, Valdaian) time. The third unit of glacial deposits (tills and glaciofluvial sediments with cover of glaciolacustrine and aeolian sands and interlayer of soil were formed in the Late Nemunas Nemunas (Vistulian, Weichselian, Valdaian) time about 18.3±2.7 – 11.0±2.7 ka BP. The two redih brown (lower) and brown (upper) till layers of this third unit in the Vilkiškės section are younger than 30,000 BP and older than 15,900 BP according TL dating (Fig. 1). Sand under soil layer (0.16 m thick) was TL dated at 15.9±2.4 ka BP.

The soil organics were dated by ^{14}C [4]. The radiocarbon age of soil layer is younger the sand laying below. The underlaying tills at the Vilkiškės outcrop may be correlated with Grūda (Brandenburgian, Lesznian) stadial and Žiogeliai (Frankfurtian, Poznanian) phasial of maximum of Late Nemunas (Vistulian, Weichselian, Valdaian) glaciation.

Glaciolacustrine and aeolion sand of upper part of the Late Nemunas (Vistulian, Weichselian, Valdaian) time were TL dated at 14,000±2,100 – 11,000±1,600 BP. Fine grained sand of massive structure in the uppermost part and subhorizontally laminated in the lowermost part compose the dated bed of sediments. The tickness of sand laminas is 10–20 cm. The sediments are brownish yellow colour.Their deposition took place in the periglacial conditions of maximum and retreat of glacial cover of Baltija (Pomeranian, Pomorzian) stadial of Late Nemunas time (Fig. 1). This uppermost part of the section belongs to Middle and Late Dryas time (Gotiglacial). Regular succession of Gotiglacial sedimentation changes was conditioned by recession of ice sheet in Latvia and Estonia [2]. The uppermost part of the section is composed of aeolian sand. The sand is covered by 30 m of recent forest soil cover.

TL dating of samples from the studied Vilkiškės section suggests that there are three chronostratigraphical units of Nemunas (Vistulian, Weichselian, Valdaian) glacial: Early, Middle and Late. The lithostratigraphical data and TL dates obtained during recent research confirm the absence of an ice sheet in Early Nemunas time. The Early Nemunas deposits are presented by periglacial silty and sandy sediments. Periglacial conditions existed in Early Nemunas time at Vilkiškės area. The rhythmic sedimentation of yellowish light grey and greenish yellow very fine sand with calcareous and organic rich laminas express the climatic fluctuations of megainterstadial of Middle Nemunas time. The glacier covered the South-eastern Lithuania in Late Nemunas time for limited number of thousand years. The extension of the Late Nemunas glaciation was short, less than 10,000 years. The studied area remained ice-free during Early and Middle Nemunas.

It is very important to take in account the results of TL method of dating, obtained recently for the Middle and Late Pleistocene sediments of the outcrop Vilkiškės. TL dates indicated the age of main climatoliths and thermomers of Middle and Late Pleistocene time in Vilkiškės outcrop. Climatostratigraphic criteria is applied for Quaternary stratigraphic subdivision [9]. Deposits formed during global warm (interglacials) or global cold (glaciation) time are attributed to the climatoliths. Deposits formed under warmer interval of time of glacial are attributed to the thermomer. Deposits formed during a colder period of glacial are named as cryomer. These terms (thermomer and cryomer) are used in our case, because palaeogeographical conditions of sediment formation are enough unclear.

To determine the rank of the climatostratigraphical unit are used the chronostratigraphical (or geochronometric) methods. To determine the age of climatoliths, thermomers and cryomers we used the TL method of geochronometric investigations.

The palaeontological, bioindicational, lithological, sedimentological and etc. methods can be applied for determination of climatostratigraphical units and changes of palaeoclimate.

The first thermomer in the Vilkiškės outcrop represents the Early Snaigupėlė Interglacial or climatolith (interval 98.0–107.5 m a. s. l.). TL dates of the sand from this interval cover the time from 280.6 ka years BP to 253.0 ka years BP. The second thermomer belongs to the Late Snaigupėlė Interglacial or climatolith (interval 107.5–113.9 m. a. s. l.). The sandy sediments of second thermomer of the Snaigupėlė climatolith have showed the TL age from 182.3 ka to 192.0 ka years BP.

The third thermomer belongs to the Merkinė (Eemian) Interglacial (climatolith). It is represented by silt and very fine-grained sand with admixture of organic. The layer of the Merkinė (Eemian) Interglacial is composed by white, yellowish light grey and greenish, horizontally laminated lake-bog sediments with lamina 2–30 mm thick.

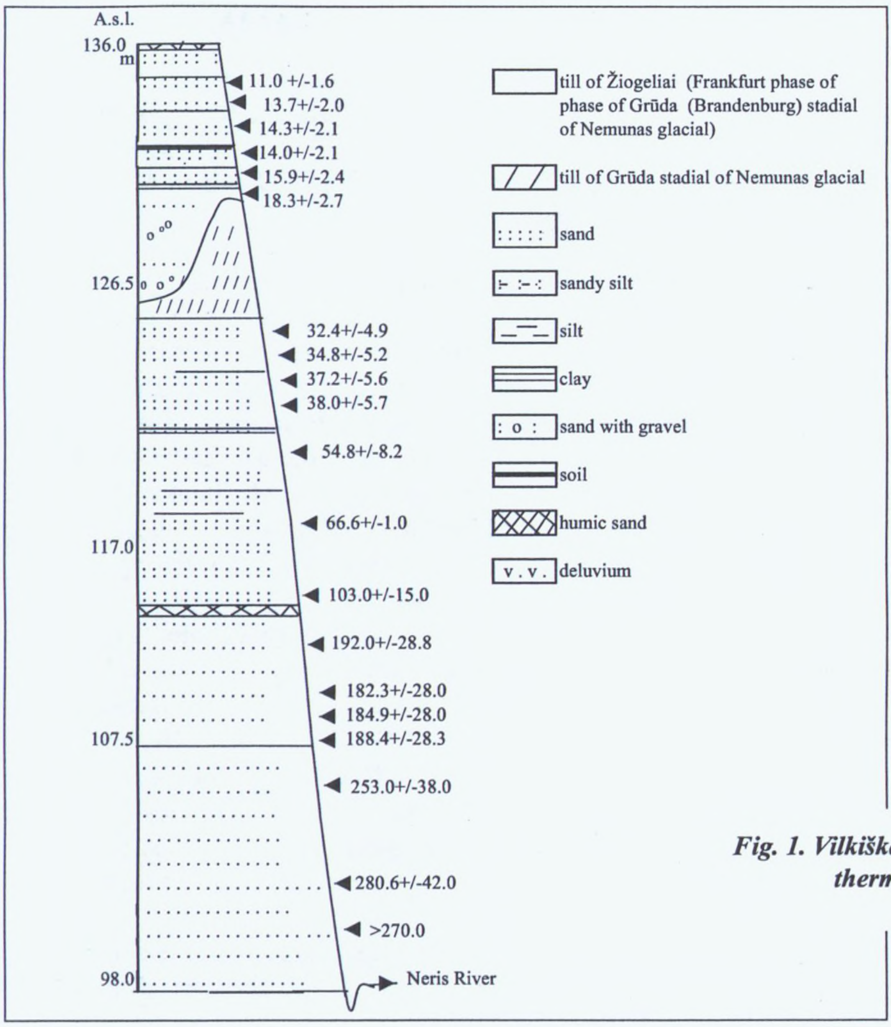

Fig. 1. Vilkiškės outcrop with results of thermoluminescence dating.

Silt in the uppermost part of layer is coloured by limonite pigmentation. The sediments in middle part of the layer are rich in mica. The admixture of lacustrine calcareous tuff are frequent in the lower band of the layer. The age of white very fine grained sand from middle part of the layer has been determined to 103.0±15.0 ka before present.

The fourth thermomer is connected with Middle Nemunas (Vistulian, Weichselian, Valdaian) megainterstadial. The Middle Nemunas megainterstadial (Rokai and Biržai) deposits accumulated about 66.6–32.4 ka years ago. The sediments of fifth thermomer have the TL age from 15.9±2.7 ka to 11.0±1.6 ka years BP. Their deposition took place in Late Glacial and final deglaciation of East Baltic region.

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RADIOCARBON AGE OF LAST GLACIATION TILLS IN LITHUANIA

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The location of the limit of the last glaciation depends essentially on the age of glacial deposits forming the topography in the extraglacial area as well as on the age of glacigenic sediments in sections of the glaciated area. Radiocarbon dating is of exlusive importance for substantiating the age of the sediments under and above the tills of the Late Nemunas (Vistulian, Weichselian, Valdaian) glaciation in Lithuania. Radiocarbon dating of the limits of the maximum advance of the Nemunas (Vistulian, Weichselian, Valdaian) continental ice in South-East Lithuania has been made earlier [5]. The limits of the maximum spread of the last ice cover in South-East Lithuania should be located in maps on the northern slopes of the Medininkai Higland. The till of Medininkai (Wartanian) glaciation forms the topography of Medininkai Highland in South-East Lithuania. Evidently, the ice sheet did not penetrate into the South-Eastern Lithuanian region during the Early and Middle Nemunas (Vistulian, Weichselian, Valdaian) interval of time.

The data already obtained from the study of the Medininkai section (borehole 117A) located in South-East Lithuania confirm the absence of ice in this region during the Early, Middle and Late Nemunas (Vistulian, Weichselian, Valdaian) [5]. The glaciolacustrine silt and clay that covered lacustrine and peat sediments were probably accumulated during the interval from 29,800 to 23,770 years BP. Their deposition probably took place just before the last glaciation. The deposits of the Nemunas glacial maximum are not of glacial origin and are presented by periglacial sediments – glaciolacustrine silt and clay (23,000–15,000 years BP). Their deposition took place during the maximum of the last glaciation of Lithuania. This silt and clay was formed in aquatic conditions under the influence of melting ice water of Late Nemunas (Grūda and Baltija stadials).

Data of the radiocarbon dating of last glaciation tills in the geological section are listed in Table 1. In different parts of Lithuania 4 sections were studied: Biržai (northern), Rokai, Balbieriškis and Girininkai (middle), Jonionys (southern) and Vilkiškės (eastern).

In the uppermost part of sandy lacustrine-alluvial strata of the Rokai section near the Kaunas city, Central Lithuania, 1.5 m below the Late Nemunas till, there lies an interlayer of sandy peat and chemogenic white carbonatic tuff. Samples of humic loam and terrestrian peat were taken from the depth of 1.50–1.60 m under the till and dated at the Radiocarbon Laboratory of Silesian Technology University and St. Petersburg University (Table 1). The upper samples taken from the depth of 1.55–1.59 m were dated to 34,930±510 (LU-3155) and 24,430±40 (Gd-6991) and the lower samples from the depth of 1.6 m to 37,590±820 (Lu-3165) and 27,800±340 (Gd-7436).

The interstadial Middle Nemunas (Middle Vistulian, Middle Weichselian, Middle Würmian) deposits in the Rokai outcrop are overlain by Upper Nemunas glacial deposits – tills forming strata up to 18–20 m thick. In the section of overlapping till beds of the maximum phase of the Grūda (Brandenburgian) stadial and of the recessing Žiogeliai (Frankfurtian) phase, the East-Lithuanian and South-Lithuanian phases of the Baltija stadial of the last glaciation are distinguished. Lithologically these are respresented by variegated boulder loams. The uppermost part of this boulder loam of the Žiogeliai (Frankfurtian) phase of the Grūda (Brandenburgian) stadial is reworked by basin waters. Tills of the Grūda (Brandenburgian) stage are separted from East-Lithuanian and South-Lithuanian phasial tills by a thin (only 0.6 m thick) bed of lacustrine yellowish-grey cross-bedded silt formed during the Grūda–Baltija Interstadial. Infrequent spores and pollen are found in this silt. Spore-and-pollen spectra are very dispersed and poor.

They should be studied in a more representative section of these deposits. A till stratum of the Upper Nemunas subformation of the Upper Pleistocene is overlapped by varved clays deposited in a periglacial basin at the edge of the glacier of the Middle-Lithuanian phase of the Baltija stadial. The boundary of glacier distribution of this phase was in the northern outskirts of the Kaunas city.

The Merkinė (Eemian, Mikulian) Interglacial deposits at the Jonionys stratotype locality, South Lithuania, are stratigraphically covered by tills of the last (Nemunas = Vistulian = Weichselian = Valdaian) glaciation [7].

Table 1. Radiocarbon dates of organics and carbonates from sediments under and above the tills of Last glaciation in Lithuania.

Site	Material	Laboratory number	^{14}C Age [BP]	References
under the tills				
Jonionys	Wood	Gd-10825	31,500+2300/-1800	Gaigalas, Pazdur, Pawlyta, 2001
Jonionys	Wood	Gd-14000	> 31,000	Gaigalas, Pazdur, Pawlyta, 2001
Jonionys	Peat	Vs-914	14,040±240	Gaigalas, Hütt, 1995
Rokai	Peat	Gd-6991	24,430±280	Gaigalas, 2000
Rokai	Carbonatic tuff	Gd-7436	27,800±340	Gaigalas, 2000
Rokai	Peat	LU-3155	34,910±510	Gaigalas, 2000
Rokai	Carbonatic tuff	LU-3165	37,590±830	Gaigalas, 2000
Biržai	Wood	LU-1633	33,600±1060	Gaigalas et al., 1992
Biržai	Wood	Vs-412	34,440±1500	Gaigalas et al., 1992
above the tills				
Vilkiškės 1	Soil	Gd-15327	10,815±160	RD, NaOH-sol
Vilkiškės 2	Soil	Gd-15392	7690±160	RD, NaOH-sol
Vilkiškės 3	Soil	Gd-12405	10,320±100	RD, NaOH-sol
Vilkiškės 4	Soil	Gd-1812	30±190	RD, whole org.
Balbieriškis 7C	Carbonate	Gd-16094	26,630±1680	RD, $\delta^{13}C$ = -1.80 ‰
Girininkai 5C	Carbonate	Gd-18011	> 26,000	RD, $\delta^{13}C$ = (-1.80 ‰)

RD = recent dating, ( ) = assumed value, NaOH-sol = fraction of soil dissolved in NaOH (humic acids), whole org. = whole organic matter

The Nemunas ice-free deposits in the Jonionys section overlie the sediments of Merkinė Interglacial. Glacigenic sediments in the Jonionys section were found in the upper part. For peat and wood remnants under the rewashed till two radiocarbon dates were obtained: Gd – 10825: 31,500+2300/-1800 and Gd – 14000: > 31,000 BP (Table 1). Its deposition probably took place just before the maximum of the last glaciation. In the Jonionys section of the Merkinė Interglacial, Early and Middle Nemunas non-glacial sediments were probably accumulated during climatic fluctuations, but without glacigenic sedimentation. The glacigenic sediments in the uppermost part of the Jonionys stratotype section are younger than 30,000 BP and belong to the Late Nemunas glacial maximum.

Lacustrine sediments, 13 m thick, have been discovered under the Late Nemunas (Vistulian, Weichselian, Valdaian) tills in a palaeokarst depression at the Biržai town in North Lithuania [9]. Radiocarbon datings of wood fragments from silt lenses obtained in two different laboratories showed: 34,400±1500 (Vs-412) and 33,460±1060 years BP (LU-1633) (Table 1). The sediments belonging to the upper till complex were formed in a lake attached to karst depressions. The tills belong to South and Middle Lithuanian phases of the Baltija stadial of Nemunas glaciation. This till forms a moraine plain with a flat surface. The tills that covered non- glacial sediments of the Nemunas (Vistulian, Weichselian, Valdaian) glacial were accumulated in Lithuania later than 37,590±830 – 24,430±280 BP. The nonglacial sediments below the tills in the sections studied are older than the Late Nemunas glacial maximum in South Lithuania.

In Lithuania, lithological research led to the recognition of two independent Late Pleistocene till horizons, which were designated as the Grūda (Brandenburgian) and Baltija (Pomeranian) glacial stadials [1].

The petrographical composition of the till of the Grūda stadial of the Nemunas glaciation indicates a Central Swedish source for erratic boulders and coarse material [2]. The Grūda till was enriched with Mesozoic sedimentary rocks. Apart from Mesozoic marl clasts, this till is rich in crystalline rock from

Central Sweden, the Åland Islands and the Baltic Sea floor. The Grūda ice sheet moved therefore from northwest to southeast. During retreat a minor readvance, the Žiogeliai phasial, occurred. In the Žiogeliai till, crystalline rocks (porphyres, mandelstones, diabases, etc.) from the northern part of the Middle Baltic Sea predominate, accompanied by numerous fragments of Yotnian sandstones.

The Baltija till deposited by the stadial glacial advance exhibits high frequencies of dolostones derived from Devonian rocks of the East Baltic region. The direction of this ice movement varied regionally and was apparently controlled by the topography. The inland ice of the Baltija stadial transgressed the area of Lithuania in three distinct lobes: West-Lithuanian, Central-Lithuanian and East-Lithuanian. Glacial deposits of each of the three lobes differ in the petrographic composition of the erratics of crystalline rocks and clasts of sedimentary rocks. However, the bulk of the ice moved from north to south over East Baltic Palaeozoic rocks. The Baltija stadial till has a specific association of Palaeozoic sedimentary rocks and crystalline rocks from South Finland.

While retreating and recessing, the Baltija stadial glacier left its phasial tills, which are distributed by tracks of East-Lithuanian, South-Lithuanian, Middle-Lithuanian and North-Lithuanian phasial tills on the surface of Lithuania. The phasial tills are spread locally, in recession zones of phasial glaciers, so their full lithostratigraphical sequence may be revealed only by a successive investigation and correlation of Baltija tills on the whole area of Lithuania [6].

The maximum of the Baltija stadial is called the East-Lithuanian phasial. In the next zone of the South-Lithuanian recessional phase, three oscillational branches of end moraines are found. Immediately after each retreat of the ice sheet, vast areas of Lithuania in interphasials were covered with big ice-dammed lakes. The peculiar conditions came to dominate in this area. At present in Lithuania some basins of glaciolacustrine clay of dammed periglacial lakes are established. Water of periglacial lakes was dammed near the edge of a recessing glacier every time. Varvometric measurements of varved clays from these dammed periglacial lakes enable to determine the geochronology of glacier recessions, varved clay sequence and peculiarities of glaciolacustrine sedimentation [11]. Geochronometric counts of varved clay bands of phasials, oscillations and facial complexes, separated by stratigraphical boundaries of different categories, allow us to evaluate the duration of recession and oscillations, progressive, stable and regressive states of of Baltija stadial glacier.

Recent radiocarbon datings have been done for sample organics from the soil overlying the tills of Late Nemunas (Vistulian, Weichselian, Valdaian) time in the Vilkiškės outcrop (Table 1). The Vilkiškės site is located in near the northern suburb of the Vilnius city (Eastern Lithuania). The dated exposure is found in the peripherial part of last glaciation.

Two reddish brown (lower) and brown (upper) till layers are found to occur. The overlying soil layer (0.16 m thick) was dated by ^{14}C to 10,815±160 years BP (Gd-15327) and 7,690±160 (Gd-15392). The till strata at the Vilkiškės section belongs to the Grūda (Brandenburgian, Lesznian) stadial and the Žiogeliai (Frankfurtian, Poznanian) phasial of the Late Nemunas glaciation maximum. These tills are overlain by glaciolacustrine and aeolian sediments. Their deposition took place in the periglacial conditions of the glacier cover maximum and retreat during Late Nemunas Baltija (Pomeranian, Pomorzian) stadial. Thermoluminescence dating of sandy deposits by St. Fedorowicz at Gdansk University Laboratory under and above the tills served as the base for determining its age [8]. Under the soil sand deposits were TL dated to 15.9 ka.

Thus, we can firmly establish that the covering till strata of non-glacigenic sediments of Nemunas glacial in Lithuania belongs to the Late Nemunas (= Vistulian = Weichselian = Valdaian) time (younger than 24,000 years). Thus, in Lithuania the stratum of Upper Pleistocene Nemunas glaciation tills is heterogeneous. Both phasial and stadial tills are found. Glacial sediments of Early and Middle Nemunas (Vistulian = Weichselian = Valdaian) time have not yet been detected in Lithuania.

The duration of the Late Nemunas glaciation was short, certainly less than 10,000 years. The recessing of the Baltija stadial glacier from the main area of Lithuania took about 3000 years. In the Late Nemunas (Vistulian, Weichselian, Valdaian), the main glacial advance, which covered the most part of Lithuania, also lasted 3000 years.

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STRATIGRAPHY OF LITHUANIA IN THE LAST DECADE: A RETROSPECTIVE VIEW

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Lithuania, as other countries surrounding the Baltic Sea, belongs to the oldest pre-Cambrian East-European Craton. Slow, differentiated tectonic movements prevailed here during the whole Phanerozoic since Late Proterozoic Riphean–Vendian time. The sedimentary cover was formed; uplifting and marine basins replaced by land changed periods of sinking. As everywhere, fossil fauna and flora remnants found in the sediments of former marine basins are a main source of information about the Earth's history events in this area.

History of stratigraphy in Lithuania goes back more than 220 years, when famous botanist Jean Emmanuel Gillibert, professor of the Principal School of Grand Duke of Lithuania, published the first geological observation in 1780s. In 1803, a year after Alcide d'Orbigny and Ignacy Domeyko Ancuta were born, a chair of Mineralogy was established in the oldest Vilnius University. In remarkable 1830, zoologist and palaeontologist Eduard Eichwald, professor of Vilnius University, discovered ammonites in Jurassic outcrops in the vicinity of Papilė, northwest Lithuania, and naturalist Frederick Dubois de Montpereux, governor of nobleman Baron von Ropp family in Pakruojis, compiled a first "geognostic" map of Lithuania.

The Jurassic system was the first one to be found in the area according to ammonite determinations made by Eduard Eichwald, Leopold von Buch and George Pusch. Devonian outcrops were described from north Lithuania during the notable expedition leaded by Roderick I. Murchison and Eduard de Verneuil to European Russia in 1841.

Later on, in the second half of 19th century, Constantine Grewingk, professor of Dorpat (Tartu) University, was the first who dated the Permian (in 1857) and Triassic (1878) sedimentary rocks in the northern part of the area. The Cretaceous, Palaeogene and Neogene outcrops and faunal remnants in erratic boulders were found in South Lithuania and neighbouring Eastern Prussia (Ostpreussen) by C. Grewingk, Anton Giedroyc, chief geologist of the Russian Geological Committee, and Gustav Berendt and Alfred Jentzsch, scientists of the Königsberg University. In 1880–1920s Jozef Siemiradzki, V. N. Rjabinin, Roland Brückmann, F. Krenkel, K. Boden, Roland Brinkmann published the first palaeontological descriptions of well-known Jurassic fossils from Papilė. Notably, that R. Brückmann's edition of Lithuanian–Kurish Jurassic foraminifera (1904) was the first application of micropalaeontological methods in Lithuania. Moreover, in this period the first wells for underground water supply were drilled. The borings brought new data on distribution of pre-Quaternary rocks of different geological age.

The data obtained during the 19th century were for the first time systematised by Juozas Dalinkevičius, Professor of Geology and Mineralogy in Kaunas University, in the 1930 and 1940s. Devonian, Late Permian, Jurassic, Mid Cretaceous fossils of various groups being studied in this period. In addition, the new boreholes drilled in Lithuania discovered thick Early Palaeozoic sedimentary series, i.e., Cambrian, Ordovician, and Silurian, beneath the Devonian, Permian and Cretaceous. Particularly important was an idea, concluded by J. Dalinkevičius in 1940s, about completeness of geological sequence in the whole Baltic Syneclise on the southwestern border of the East-European Craton, and about oil prospecting in the Lower Palaeozoic sediments.

The fruitful idea of oil prospecting of the Lower Palaeozoic has stimulated a lot of geological works in Lithuania after 1945, the Second World War. Particularly, many geological mapping and deep oil prospecting boreholes were drilled during a Soviet period (1945–1991). As it was recognised, the basement of pre-Cambrian crystalline rocks in Lithuania, in the Baltic Syneclise basin, lies at the depths of 600–2000 m, deepening southwestwards to 2900–3000 m in Kaliningrad and to 4000–6000 m in north Poland. Hence, the total thickness of Vendian–Phanerozoic sequence exceeds 3,000 m in the area.

A highly rich geological and palaeontological material was brought from boreholes for scientific research. This situation allowed to develop detail biostratigraphy of all the geological systems, and many available methods of palaeontology, micropalaeontology and palaeobotany were used. The results achieved could be difficult to imagine.

Therefore, in the second half of 20th century, the stratigraphy of Lithuania made major step forward. A detailed stratigraphic subdivision of all of the geological systems was elaborated, based on extremely well preserved fossils from deep boreholes. Around 100 monographs were devoted to palaeontology and biostratigraphy. Detailed stratigraphic charts were compiled, based mainly on theories of zonal stratigraphy, ecostratigraphy and basin analysis. A Lithuanian Stratigraphic Guide has now been compiled, and published in 2002.

In a long run of the geological time, the sedimentary sequence is stratigraphically most complete in a whole East European Craton. In spite of rather intense changes in palaeogeographic conditions, the sedimentary cover is completed by a broad spectrum of marine, lagoonal and continental deposits. The international (global), regional and local stratigraphic units (stratons) are determined in the stratigraphic charts of every geological system. Regional stages are subdivided into the units depending on methods used and purposes of an end-user:

- Lithostratigraphic units – groups, formations, members, beds;
- Biostratigraphic units – zones (chronozones);
- Sequence stratigraphy units;
- Magnetostratigraphy units;
- Climatostratigraphy units (Quaternary);
- Crystalline rock units – complexes, groups, formations.

A relative age of smaller stratons is most often determined by methods of biochronology, i.e. relative stratigraphy. The rules of a rather complicated stratigraphic classification and terminology are followed after recommendations of the International Stratigraphic Guide (Salvador, Ed. 1994) and Lithuanian Stratigraphic Guide (Grigelis, Ed. 2002).

As it was mentioned before, Professor Juozas Dalinkevičius elaborated the background of Lithuanian stratigraphy in 1928–1960. In the next decades (1960–1990) the main fundamental investigations were made by T. Jankauskas (Vendian–Cambrian), J. Paškevičius (Ordovician–Silurian), V. Karatajūtė-Talimaa (Lower–Middle Devonian), S. Žeiba (Devonian–Carboniferous), P. Suveizdis (Permian), J. Kisnėrius (Triassic), L. Rotkytė (Jurassic), A. Grigelis (Jurassic, Cretaceous, Paleogene).

In two latest decades (1980-2000) the investigations made by L. Paškevičienė, T. Jankauskas, N. Sidaravičienė, A. Brazauskas, E. Laškovas, P. Lapinskas, P. Musteikis, V. Narbutas, Val. Katinas, P. Šimkevičius, R. Mertinienė, Vl. Katinas, V. Baltakis et al., enabled to establish the very detailed stratigraphy and on this basis, to develop the basin models and palaeogeographic reconstructions of all the geological periods of Phanerozoic in Lithuania.

The last decade (1990–2000) is especially notable for publication of many abstracts of conferences, magazine articles and several monographs. They are listed below as a supplement. Some IGCP international and bilateral projects were fulfilled. Several scientists – stratigraphers – were awarded the national prizes:

Jurgis Kisnėrius	*Academician Juozas Dalinkevičius Prize, 1993 (after death)*	*Achievements in Triassic stratigraphy and compiling of geological maps*
Algimantas Grigelis, Vytautas Narbutas, Valentinas Kadūnas, Juozas Paškevičius, Povilas Suveizdis	*Lithuanian State Science Award, 1996*	*Monograph "Geology of Lithuania"*
Valentas Katinas	*Young research fellow award by Lithuanian Academy of Sciences, 1997*	*Palaeomagnetic research of Lower Triassic in West Lithuania's boreholes*
Juozas Paškevičius	*Academician Juozas Dalinkevičius Prize, 1997*	*Achievements in Silurian Stratigraphy and graptolite studies*
Algimantas Grigelis	*Academician Juozas Dalinkevičius Prize, 2001*	*Achievements in Mesozoic Stratigraphy and micropalaeontological studies*

A particular mention should be made about a basic work done by N. Sidaravičienė "Stratigraphic Units of Lithuania" (1999) and the newest "Lithuanian Stratigraphic Guide" (completed by A. Grigelis, 2002). These books should be in use by every geologist, student, teacher, and decision maker.

The completeness of publications on stratigraphy in general is shown on Fig. 1.

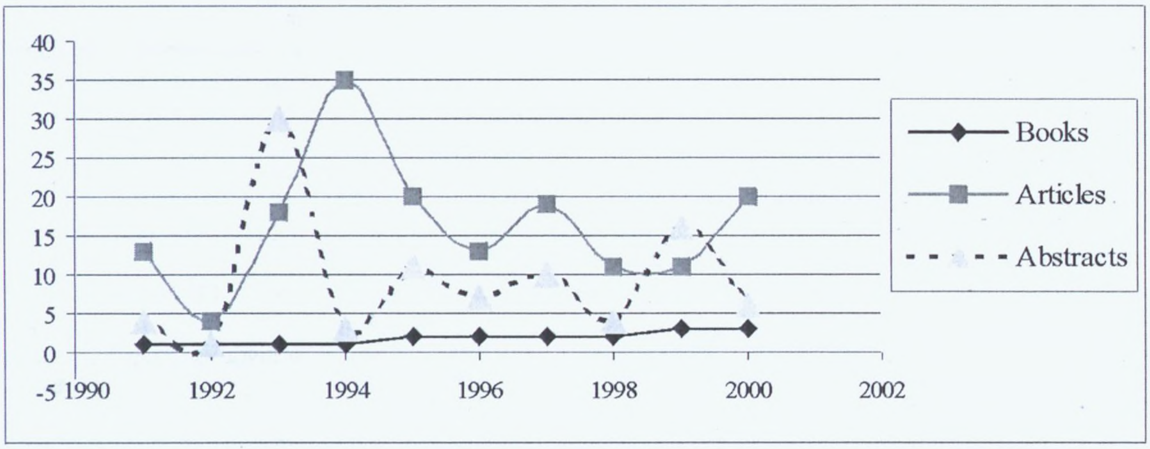

Fig. 1. Number of publications on Lithuanian stratigraphy in 1990–2000.

Concluding, it could be said that modern Lithuanian Phanerozoic stratigraphy has used main possibilities offered by this method, viz., subdivision of sedimentary rock sequences of different geological age is done in such a detailed manner that is possible to be reached by palaeontological and lithological investigations. Elaboration of the detailed stratigraphic charts relies mainly upon the concept of zonal stratigraphy, which is based on detailed research of fossil remnants. Otherwise, not every geological sequence contains enough fossil findings. Taking it into account, peculiarities and common relations of sedimentary rocks of different facies and different genesis (marine, lagoonal, continental) have been studied, and ecostratigraphic, faunal community, lithostratigraphic, palaeoecological and – at least – basin analysis methods have been used.

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MARGINAL POSITIONS OF THE LATE NEMUNAS (LATE WEICHSELIAN) GLACIATION IN LITHUANIA

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New information appeared in Quaternary geological and geomorphologic maps at a scale 1:200,000 that were compiled after the revision of surface landforms in Lithuania. As a result, the geomorphologic image of Lithuanian surface, obtained from aerial photos, differs somewhat from the traditional model of deglaciation in Lithuania, in which Grūda (Brandenburg), Žiogeliai (Frankfurt) and Baltija (Pomerania) stages as well as Eastern Lithuanian (Aukštaičiai), South Lithuanian, Middle Lithuanian and North Lithuanian phases of the Baltija stage were distinguished (Gaigalas, 1995).

Summarizing all the new data on the surface geology and geomorphology of Lithuania, presented in the new digital Quaternary geologcal and geomorphological maps at a scale 1:200,000, following conclusions may be drown (Figure).

A relief of the Asmena Highland including the Lyda plateau, Medininkai and Buivydžiai heights, have been formed during the pre-Last (Medininkai) Glaciation. According to the palinologic evidences and radiocarbon (^{14}C) dating the Middle Pleistocene age was confirmed for the Medininkai height.

The maximum extension of the Last Glaciation in Vilnius area has been traced to the foot of the Medininkai height, in the Vilnia depression, while in southern Lithuania maximal outspread of the Last Glaciation is probably outlined by Urstrom valley Vokė–Merkys–Nemunas and does not reach foots of the Lyda plateau.

Maximum extent zone of the Last Glaciation in Lithuania corresponds to the Leszno–Pomeranian Stadial limit in Suwalky Lakeland, drown by T. Krzywicki (2001) and is described more detail in A. Ber's study (Ber, 2000).

According to many geomorphologic features Last Glacial Maximum in southern Lithuania is outlined from the south by the Merkys River (Figure). The previous boundary of the Last Glacial Maximum in earlier published maps of Belarus and Lithuania reflect only distal margins of outwash sediments of the Last Glaciation. The defined boundary of the Last Glacial Maximum in southeastern Lithuania fully coincides with the newly traced maximum extent limit of the Poozerie Glaciation in Belarus (Matveejev & Pavlovskaya, 2001).

The last (Late Nemunas) ice advance in Lithuania is considered to comprise two stages, a Grūda Stage (the older) and the Baltija Stage. Glaciomorphologic complexes of the Grūda Stage occupy southeastern slopes of the Baltija Highland, southwestern slopes of the Švenčionys Highland and the Southeastern Lowland. The major part of the present topography of Lithuania was formed during the the Baltija Stage.

Marginal moraine complexes, which may indicate the Maximal phase, Aukštaičiai phase and its East Aukštaičiai oscillation and Middle Lithuanian phase with the North Lithuanian and Pajūris oscillatory stops of the retreating of the Baltija ice sheet are determinated.

Maximum of the Baltija (BL) stage (Eastern Lithuanian Phase) has to be correlated with the Wygry (W) subphase and Pomeranian (Pm) phase according to T. Krzywicki (Ber, 2000). The correlation of the Pomeranian phase in Poland with the South Lithuanian phase in Lithuania (Uscinowicz, 1999; Marks, 2002 etc.) has no geological and geomorphologic background. A distinction of the South Lithuanian phase is problematic at all. Recession stage previously distinguished as the South Lithuanian Phase, is recommended to treat as oscillatory one.

Maximum limit of the Baltija stage, introduced as the Aukstaičiai Phase (A) in eastern Lithuania, would coincide with the Lepel phasial complex of the Late Poozerie stage of the Poozerie Glaciation in western Belarus (Matveejev & Pavlovskaya, 2001). The Braslav (in Belarus) and newly suggested Eastern Aukstaičiai (RA) glaciomorphological complexes have the same characteristic geomorphologic and geological features (Figure). However, the above correlation of the glaciomorphologic complexes in eastern Lithuania and western Belarus is possible only if Lithuanian and Latvian (Marks, 2002) ice streams of the Last Glaciation were asynchronous.

Middle Lithuanian Phase (VL). Overall, it is a prominent moraine ridge occupying an area between Seda, in the west, to Rokiškis, in the east. The ice-pushed ridge up to 8 km wide consists of small, low hills. Small outwash plains are related to the ridge near Šiauliai and Anykščiai. The Middle Lithuanian phase limit is drawn along the the distal slope of the ridge. The boundary, however, is less-clearly expressed to the northeast of Rokiškis and to the northwest of Seda, since the hilly ridge "merges" here with the adjacent hilly relief, or more exactly – as if splits off it. There are no classic marginal ridges, which could show existence

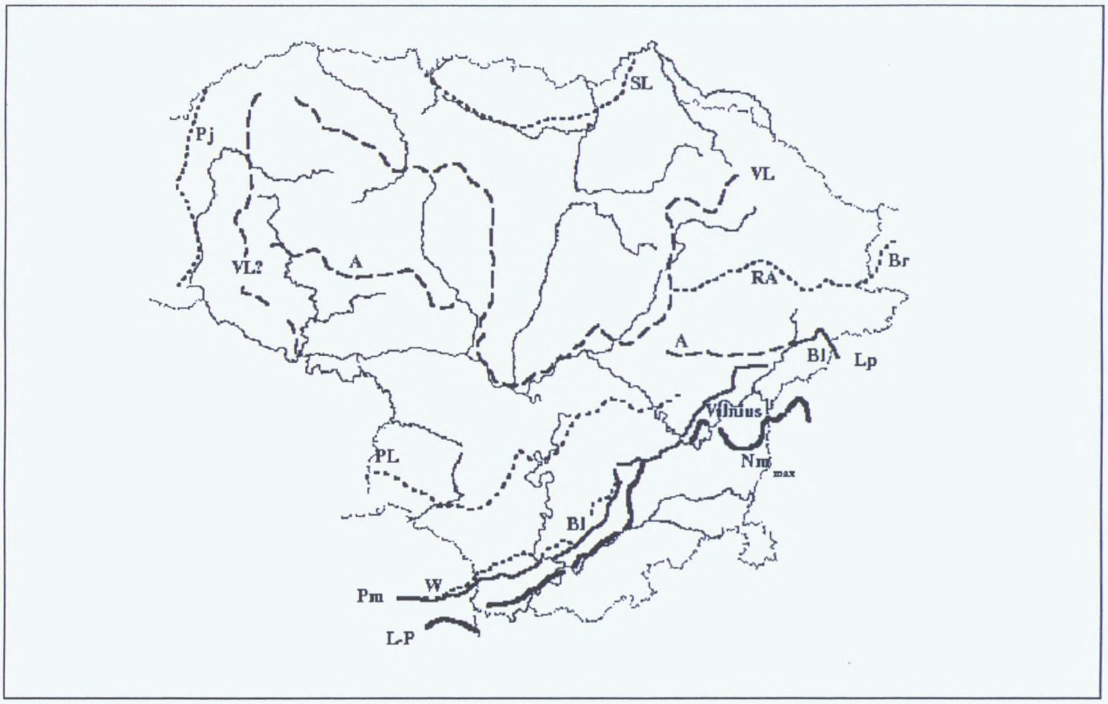

Figure. Ice sheet marginal positions in Lithuania: Nm max – limit of the Last Glacial Maximum; Bl – Baltija stage; PL – South Lithuanian oscillacion; A – Aukštaitija recession phase of the Baltija stage; RA – Eastern Aukštaitija – oscillation; VL – Middle Lithuanian Phase; SL – North Lithuanian phase (oscillation of the VL Phase); Pj – Pajūris phase?, (oscillation?). In Poland: Pm – Pomeranian Phase; W – Wygry subphase; L-P – Leszno–Pomeranian stadial after T. Krzywicki (2001). In Belarus: Br – Braslav and Lp – Lepel phases of the Late Poozerie Glaciation after I. Pavlovskaya (Matveejev & Pavlovskaya, 2001).

of ice margin at the western part of the Žemaitija Highland. VL phase limit here is outlined by north-south trending belt of loaf-shaped hills and by an end moraine fragment known as Vilkyškiai ridge to west of Jurbarkas.

North Lithuanian Oscillation (ŠL). The North Lithuanian Phase is represented by an arcuate ridge that bends towards the south, which is about 10 m high and 2 km wide. It is well expressed in the present lanscape and is known as the Linkuva Ridge. This ridge is considered to represent the limit of the North Lithuanian Phase in maps and publications. The whole ridge is in the Lithuanian area and only ends of this bow-shaped formation protrude in Latvia. General geological situation seems to testify that there was a local inflow of glacier's ice mass during the Middle Lithuanian phase ice advance only along the Middle part of the Žiemgala Lowland southwards as far as Linkuva ridge. Fragments of a kame terrace in the distal part of the ridge could be an evidence that Middle Lithuanian Lowland was not covered with ice, hence, the Linkuva ridge should be considered to be oscillatory formation of Middle Lithuanian phase ice sheet.

Pajūris Oscillation (Pj) can be synchronous with the North Lithuanian oscillatory advance (or even younger). The phase is represented by a subdued, low ridge along the Baltic Sea coast. The Pajūris oscillation boundary is drawn along the eastern margin of this end moraine ridge.

The ice sheets limits have been traced on the basis of geological and geomorphological criteria, however the precise stratigraphical rank of the established ice sheet retreat advances are still unknown.

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BIOSTRATIGRAPHIC EVIDENCE AND PALAEOGEOGRAPHIC IMPLICATIONS FOR THE LATE HOLOCENE BALTIC MYA-STAGE IN SEDIMENTS OF THE GREIFSWALDER BODDEN (NE-GERMANY, SOUTHERN BALTIC SEA)

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The Greifswalder Bodden, one of the numerous bays bordering the southern shore of the Baltic Sea, is situated between the island of Rügen and the western Pomeranian mainland. Having an extension of 510 km² its depth does not exceed 13.8 m. The characteristics of both the shorelines and the sea floor topography originate from the configuration of the moraines built up during the last glaciation, i.e. Weichselian: Pomeranian (W2) and Mecklenburgian (W3) Stages (Rühberg et al. 1995). The depressions of this Pleistocene relief were filled with water and sediments during three stratigraphic und depositional phases (Verse et al. 1998): (1) Pleistocene glacial/ice decay landscape from High- to Late Weichselian showing tills, diamictites, gravels, sands, clayey silts which, in places, are deformed by gravitational processes, (2) Early to Middle Holocene, i.e. Preboreal/Boreal to Older Atlantic, phases of the Baltic Ancylus Stage originating in a landscape of small lakes which could be linked by ephemeral fluvial activities (small rivers, creeks, channels), (3) Middle to Late Holocene (Older Atlantic to Subatlantic) phases which were characterized by a stepwise Baltic sea level rise (Littorina–Transgression) and changing the former terrestrial into a marine-brackish environment by widespread transgressive flooding processes. The deposits of the latter phases mainly consist of mixed organo-clastic sequences (mud, peat, silt, sand). Especially, the marine-brackish transgressive sandy sediments formed long beach barrier systems which linked the Pleistocene morainic hills of the neighbouring islands of Rügen and Usedom.

For the Greifswalder Bodden, it is assumed that a lake level lowstand during Ancylus time (phase 2) reached 12 to 14 m below present sea level (mbsl). There is evidence from drill cores (Verse et al. 1999) showing strong water level fluctuations and rapid changes in the sedimentation processes being controlled not only by the eustatic sea level rise but also by storms. The latter are recorded in tempestitic sequences which might originate from allocyclic sea level variations of higher order (1–3 m). A maximum flooding surface detected in drill cores revealed a relative sea level highstand at the end of the Atlantic Stage. A sea level curve established by Janke & Lampe (2000) covering the last 8,000 years indicates phases of land uplift along the Pomeranian coast, mainly between 5,800 and 5,000 BP, which resulted in a seaward shoreline displacement. Only in Subboreal to early Subatlantic times the sea level has reached its former position again. For the last 2,000 years a general phase-like sea level rise is suggested which might involve both local sea level fall and storm-controlled superimposition of eustatic sea level rise. Further neotectonic control by isostatic movements may not be excluded even for the latest submodern/modern times. This will be shown and discussed by the occurrence of the stratigraphic Baltic Mya-Period in the deposits of the Greifswalder Bodden.

57 box cores have given insight to the stratigraphy of the deposits of the Greifswalder Bodden up to 40 cm below sediment surface. Two areas are evident (Niedermeyer et al. 1993, 1994, 1995): (*i*) a muddy basin area of 8–9 m depth in the western part; (*ii*) shallow areas with mostly fine sandy sediments between 2 and 8 m depth both in the marginal parts of the bay and the ridges in the eastern part. The ridge area separating the Greifswalder Bodden from the open Baltic Sea is dissected by several troughs reaching depths of 12 m and having the sedimentary character of the basin area in the west. Additional 20 box cores from the westerly adjacent slope to the Pomeranian Bay also exhibit a sedimenntary succession from coarse sandy to muddy in depths exceeding 10 m.

The most conspicuous feature in the sediments of the Greifswalder Bodden and the adjacent area is the occurrence of *Arenomya arenaria* in the upper few decimetres of the sediment column (Hertweck 1994). In the basin area in the western part of the bay the *A. arenaria* population is less abundant and represented only in some of the box cores. In the marginal belt and in the ridge area of the eastern part most of the sites are populated by living *A. arenaria* down to a sediment depth of about 10 cm. Dead *A. arenaria* shells in life position extend to sediment depths not exceeding 20 cm. Only in two samples shells of deeper extension were found, i.e., down to 25 cm in core GB 115 and 36.5 cm in core GB 128. Thus, only the upper 20 (to 40) cm of the profile exhibited in the box cores represent the Mya-Period which has established in the Baltic Sea after the Littorina and Lymnaea Periods. However, this representation is not valid for the sediment surrounding dead *A. arenaria* shells in life position found in deepest position in the cores. For endobenthic bivalves contemporaneous sediment is situated on the level of the siphon passage openings, i.e., considerably above burrowing depth. Accordingly, the "deepest *Mya* shell" from core GB 115 was sticking in older limnic deposits probably belonging to a pre-Littorina stage of the Baltic Sea.

Reworked *Arenomya* shells have been found only in box cores taken in water depths of less than 5.8 m. Three reworking horizons in 15, 8 and 2 cm sediment depth suggest an origin by three century-scale high magnitude storm events during the whole Mya-Period. In the adjacent area of the western Pomeranian Bay such reworking horizons occur down to a water depth of about 9 m, according to a considerably higher wave energy in the open Baltic Sea.

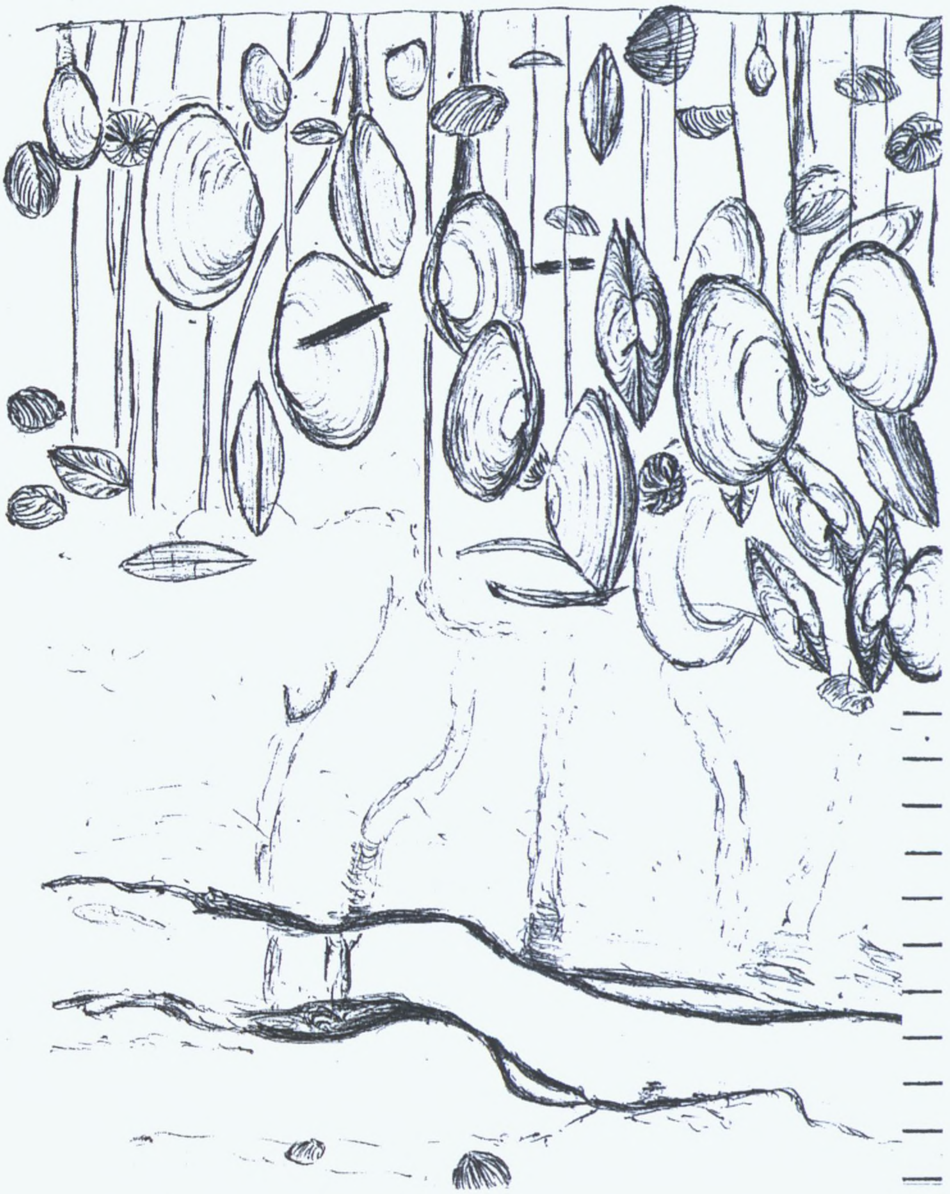

Figure. Typical example of the Mya-Stage deposits in a box core from the Greifswalder Bodden. The Mya layer shows living specimens of Arenomya arenaria in the upper part and dead shells in life position in the middle and deeper parts. Active polychaete burrows extend from the surface to about 12 cm depth. In the lower half of the core old sand-filled polychaete burrows occur which partly penetrate two mud flasers present at about 18 to 23 cm depth. – Scale 10 cm.

Along with *Arenomya. arenaria* living specimens and dead shells of *Cerastoderma edule* and *Macoma balthica* were found in the upper portions of the box cores from the Greifswalder Bodden. Shell material of these two species also occurs in the deeper portion of the cores, together with shells of *Scrobicularia plana*. This species regularly populates the western Baltic Sea which has a higher salinity than the average value of 7.3 measured in the Greifswalder Bodden. Its absence in contemporary southern Baltic and in the sediments of the Mya-Period points to a decrease in salinity at the end of the Littorina-Period in this region.

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DATABASE FOR THE GEOLOGICAL COLLECTIONS HELD AT THE INSTITUTE OF GEOLOGY AT TALLINN TECHNICAL UNIVERSITY

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It was understood already in the 1960–1970s that computers could largely facilitate the storage and processing of enormous amounts of data related to various collections, including geological collections (Brunton, 1979). Since then the use of computers and electronic databases has increased considerably. The Institute of Geology at Tallinn Technical University is one of the owners of large geological collections *sensu lato* (i.e., palaeontological specimens, rock samples, drill cores, etc.) in the Baltic region. The very first computerization attempts made at the institute clearly pointed to the need for a further effort in this field.

The basic requirements for an institutional database were defined as to register and keep track of, and perform various queries on (a) *individual objects*, such as fossil or mineral specimens, the type- and other published specimens given the priority; (b) *samples*, microfossil preparations, etc.; (c) institute's *drill cores*; (d) *localities* and sites (boreholes and outcrops) where objects or samples derive from, or which are otherwise geologically important; (e) *bibliographic information* (publications, manuscripts, maps, etc.) related to objects or localities.

While looking for an appropriate software solution there were also some other important aspects to keep in mind. The system should (a) be scalable and secure; (b) be easily customizable, especially what concerns generating specific queries and designing reports; (c) handle certain information in two languages (Estonian and English); (d) have output to the Internet; (e) have support for linking multimedia-, text- and other files; (f) not be pricey.

The principal decision to make was whether to go with some of the standard museum cataloguing programs or start developing a separate system. There are quite a few such programs around that are efficiently used by the largest museums in the world. Some of those like *Specify* are available for free use (yet without the possibility of modifying the program), whilst the others like *KE EMu* are commercial products. Also, Estonian Ministry of Culture supports the development of a special program called *KVIS* to be used in various Estonian museums.

However, the collections of the institute are rather heterogeneous. Besides properly stored and organized specimens, they often contain material that has been collected for practical reasons such as geological mapping and is not so well documented. This material is usually organized by localities or stratigraphical levels rather than systematically. Also the samples and drill cores are not a typical part of a geological museum, being instead a characteristic component of a geological survey collection. Geological surveys, unlike museums, usually develop their own information systems, which are not made available to others. Consequently, none of the programs on hand appeared to be perfect for our needs and it was considered expedient to develop a system from the scratch to meet the required functionality. It was still kept in mind that possible migration to another, more sophisticated system in the future must be as straightforward as possible.

The client-server architecture of such a database system provides several advantages over simpler solutions such as file-server database and is definitely the way to go. At the time of this writing, however, moving from the file-server approach to client-server system as depicted in Fig. 1 is in halfway depending upon certain hardware upgrade.

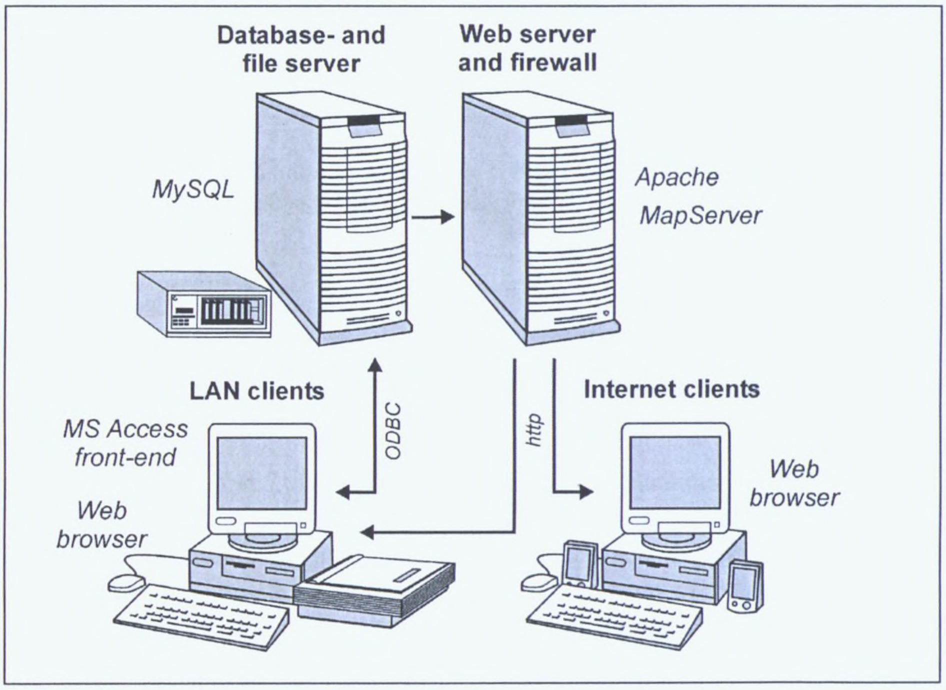

Fig. 1. Schematic architecture of the system showing system components and data flows.

To keep the budget low, the main software components of the system were assembled from open-source software and/or programs that are installed in computers independent of the database usage.

Open source *MySQL* was chosen as the database engine. It is free for non-commercial usage and, although its functionality is partly inferior to such commercial solutions as MS SQL Server or Oracle, it is very fast and reliable and is developed rapidly.

For data entry, sophisticated queries and various reports like printing labels or catalogues a front-end application based on *Microsoft Access* has been developed. Access is a part of the MS Office bundle, which is used in most of the institute's computers and therefore required no additional licensing. Moreover, as the application is consistently improving, usage of some more sophisticated programming environment would have been impractical at this stage. The connection between the database server and front-end is realized using *ODBC* (Open Database Connectivity).

Great part of the information to be stored in the database, data on the type- and other published fossils in particular, must be available to researchers worldwide. As pointed by, e.g., MacLeod & Guralnick (2000), the World Wide Web is the best way for distributing such data. Output to the Internet was therefore considered vital for the database. By employing an appropriate privilege- and authentication system, some sensitive information like museum locations or data on unpublished samples and drill cores could also be queried using the web interface, which is a way easier to use than the Access-based application. It needs no other programs but a *Web browser* (most common options are *MS Internet Explorer*; *Netscape/Mozilla*, and *Opera*). Presently the web interface is programmed using *PHP scripting language*, which is a usual selection with *Apache* web server. An important part of the system is also *MapServer*, an open source web map server, which, being not a full-featured GIS, is performing well for plotting GIS data, e.g., spatial distribution of a fossil. Some functionality of the Access front-end, like displaying images and maps also relies on the web server and hence requires a browser.

It was clear from the very beginning that the relational database model suited best for the specific needs of the appropriate information system. It is efficient in terms of disk storage, helps to keep minimal the data redundancy, which is the main cause of data inconsistency, and greatly facilitates the data entry procedure.

On the other hand, a too strict model would need more effort and time during data input. Thus the number of strong relationships was kept relatively low, using additional free-form fields to complement the hard-coded fields that must have related records in other tables. From time to time, such entries are automatically grouped and updated using special queries. Using this methods helps to speed things up especially when non-geologists are involved in data entry.

The simplified data model is illustrated in Fig. 2. It consists of over 15 main tables, which are supported by several other tables, such as linkage tables in case of many-to-many relationships, and simple tables providing predefined lists of values that can be entered into a particular field.

The ***object*** table can be regarded as the most important part of the system. It contains data on all items to be given individual unique specimen numbers and identifications. It is related to many other tables, most importantly *history*, *locality*, *reference*, *museum location*, *classification*, etc.

Every object belongs a certain collection, which might be defined by different criteria (e.g., is published in one paper, is collected by one person, belongs to the same taxonomic group). Collections' details are stored in the ***collection*** table.

The ***sample*** table holds data somewhat similar in the object table. However, a sample is not something on its own, it has been collected for further research like mineralogical or geochemical analysis or microfossil content. It may be a piece of rock, but it may also no longer exist, being just a source for some kind of information, or, for instance, a microfossil slide. Samples are related to microfossil preparation table, which can hold data on the occurrences or counts of particular taxa. Using crosstab queries, one can produce taxa/sample matrices, which in turn may be used for plotting faunal logs. Thus, this module has some of the functionality that some more specific micropalaeontology programs like CHITINOS (see Achab et al., 2000) have.

Drill cores are registered by individual core boxes. Also information about various activities on cores, such as sampling events or photographing, is stored in the database. Drill cores as well as samples and related information are mostly for internal application.

Localities are boreholes, outcrops, or eventually geographical areas where objects, samples or cores derive from. Their position is determined by the geographical coordinates, which makes it possible to plot them on the map, or perform certain queries using other map layers available, such as geological- or administrative maps.

The ***reference*** table contains information about publications as well as manuscripts, such as field notebooks, providing some sort of data on objects, localities, drill cores, etc.

For a better performance, photos, text- and other files are stored separately from the database. However, every such file must have a corresponding record in the ***file*** table that also stores the information as to which table and particular row the file is corresponding.

The ***museum location*** table holds information on individual museum locations such as rooms, cabinets, drawers, shelves or boxes. Brief description of the content of a location can be used to record basic information about groups of samples or objects prior to their one-by-one registration.

The ***classification*** and ***stratigraphy*** tables are built hierarchical, so there is no need for specifying, e.g., Phylum, Class, Order, Family, etc. for every fossil. It is sufficient to record the lowest level present in the

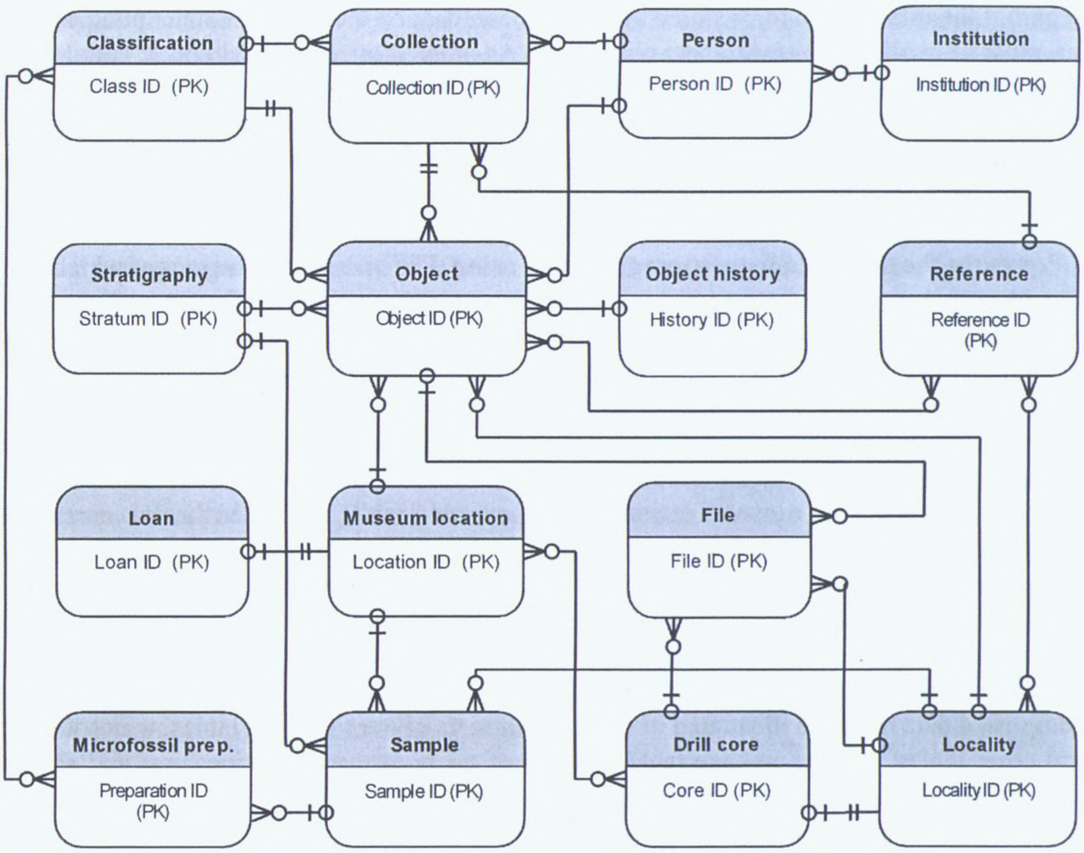

Fig. 2. Simplified data model, showing the main objects in the database and principal relationships between them.

predefined hierarchy and it would later be possible to search by any hierarchy level. This method puts indeed a great responsibility to the designer of the hierarchy.

There are several other aspects of this data model but describing all these is beyond the scope of this overview.

As of April 2002, some 22 000 individual objects, including nearly 8 000 type-, figured-, and cited specimens are entered into the database. In addition, some 20 000 samples, the drill cores (including over 1000 records on individual core boxes), 10 000 microfossil preparations, etc., and the accompanying information are stored. This is still less than 10% of all data to be inserted into such an institutional database. Hence, in the forthcoming years the main effort must be put into the data entry. Also the application itself needs several improvements, which would increase data consistency and overall functionality of the database. There is, however, no doubt that even this relatively small amount of data currently in the electronic form has been beneficial for the curatorial practice. The advantage of electronic database will indeed further increase with the increasing amount of data.

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SOURCES OF KIMBERLITIC MINERALS IN CLASTIC SEDIMENTS OF LATVIA AND SOME PROBLEMS IN SUCCESSION OF FORMATION OF KIMBERLITE

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Kimberlitic minerals are known in the composition of sandy deposits of the Ogre Formation (outcrops in valley of the Abava River) and of the Ketleri Formation (outcrops in valleys of the Venta, Ciecere and Paksīte Rivers) of the Upper Devonian in Kurzeme (Sorokin, Krivopalov, Mūrnieks, Savvaitova and Samburg 1992) and as well in the composition of the Quaternary sediments. During the last decade numerous finds of kimberlitic minerals were discovered and researched in detail as in the beach placers both of the Baltic Sea and the Gulf of Rīga and as well as in the alluvium of the left tributaries of the Gauja River (Savvaitov, Veinbergs, Nulle, Samburg and Stinkulis 1998, Savvaitovs A., Savvaitova L., Veinbergs I. 1999, Hodireva, Nulle, Samburg, Savvaitov, Veinbergs 2000, Hodireva V., Samburga Ņ., Savvaitova L., Savvaitovs A., Veinbergs I., Nulle U. 2000, Savvaitovs, Veinbergs, Samburga, Hodireva and Nulle 2000, Korpečkovs, Samburga, Savvaitovs, Hodireva 2001, Hodireva, Korpečkovs, Samburga, Savaitovs 2002). According to them, existing notions about the composition and distribution of kimberlitic minerals in area of Latvia are broaden and supplemented. These finds reflect the features of the concentration and composition of kimberlitic minerals containing in eroded tills. Therefore its characterise petrological and mineralogical signs of the kimberlites also, the bodies of which had been destroyed, their material had been transported and accumulated into tills by advancing ice covers.

Mainly signs distinguishing different kimberlitic mineral assemblages containing in clastic deposits and sediments of Latvia are discussed below. The reconstruction of the locations of supposed kimberlites is given also.

There are five types of kimberlitic mineral assemblages. Two of them are reflected in the composition of kimberlitic assemblages containing in the Upper Devonian deposits: the first – in <u>the Ogre Formation</u> and the second – in <u>the Ketleri Formation</u>. The chrome-pyrope containing in kimberlitic mineral assemblages of these two types had been researched in detail only. The chrome-pyrope of their in comparison with other types is more magnesian and is characterized by higher content of the pyrope molecule. The chrome-pyrope occurring in the Ketleri Formation in contrast to the chrome-pyrope from the Ogre Formation and from other types is distinguished by poor content of chrome. Besides that, the chemical composition of the chrome-pyrope from the Ketleri Formation not content the knoringite molecule. The few grains of the chrome-pyrope from this Formation are located in field (Cr_2O_3–CaO diagram of Sobolev) of dunite-harcburgite association. The chromespinelide found in the deposits of the Ketleri and Ogre formations also.

Three other types of kimberlitic mineral association are observed in the composition of above-mentioned Quaternary sediments. Each fixed type of kimberlitic association has own zone of the distribution. There are three such zones: (1) the Western Kurzeme zone, (2) the zone of the Gulf of Rīga and (3) the Northern Vidzeme zone.

Kimberlitic minerals found in the composition of the beach placers of the Baltic Sea characterise <u>the Western Kurzeme zone</u>. This zone is distinguished by higher content of the grains of the chrome-pyrope. The chrome-pyrope in contrast to the chrome-pyrope from the Ogre and Ketleri Formations is less magnesian and the pyrope molecule is contented in decreased quantity. In addition the chrome-pyrope of this area is richer in chrome. The grain of the chrome-diopside found in the coast of the Baltic Sea around Ulmale and Labrags. In the Western Kurzeme zone the chromespinelide occur in a little number. The beach placers in the coast of the Baltic Sea are richer in the olivine, pyroxene, magnetite, than in other sites of kimberlitic minerals.

Kimberlitic minerals found in the composition of the beach placers of the Gulf of Rīga characterise common <u>zone of the Gulf of Rīga</u>. Here, the content of the chrome-pyrope is smaller, than in the Western Kurzeme zone. The chrome-pyrope from this zone to contrast of the Western Kurzeme zone contain increased amounts of the manganese, chrome, knoringite molecule and decreased content of the pyrope molecule. It is important, that the few grains of the chrome-pyrope correspond to field (Cr_2O_3–CaO diagram) of the dunite-harcburgite and diamond associations. The chromespinelide among heavy minerals is contented in more quantity, than in the Western Kurzeme zone. Some grains of the chromspinelide contain Cr_2O_3 in more than 60 mas.%. The chrome-diopside occurs around Timmāji and Gauja only.

Kimberlitic minerals containing in the composition of the alluvium of the left tributaries of the Gauja River characterise common zone of the Northern Vidzeme. Here, the list of the kimberlitic minerals is broader. The chrome-pyrope, chromespinelide, chrome-diopside and muassonite there are represented. The grains of the chrome-pyrope have often dark-violet and crimson colours in this zone. The chrome-pyrope usually is characterised higher content of the chrome and knoringite molekule too, but the grossular molecule is absent. The amount of the pyrope molekule is decreased. The some grains of the chrome-pyrope correspond to field (Cr_2O_3–CaO diagram) of the dunite-harcburgite and diamond associations. The chromespinelide in contrast to other zones is contained in more quantity and some grains of it contain Cr_2O_3 in more than 60 mas.%. It is the signs of the kimberlites with diamond. Here, the chrome-diopside in contrast to other zones found more often. The halos of the chrome-pyrope are observed on researched area of the Northern Vidzeme zone. The finds of chromspinelide with higher contents of Cr_2O_3, chrome-diopside and muassonite concur with them.

According to the situation of each established zone of kimberlitic minerals and as well to the sites of the Upper Devonian deposits with kimberlitic minerals and to observed features of composition and concentration of kimberlitic minerals also, the reconstruction of the distribution of supposed kimberlites is done. The field of distribution of kimberlites may to suggest on the bottom of the Baltic Sea to Northwest from recent shoreline. This field locate not far. Therefore the concentration of chrome-pyrope from researched beach placers of the coast of the Baltic Sea is rich. The second supposed field of kimberlites, which locate not far from the finds of kimberlitic minerals, is the area of the Northern Vidzeme to north and Northwest from the finds of the kimberlitic minerals in alluvium of left tributaries of the Gauja River. The finds of the chrome-diopside from the Quaternary sediments both in the Western Kurzeme and Northern Widzeme zones of kimberlitic minerals are important signs of near locations of kimberlites. The third field of kimberlites is located not far from site of the Paksīte River contenting the highest amount of the grains of the chrome-pyrope in the Ketleri Formation. This field is situated in central part of the Eastern Kurzeme. The location of kimberlites, the material of which had been reflected in composition of kimberlitic association in deposits of the Ogre Formation (Abava River) to determine difficult until. Sources of fixed kimberlitic minerals within zone of the Gulf of Rīga probably located on a long distance. Found kimberlitic minerals in zone of the Gulf of Rīga had been transported from far regions.

The age of the kimberlites is post-Middle Devonian. However, it is possible that there are two varieties of the kimberlites distinguishing by geological age.

The scheme of the distribution of kimberlitic minerals and supposed situation of the kimberlites in Latvia is shown in Fig. 1.

Authors dedicate this paper to memory of Prof., Hab. Dr. Visvaldis Kuršs and Hab. Dr. Ints Veinbergs.

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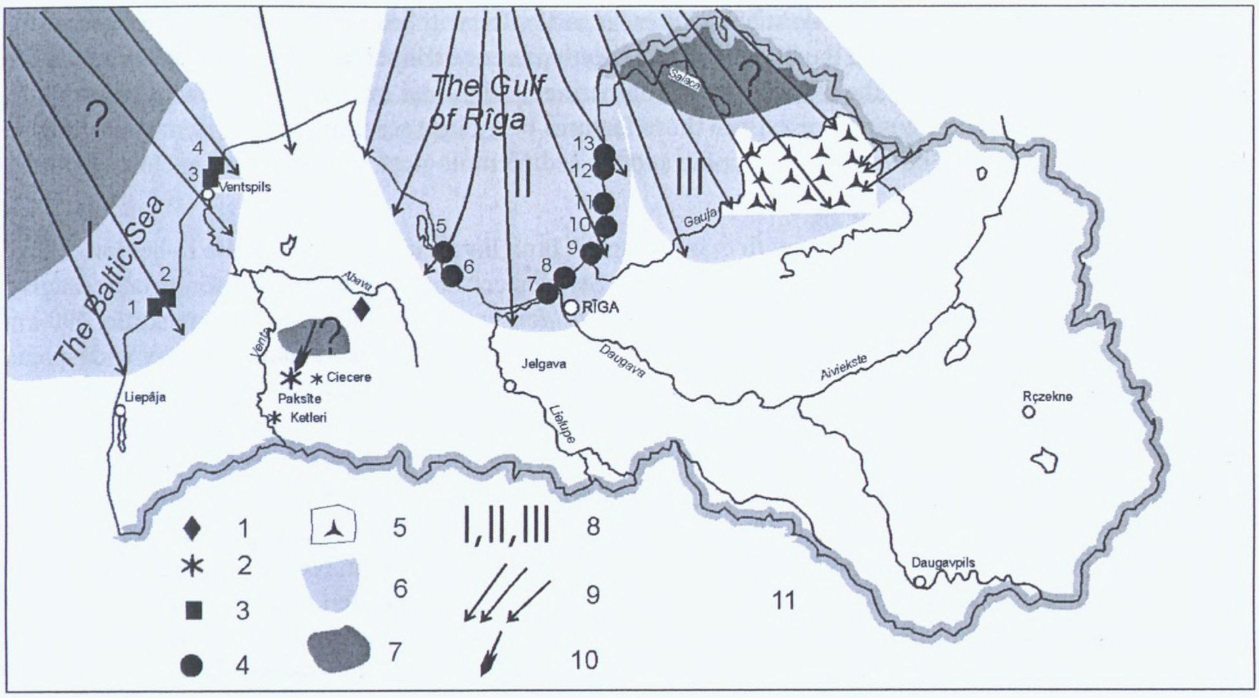

Fig.1. The scheme of the distribution of kimberlitic minerals and supposed situation of the kimberlites in Latvia. 1 – site of kimberlitic minerals in the Ogres Formation, 2 – sites of kimberlitic minerals in the Ketleri Formation, 3 – beach placers of the Baltic Sea containing kimberlitic minerals (1 – Ulmale, 2 – Labrags, 3 – Staldzene, 4 – Liepene), 4 – beach placers of the Gulf of Rīga containing kimberlitic minerals (5 – Engure, 6 – Ķesterciems, 7 – Buļļusala, 8 – Mangaļsala, 9 – Gauja, 10 – Timmāji, 11 – Korbiņi, 12 – Ķurmrags, 13 – Ēžurga), 5 – researched region of left tributaries of the Gauja River, in alluvium of which found kimberlitic minerals, 6 – supposed zones reflecting the drift of kimberlitic minerals, 7 – regions of distribution of supposed kimberlites, 8 – indexes of zone of kimberlitic minerals: I – the Western Kurzeme, II – the Gulf of Rīga, III – the Northern Vidzeme, 9 – directions of glacial drift, 10 – direction of transportation of kimberlitic minerals in the Ketleri basin, 11 – the northern boundary of the Ketleri deposits distribution.

K-BENTONITES AS CLUES TO PALEOGEOGRAPHIC RECONSTRUCTION

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INTRODUCTION

Fallout pyroclastic ash beds (K-bentonites) preserved in the Ordovician stratigraphic record are intuitively treated as event beds because of their many similarities to contemporary deposits formed from explosive volcanic eruptions. Extending that analogy, these beds are generally considered to represent chronostratigraphic surfaces whose time frame of origin is geologically instantaneous, though the actual formative events probably lasted from days to weeks. Regional correlations of enclosing strata often rely on K-bentonite stratigraphy since they may represent both a higher degree of resolution than established biozones and they are thought to carry less ambiguity with respect to time correlative horizons that either biozone or sequence stratigraphic boundaries. For the most part, these are reasonable assumptions to make, however, our continuing studies of individual bed characteristics on a regional scale have revealed several features of K-bentonite beds that add complexity to their application as event stratigraphic deposits. The complexities arise from several features of K-bentonite beds that have long been recognized but have not been fully integrated into their interpretation. Two such features, in particular, will be discussed here. One concerns the composite nature of some beds and the likelihood that they represent multiple rather than single accumulation events, and the other concerns the association of submarine hardground or omission surfaces upon which some K-bentonite beds are found to rest. Both these features bear close scrutiny since they may easily give rise to misinterpretation of both the stratigraphic and the sedimentological records of the enclosing strata.

MULTIPLE EVENT BEDS

We will illustrate this phenomenon with two examples, both involving the Kinnekulle K-bentonite bed. The Middle Ordovician section at Röstånga in southern Sweden contains eighteen K-bentonite beds ranging from 1–67 cm in thickness, and all occur within the *D. multidens* graptolite biozone. At Kinnekulle, 290 km to the north, this interval includes the type section of the Kinnekulle K-bentonite, which is very widespread and has been correlated throughout northern Europe and across Iapetus to North America (Bergström et al., 1995). In most sections of this age the Kinnekulle K-bentonite can be recognized by distinctive geochemical fingerprints, its prominent thickness, and by its biostratigraphic and lithostratigraphic position. However, at Röstånga whole rock chemistry is inconclusive at identifying which of the eighteen beds is the Kinnekulle K-bentonite. Several beds at Röstånga correlate equally well with the Kinnekulle bed (Bergström et al., 1997). Other, more definitive methods were used to try to resolve the problem. Nineteen samples representing both the Röstånga and the type Kinnekulle section in south central Sweden were wet sieved using 200 mesh and 70 mesh screens. Primary and authigenic phenocrysts were mounted and polished for petrographic study. For the Kinnekulle K-bentonite and most of the Röstånga K-bentonite beds, biotite is the primary phenocryst ranging from 22% to 90% of the total phenocryst population. Quartz and feldspar make up the bulk of the remaining phenocrysts, with lesser amounts of apatite, zircon, Fe-Ti oxides, and amphibole. Feldspar is dominantly orthoclase with occasional microcline and plagioclase.

Based on mineralogical percentages and phenocryst composition the Kinnekulle bed equivalent at Röstånga may be interpreted to be represented by several closely spaced K-bentonite beds perhaps in the 3–5 m interval, as suggested initially by whole-rock chemical fingerprinting. These beds are separated by thin, biotite-rich limestone layers indicating *in situ* reworking of the ash, selective winnowing of micas, and mixing with carbonate-rich sediment. Additional work on the Röstånga section is underway (Pålsson et al. this volume), including an evaluation of possible abbreviation of the Röstånga section by faulting.

The Middle Ordovician Millbrig and Deicke K-bentonite beds occur throughout the eastern Midcontinent of North America (Kolata et al., 1998). On the basis of biostratigraphic and geochemical data Huff et al. (1992) proposed that the Millbrig and Kinnekulle beds were equivalent and represented components of the same eruptive event. An implication of that correlation was that the beds were chemically homogeneous, which permitted them to be correlated, and that the unaltered phenocrysts of primary minerals must therefore be similarly uniform in both North America and Baltoscandia. Haynes et al. (1995) tested that hypothesis by comparing the composition of unweathered biotite grains from each of the three beds. They found that biotite from the Millbrig and Kinnekulle beds, while similar in some respects, was sufficiently different that the proposed equivalence of the two beds could not be confirmed.

Additional microprobe data on biotites from multiple localities of the Deicke, Millbrig, and Kinnekulle beds, and also from multiple samples within the Millbrig and Kinnekulle beds (Fig. 1). These data show clearly that the Deicke and, to a lesser degree, the Kinnekulle beds represent deposits from compositionally homogeneous magmas. However, the Millbrig shows considerable within-bed variability and has biotites that range from only slightly different from Deicke to indistinguishable from Kinnekulle. Figure 1A and 1B show Ti plotted against total Fe/Mg, and Mg against Al. Both illustrate the diverse nature of Millbrig biotite compositions. Figures 1C and 1D plot what is known as the Mg number, which is the ratio of Mg against Fe+2 plus Mg. The Mg number is frequently used by igneous petrologists as an index of fractional crystallization, with values decreasing from 1.0 as magmas become more highly evolved and thus more Si rich. Both plots of Mg# against Mg and against Ti show a trend in the Millbrig biotites from a slight overlap with the least evolved Deicke K-bentonite to the most highly evolved Kinnekulle. The uppermost portion of the Millbrig is indistinguishable from the Kinnekulle when viewed from the standpoint of biotite composition. Our data suggests that the Millbrig consists of multiple ash falls, the last of which is essentially indistinguishable from the Kinnekulle. This finding is consistent with the field observations of the Millbrig which, in the southern Appalachians where it has its maximum thickness, generally is seen to consist of several fining upward subunits with coarse biotite and feldspar at the base of each layer (Haynes 1994).

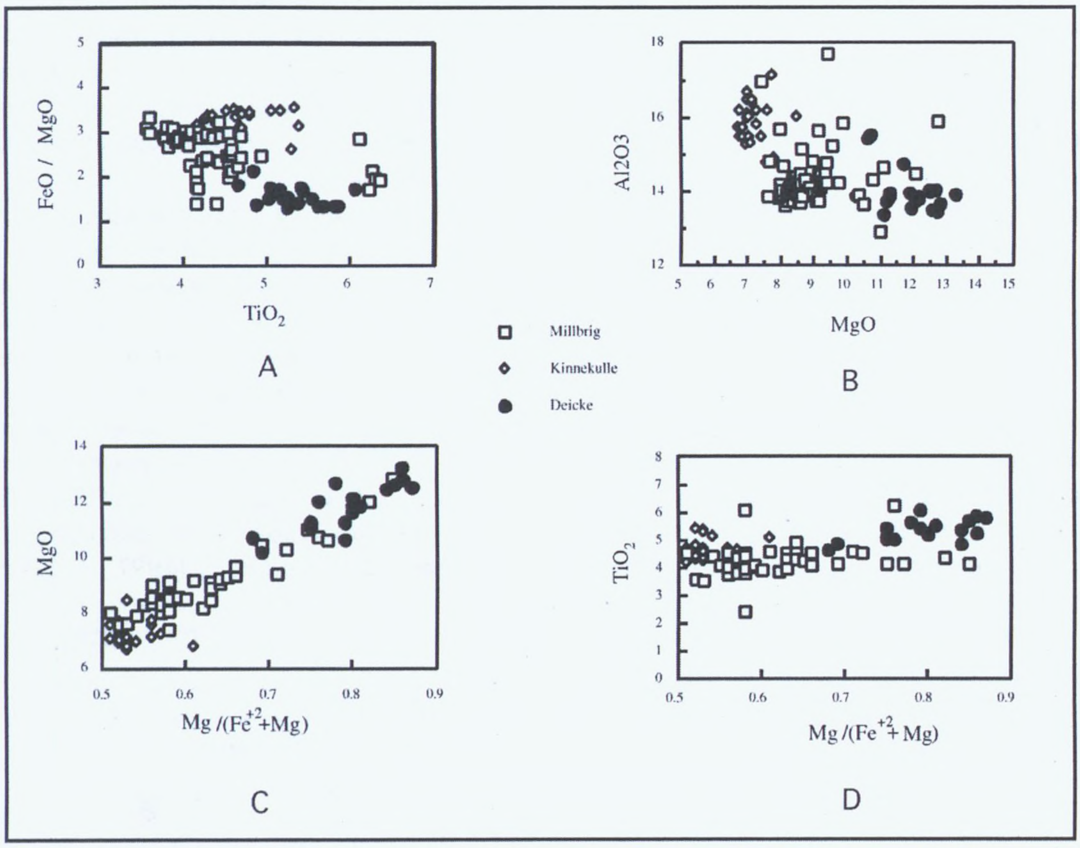

Fig. 1. Bivariate plots of microprobe data from biotite in the Millbrig, Deicke, and Kinnekulle K-bentonites. A and B illustrate the variable nature of Millbrig biotite composition compared with the other two. C and D show the variation in the magnesium number between the three beds.

TRACE ELEMENT GEOCHEMISTRY

The use of immobile trace elements in altered and unaltered volcanic rocks to discriminate compositions erupted in different tectonic settings is well known (Harris et al., 1986; Pearce and Cann, 1973; Pearce et al., 1984; Winchester and Floyd, 1977; Wood, 1980) Discrimination diagrams which are derived empirically from many analyses of lavas must be cautiously applied to the interpretation of altered pyroclastics, however, and with due consideration for the possible effects of phenocryst fallout and other lateral changes on whole rock composition. These problems are minimized in the case of rapidly formed vitric ash beds since pyroclastic glass tends to maintain a constant composition with distance from its source (Sarna-Wojcicki et al., 1987). Preliminary studies of high field strength (HFS) elements in K-bentonites (Huff et al., 1991; Merriman and Roberts, 1990; Roberts and Merriman, 1990) indicate that they largely behave as immobile elements and do, within the range of sensitivity offered by empirical discrimination diagrams, preserve characteristics of the original magma. They can thus be useful in reconstructing the setting of subduction-related island arc and plate margin volcanism. While not providing conclusive proof of origin, immobile element compositions of K-bentonites can reveal important information concerning both the compositional characteristics of the parent magmas and the probable tectonic environment in which the eruptions occurred. Roberts & Merriman (1990) studied the mineralogy and trace element geochemistry of 3 Cambrian and 1 Ordovician (Caradoc) K-bentonites from the Welsh Basin, and concluded all were derived from felsic magmas in a within-plate tectonic setting. In a companion study (Merriman and Roberts, 1990) of Upper Ordovician–Lower Silurian K-bentonites from the Southern Uplands they found evidence that the tectonic setting of the source volcanoes changed from an ensialic arc transitional to a back-arc setting and involved attenuated sialic lithosphere and depleted mantle in the generation of the parent magma.

ARGENTINA

During the past few years, many Lower and Middle Ordovician K-bentonite beds have been found in the Precordillera of western Argentina; indeed, this region has some of the most abundant ash beds known from that period anywhere in the world. Ordovician K-bentonites have been recorded from more than 20 localities in a region extending about 250 km in a north-south direction across parts of San Juan and La Rioja Provinces in Argentina. Most of the known K-bentonites occur in the eastern thrusts of the thrust and fold belt, where they are quite common in the upper section of the San Juan Limestone and in the overlying Gualcamayo Formation. A few ash beds are also known from the central thrusts.

Probably the oldest known Paleozoic K-bentonites in the Precordillera occur in the middle Arenig in the topmost part of the Lower Ordovician San Juan Limestone in outcrops along the Gualcamayo River in the Guandacol region. In the overlying Gualcamayo Formation, which ranges from the Upper Arenig to at least the Middle and probably Upper Llanvirn, there are numerous ash beds. This is by far the most extensive suite of Lower Paleozoic K-bentonites recorded from an outcrop anywhere in the world. Most of the ash beds, which range in thickness from less than a mm to more than 50 cm, occur in the upper Arenig *U. austrodentatus* Zone (Brussa and Astini, 1997) but many beds are present also in slightly older and slightly younger strata.

In the Jáchal area, there are about 30 individual K-bentonite beds in the Gualcamayo Formation (= lower part of the Los Azules Formation) at Cerro Viejo (Bergström et al., 1996; Cingolani et al., 1997; Huff et al., 1995) and up to a dozen such beds have also been recorded from the upper part of the underlying San Juan Limestone. Concordant zircon ages from Cerro Viejo were reported by Huff et al. (1997) as 464 ± 2 Ma which they considered to be the age of the base of the *U. austrodentatus* Zone.

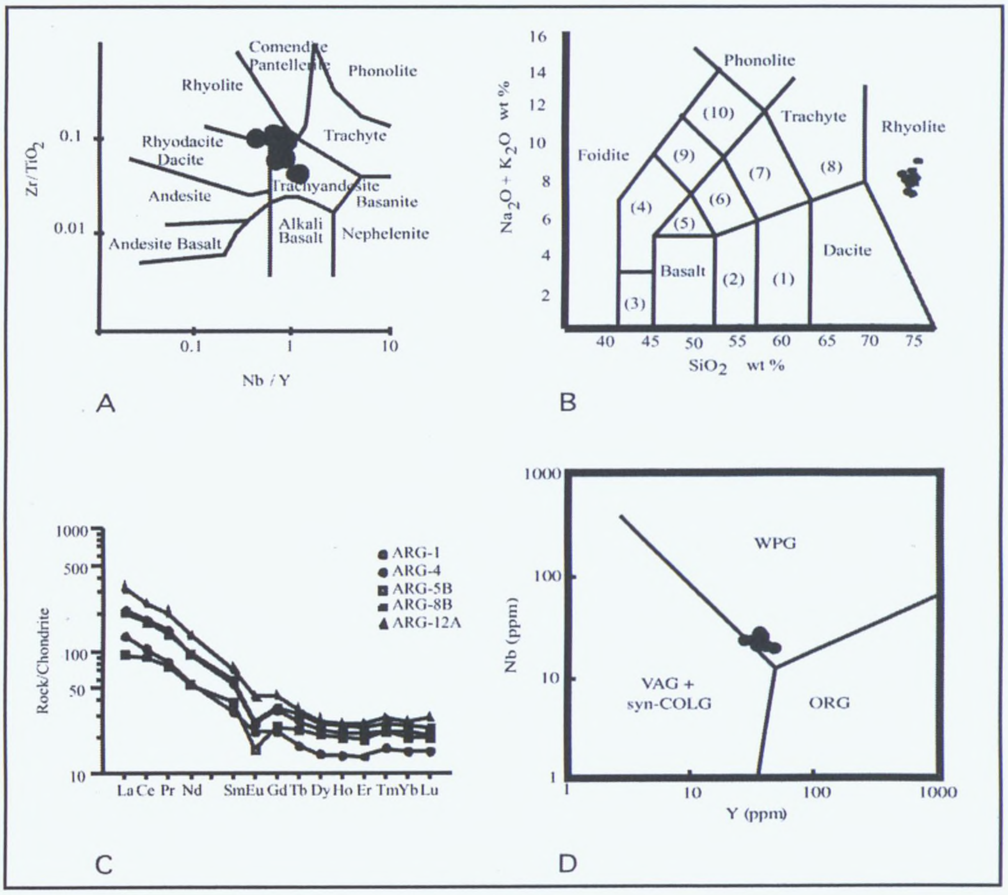

Fig. 2. (A) Zr/TiO_2 vs. Nb/Y plot after Winchester and Floyd (1977) showing the magmatic composition of the Cerro Viejo samples based on immobile element ratios. (B) Quartz-hosted glass melt inclusions from Cerro Viejo were analyzed by electron microprobe and the data plotted on a total alkalies vs. silica (TAS) diagram. (C) A chondrite-normalized rare earth element (REE) plot of 5 Cerro Viejo samples. (D) Tectonic discrimination diagram from Pearce et al. (1984) showing the position of the Cerro Viejo volcanics in terms of granitic origins. WPG = within plate granite; ORG = ocean ridge granite; VAG = volcanic arc granite; syn-COLG = syn-collision granite.

K-bentonite geochemistry may provide highly significant information about the tectonomagmatic nature of the source area, and the distribution patterns of individual ash beds or complexes of such beds may shed additional light on the former positions of continental plates. K-bentonite beds also provide important evidence regarding the timing and location of orogenic events because they typically originate from source volcanoes situated at or near tectonically active plate margins (Huff et al., 1992; Kolata et al., 1987). Further, dating of K-bentonite beds or complexes of beds provides precise age dates on periods of volcanism associated with continental margin subduction events.

CONCLUSIONS

Palaeogeographic reconstruction for the early Palaeozoic is, at best, a "weight of evidence" process, in which the more pieces of information there are available, the more solid the arguments for a particular scenario. K-bentonite stratigraphic and geographic distribution patterns can be helpful in this process by providing direct evidence of subduction-related explosive volcanism and the accompanying palaeowind directions, sedimentary conditions and magmatic processes which existed at the time. In Argentina, for example, it is clear that subduction-related volcanism in the Precordillera commenced in the mid-Arenig and continued through the late-Llanvirn, coeval with volcanism in the Famatina range. Based on the tectono-stratigraphic affinity of the Cambrian to early Ordovician carbonate sequence in the Precordillera that exhibits similarities to sections along the Laurentian margin it has been proposed that the Ocloyic event of the Famatinian orogeny in South America was a continuation of the Taconic orogeny in North America (Dalla Salda et al., 1992; Dalziel et al., 1994). However, this scenario is not supported by the K-bentonite record. That is, whereas the Precordilleran K-bentonites are Arenig to early Llanvirn in age, those associated with the Taconic orogeny are of mid-Caradoc age. While the Precordillera may well have had it origins on the Laurentian margin we do not find supporting evidence that it was still there during mid-Ordovician time. Rather, we find the ash pattern is more consistent with the palaeogeographic reconstructions of Mac Niocaill et al. (1997) who envision drifting of the Precordillera in fairly close proximity to one or more additional volcanic arcs with eventual collision along the Andean margin of Gondwana during the Ocloyic orogeny in Middle Ordovician time. The extensive K-bentonite record in the Guandacol region suggests that the source volcanoes may have been located north or northeast of the Precordillera. Conceivably, the Puna–Famatina terrane could have been one of these volcanic arcs and might have served as one source of the K-bentonite ashes, possibly in concert with active arc magmatism on the Gondwana plate itself.

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GIVETIAN – FAMENNIAN PHOEBODONT ZONES AND THEIR DISTRIBUTION

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Phoebodont zones suggested by Ginter & Ivanov (1995) and traced in the different regions of the world presently need the adjustment. A first zone *Phoebodus sophiae* and accordingly the *Omalodus* assemblage occur beginning from the Middle *varcus* conodont Zone in the Givetian of Kuznets Basin, Siberia, Russia; Holy Cross Mountains, Poland; Germany; Mauritania; USA (Ivanov and Derycke, 1999; Turner et al., 2001). The latest occurrence of *Phoebodus sophiae* St.John & Worthen are known from the *disparilis* conodont Zone.

Next Frasnian zone, *Phoebodus latus* Zone corresponded to the interval of the Late *hassi – jamiaea* conodont Zones by the previous first find of index-taxon in the South Urals (Ginter & Ivanov, 1995). Recently *P. latus* Ginter & Ivanov was found in the vertebrate assemblage of *falsiovalis* conodont Zone in the Givetian/Frasnian sections of Middle Urals. The assemblage are represented by the microremains of turiniid thelodonts, chondrichthyan *Phoebodus fastigatus* Ginter & Ivanov, *P. latus*, *Wellerodus*, *Ohiolepis*, *Cladolepis*, placoderms, acanthodians, sarcopterygians and palaeniscoids. Thus, the lower limit of *P. latus* Zone and the boundary between the *P. sophiae* and *P. latus* Zones can be recognised at the base of the *falsiovalis* conodont Zone. The assemblage of this zone is distributed not so wide as the previous zone, it was found in the Middle and South Urals, Russia, and in the Holy Cross Mountains, Poland.

The latest Frasnian phoebodont Zone, the *Phoebodus bifurcatus* Zone, corresponds to the *rhenana – linguiformis* conodont Zones. The vertebrate assemblage of this zone occurs mainly in the *rhenana* Zone and are reported from the South Timan, Central Devonian Field, South Urals, Gorniy Altai, Russia; China, the Holy Cross Mountains, Poland; Moravia; Utah, USA.

The first Famennian *Phoebodus typicus* Zone was correlated with the interval of Late *triangularis – rhomboidea* conodont Zones. A first occurrence of *Phoebodus typicus* Ginter & Ivanov is varied in the different regions (Ginter, 2001): from the Late *triangularis* conodont Zone of South Urals, Russia; from the *marginifera* Zones of Queensland and probably New South Wells (Jones & Turner, 2000), Australia. It is possibly due to the unequal appearance of this species after the Kellwasser Event. *P. typicus* was found probably earlier than in the Late *triangularis* Zone in the Tom' River section, Kuznetsk Basin, Western Siberia, Russia. Perhaps this zone rises from the lower stratigraphic level.

Phoebodus gothicus Ginter, an index-taxa of next zone, was found possibly earlier than was known, in the *crepida* Zone from Iran (Ginter, 2001). The boundary between the *P. typicus* and *P. gothicus* Zones could be below the base of *marginifera* conodont Zone. The assemblage of *P. gothicus* Zone is wide distributed in the *marginifera – postera* interval of South Urals, Russia; the Holy Cross Mountains, Poland; Iran; Morocco; Iowa and New Mexico, USA.

The latest Famennian phoebodont *Phoebodus limpidus* Zone corresponds to the *expansa* – Middle *praesulcata* conodont Zones. Their boundaries, as well those of the *P. bifurcatus* Zone, are unchanged as was suggested by Ginter & Ivanov (1995). The taxa of this zone occur in the South Urals, Russia; the Holy Cross Mountains, Poland; Thuringia, Germany; Montagne Noire, France; Carnic Alps, Italy; South China; Western USA.

Most of phoebodont zones have a wide palaeogeographic distribution, only the *P. latus* and *P. typicus* Zones are recorded in a few regions.

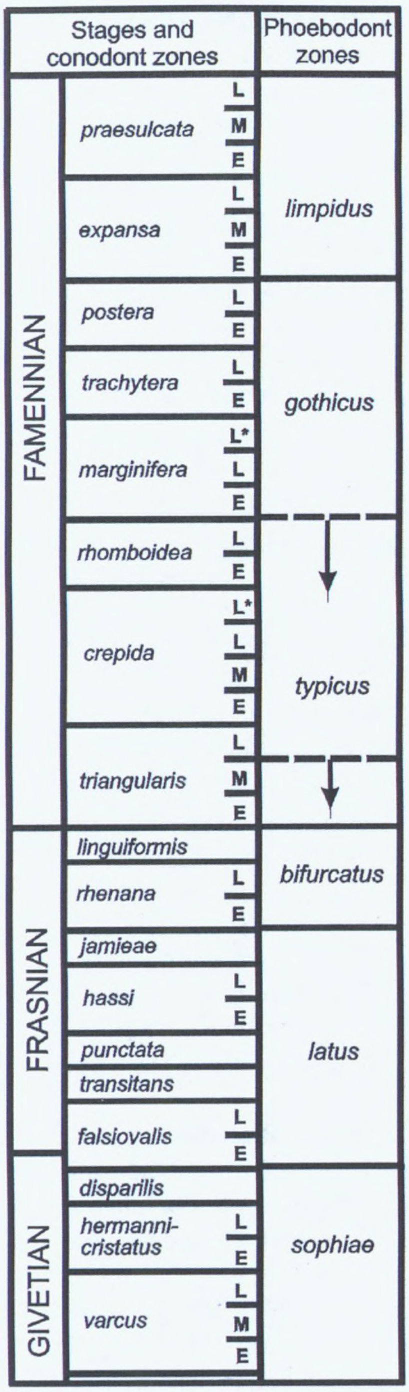

Figure. Phoebodont zonation of the Middle–Upper Devonian, based on Ginter & Ivanov (1995), modified according to a new data; arrows – possible change of zone boundary.

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LITHOSTRATIGRAPHIC SUBDIVISION OF THE VENDIAN DEPOSITS IN LITHUANIA

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Vendian sediments are distributed only in eastern part of Lithuania and rest on a crystalline basement. They were accumulated in the western marginal zone of the vast marine basin of the Moscow Syneclise. They consist of continental, lagoon and shallow marine clastic sedimentary rocks.Their thickness gradually increases eastwards from 0 to 200 m. (Jankauskas, Sakalauskas, 1997). Any palaeontolocical data are known in Vendian sediments of Lithuania. Therefore stratigraphical subdivision and correlation are based on lithological criteria.

A lithostratigraphic classification of Vendian sediments was worked out by V. Sakalauskas (1966). In a stratigraphic ascending order he recognized some formations. The oldest Merkys Formation is composed of the hillside wastes and proluvium ("Fanglomerate") and is correlated with the Wolyn Group in East European Platform. The younger "Arcose Series" are subdivided by V. Sakalauskas into Jašiūnai and Rūdninkai formations and recently united by T. Jankauskas (1993) to one Šalčia Formation (with two subformations). This interval is correlated with the lower part of the Valdai Group. Upper part of the Vendian deposits ("Sandysilty Series" of V. Sakalauskas) later was subdivided into the Skynimai and Vilkyškiai formations and correlated with upper part of the Valdai Group of East European Platform. T. Jankauskas (1993) proposed to consider two mentioned units as subformations within the Kleva Formation. According to him, the Vilkyškiai unit may be interpreted as a weathering crust on the top of the Vendian strata.

Recently T. Jankauskas and V. Sakalauskas (1997) suggested the more precise lithostratigraphical subdivision into two local groups and formations. The older Šalčia Group occupies the Merkys, Jašiūnai and Rūdninkai formations (last comprises only lower part of the former Rūdninkai Formation). Younger Kleva Group is subdivided into the Pagiriai (upper part of the former Rūdninkai Formations), Skynimai and Vilkiškės formations. Both Šalčia and Kleva groups have different lithological characteristics and slightly different geographical areas and are delimited by clear erosion gap (Table). The stratotipe of the new Pagiriai Formation is indicated in the borehole Vilkiškės-66, depth interval 411.5–432.8 meters.

As a whole the Vendian deposits of Lithuania from a thick rhythmically composed sedimentary cycle reflects two stages of sedimentation: Šalčia and Kleva groups. Each of them differs by its lithological features, but obvious predominance of clastic material is characteristic of the whole sequence. Furthermore the medium size of the clastic grains gradually decreases upwards throughout the section. The sorting of grains and maturity of rocks improve in the same direction.

The oldest Merkys Formation rests directly on the crystalline basement (Lower Proterozoic), measures 0–25 meters and is made up generally of coarse clastic material. In the stratotype area (southeastern Lithuania) the main component is a continental red-colored clayey sandstone-fanglomerate facies. In the basalian layers predominates fanglomerate. It is massive clays with abundant pebbles and sandy material. The median diameter of the pebbles is 1–3 cm, sometimes 5–7 cm. They are poorly rounded and consist of the granite, amphibolite, quartz or feldspar. Pebbles and sandy grains are covered with thin hydrous iron oxide film. The size of pebbles gradually decreases towards the top of formation and lateraly towards the northeastern part of Lithuania. In the Drūkšiai Lake area Merkys Formation consists of the coarse-graned sandstone, siltstone and gritstone. The color of rocks is reddish, brown, darkly-brown. The main components of clastic minerals are quartz, feldspar and mica. Separated layers include a abundant pebbles (up to 5 cm) of rocks of the some composition as the crystalline basement. Grain size in the sandstone ranges generally between 0.1–2–3 mm, but grains more than 1 mm are the most frequently occurring forms. The sandstone is badly sorted, roundnes ranges between angular. The quantity of quartz in sedimentary layers is greater than in underlying crystalline basement. The cementing material consists of clay, primary hydrous iron oxide and secondary dolomite and gypsum.

A stratigraphical position of the Merkys Formation is not clear. Traditionally we correlate this formation with the Volyn Group of the East European Platform (Resheniya..., 1978).

The younger strata are subdivided into Jašiūnai and Rūdninkai formations. As it was mentioned above, it conformably convers more older sediments, thus it is difficult to recognize the base of this units in the core material. Both formatios are two stages of arcose sedimentation and consist of alternating sandstones and siltstones of brown, red- brown color from hydrous iron oxide. Its structure is rhythmical. Individual rhythms are 0.1–5 m thick, their bases are sharp. They are composed of coarse-grained sandstone or gravelite at the bottom and of siltstone or clayey siltstone at the top. Most of them are parallel-laminated, occasionally they are cross-stratified. The bedding is usually distinct because of granulometric stratification.

Sandstones consist generally of scattered grains of quartz (45–63%), feldspar (10–25%) and mica (1–10%). Numerous grains of dark hevy minerals can be recognized on the surface of separated layers. The consolidation is poor, sorting is bad. Grain size ranges from 0.1 to 2–3 mm, with 0.3–0.4 mm fraction pervailing. The medium size of clastic components decreases upwards. The cementing minerals are as follows: kaolinite, chlorite and secondary dolomite and gypsum. Mixed-layered minerals occur in the Jašiūnai Formation.

Siltstones interbedded with sandstones and clayey layers are thin horizontally bedded and well consolidated. Their mineralogical composition is similar to that of sandstones. Thin clayey layers consist of illite, kaolinite, chlorite. The quantity of clayey layers increases upwards. The topmost layers (0.2–0.3 m) of the unit display traces of subaeral weathering. They are white of color and contain a big quantity of the kaolinite.

The unconformably overlying Pagiriai Formation forms upper part of the "Arcose Starta", but is separated by the hyathus from the older strata and have some wider area. Formation consists of the some rocks and reaches its maximum thickness of 21.3 m.

The Skynimai Formation conformably overlies older strata and is clearly different, by its granulometry. The unit consist of brown, red-brown or gray thin laminated sandstones, siltstones and occasionally claystones. The distribution of the interbedded silty and sandy layers varies throughout. The thickness of individual layers are normal 1–3 cm, occasionally till 10 cm. The clastic component consists of quartz, feldspar, mica. The cement basalian consists of illite, kaolinit, chlorite, iron oxide and secondary dolomite and gypsum. Sorting appears poor, grains are angular or subrounded. The most upper layers are weathered, they are light-yellow because of secondary hydrous iron oxide and kaolinite.

The Vilkiškės Formation completes the Vendian section of the Lithuania and consists of white, greenish-gray, occasionally multicolored siltstones, quartzitic sandstones and claystones. The thickness of the unit ranges between 15–23 meters. Thickness of the individual layers ranges in 2–5 cm, sometimes thin laminated layers occur. The unit displays horizontally laminated structure. Clastic component consists of quartz, mica, orthoclase and microcline. Grain size of the sandy layers ranges between 0.1–0.2 mm. The outlines of the greatest grains show evenly angular, subangular and rounded shape. However separated layers contains gravel material consisting of subrounded quartzous grains of 2–5 mm. The cementing material and clayey layers consists of kaolinite, illite, rarely carbonate. All rocks of the unit display a high grade of weathering. This deep weathering crust was formed during the pre-Baltian continental break. Only separated layers are not weathered, they are brown or greenish-gray colored and similar to the older strata. As it was mentioned above, this unit can be considered to be a weathering crust of the Vendian deposits.

Table. Lithostratigraphic subdivision of the Vendian deposits in Lithuania.

<table>
<tr><th colspan="2">Sakalauskas, 1968</th><th colspan="2">Resheniyja ..., 1976</th><th colspan="3">Jankauskas, Sakalauskas, 1997</th></tr>
<tr><th>Group</th><th>Lithostratigraphic classification</th><th>Group</th><th>Formation</th><th>Regional stage of the East European Platform</th><th>Group</th><th>Formation</th></tr>
<tr><td rowspan="4">Valdai</td><td rowspan="2">Sandy-silty Strata</td><td rowspan="5">Valdai</td><td>Vilkiškės</td><td rowspan="3">KOTLINIAN</td><td rowspan="3">Kleva</td><td>Vilkiškės</td></tr>
<tr><td>Skynimai</td><td>Skynimai</td></tr>
<tr><td rowspan="3">Arcozic Strata</td><td rowspan="2">Rūdininkai</td><td>Pagiriai</td></tr>
<tr><td>REDKINIAN</td><td rowspan="3">Šalčia</td><td>Rūdininkai</td></tr>
<tr><td rowspan="2">Volyn</td><td>Jašiūnai</td><td rowspan="2">DREVLIANIAN</td><td>Jašiūnai</td></tr>
<tr><td>Fanglomerate Strata</td><td>Volyn</td><td>Merkys</td><td>Merkys</td></tr>
</table>

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THE IMPORTANCE OF POLLEN AND DIATOM ANALYSIS DATA FOR STRATIGRAPHY OF LATE GLACIAL AND HOLOCENE DEPOSITS IN VARIOUS PARTS OF SOUTHEASTERN BALTIC

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Studies presented in this paper focus on the biostratigraphy of the Late Glacial and Holocene deposits from various parts of the SE Baltic. The main aims of the study are: 1) to look over the composition of pollen in deposits from various parts of SE Baltic, to correlate palynological data from investigated cores in this area and on the basis of the pollen data to divide up the off shore Late Glacial and Holocene deposits (Table 1) and the sediments from coastal zone of Lithuania and Kaliningrad Region (Kabailienė, 1999a); 2) to discuss the importance of diatom analysis data in stratigraphical studies of Late Glacial and Holocene deposits of the Baltic. As a rule diatom analysis data are using for reconstruction of palaeoecological conditions in diverse basins. The authors studies on diatoms from deposits of Baltic during many years confirm, that these data are especially significant for biostratigraphy (Table 1). The diatom data often are more usefull for subdivision the deposits of Baltic Sea than pollen data.

Some tens sediments core with pollen and diatom data are included in the studies presented here. The cores are located at different water depth in various parts of Gdansk and Gotland depressions, Kuršių – Sambian plateau, Kuršių Nerija Spit, Kuršių Marios Lagoon, on the territory of Nemunas delta and so on.

In the paper diatom analysis data are obtained mainly by author. Pollen analysis from deposits of the deep depressions has been done by O. Kondratienė (Kabailienė et al., 1978 and 1990), from littoral sub-aquatic zone – mainly by D. Ūsaitytė (2001). The pollen analysis data from deposits of many deep cores on the Spit of Kuršių Nerija, Kuršių Marios Lagoon on the Southern coast of this Lagoon in Kaliningrad Region, at Šventoji settlement and Šventelė peatbog has been done by author. Pollen in the deposits of cores on the territory of Nemunas Delta, some cores from settl Šventoji and two new cores in Šventelė peatbog were studied by M. Stančikaitė (Kabailienė et al., 1998, Stančikaitė and Kabailienė, 1998).

POLLEN ANALYSIS DATA. OFF-SHORE DEPOSITS

All pollen spectra from the off-shore deposits cores may be characterized by the following general features: a) predominance of arboreal pollen (up to 99%) and prevalence of *Pinus* pollen (60–97%) among the main tree taxa; b) low amounts of non-arboreal (about 5%), increasing in the Late Glacial and Subatlantic parts of the diagrams (up to 10–20%), and predominance of only a few herb taxa (*Artemisia*, Poaceae, Cyperaceae, Chenopodiaceae and Asteraceae); c) pollen from plants indicating human impact, such as Cerealia, is registered in the uppermost sediments of the cores beeing close to coast; d) rebedded pollen grains and spores of Jurassic and Cretaceaus age occur in the lower parts of several cores (up to 20% and more); e) pollen curves have a fairly smooth and monotonous character through the diagrams.

Pollen spectra of off-shore deposits represent a large "catchment area". Interpretation and subdivision of pollen diagrams from the off-shore sediments are complicate and difficult.

Table 1. Stratigraphic subdivision of Late Glacial and Holocene off-shore deposits in southeastern part of the Baltic Sea (compiled by M. Kabailienė, 2002)

Chronological scale in ka BP	Stage	Substage	Chronozone	Index	Palynozone	Baltic Sea stages	Index	Characteristic of diatomic complexes	Definition of boundary in ka BP
	Holocene	Upper	Subatlantic	SA	Pinus-Picea-Alnus-Betula	Postlitorina (Limnea)	PL (Lim)	High concentration of diatoms, brackish and marine forms of littoral and deep zone prevail	
2,5; 4,0		Middle	Subboreal	SB	Picea-Carpinus-Quercus-Pinus				4,0
5,0; 7,8			Atlantic	AT	Alnus-Tilia-Ulmus-Quercus-Corylus	Litorina	L	High concentration of diatoms, prevail marine and brackish diatoms of deep zone	8,0
8,1		Lower	Boreal	BO	Pinus	Ancylus	A	High concentration, freshwater oligotrophic of deep zone diatoms	
9,0			Preboreal	PB	Betula-Pinus-Artemisia	Yoldia	J	Middle concentration, brackish, halophilous and freshwater diatoms of littoral zone	9,5; 10,0
10,0	Pleistocene	Late Glacial	Younger Dryas	DR3	Betula-Artemisia	Baltic Ice Lake	BIL	Low concentration of diatoms, freshwater oligotrophic, diatoms of deep zone prevail	
10,9			Alleröd	AL	Pinus				
11,9			Older Dryas	DR2	Artemisia-Chenopodiaceae-Pinus-Betula				12,0
12,3			Böling	BÖ	Betula-Pinus	Local Ice Lakes	lg	Very low concentration of diatoms, freshwater forms prevail	

COASTAL DEPOSITS

Pollen spectra from the deposits of cores on the coast (Kabailienė, 1999a) differ from those from off-shore: a) although arboreal pollen prevails but values of *Pinus* pollen are considerably lower; b) amounts of non-arboreal pollen increase greatly not only in Late Glacial deposits but in deposits of Subboreal--Subatlantic chronozones; c) in pollen diagrams the curves of *Picea*, *Alnus*, *Betula*, *Tilia*, *Ulmus*, *Quercus*, *Carpinus* and *Corylus* oscillate remarkably and often show well-defined peaks.

These pollen spectra reflect regional and local environments and vegetation.

DIATOM ANALYSIS DATA

In each studied site of Late Glacial and Holocene deposits from the Baltic, diatom species were grouped according to various criteria: salinity, living conditions and others. The changes in quantity of these ecological groups of diatoms are significant for stratigraphical conclusions. The data on concentration in deposits of diatoms were used too (Table 1).

In the layer of Baltic Ice Lake concentration of diatoms are low, prevail freshwater planktonic oligotrophic species. *Aulacoseira islandica* morph. *helvetica* were prevalent. In near-shore zone in some cases dominants were *Opephora martyi*, *Navicula scutelloides*, *Fragilaria inflata*. The freshwater Baltic Ice Lake transition to the weakly brackish Yoldia Sea was rather synchronous and occurred quite quickly at 10 ka BP. Yoldia Sea deposits have not been detected on the coastal zone of Lithuania and Kaliningrad region, since the water level was very low. The salinity and sedimentary conditions seem to have varied greatly. This is reflected in the changes in the quantities of brackish and marine diatoms which vary from single frustules (Gdansk depression) to 20–40% (Gotland depression, the Gulf of Riga and the SE littoral submarine zone). Halophilous (*Stephanodiscus rotula*, *Cyclotella meneghiniana*, *Epithemia sorex*, *E. turgida* and others) are the dominant species in many cases. This interval of deposits contains freshwater species too.

Freshwater diatoms characteristic of large oligotrophic water bodies prevail in the Lake Ancylus deposits of all the sites studied. Concentration of diatoms is high. During the increase of Lake Ancylus level (first half of the Boreal chronozone) planktonic species were found to prevail, whereas during the regression (second half of the Boreal) benthic and epiphytic diatoms were more abundant. The transitional stage between Ancylus Lake and Litorina Sea characterized by a low salinity, is known as the Mastogloia Sea stage (Hyvärinen, 1988). However this stage has not been defined in off-shore facies (Ignatius et al., 1981, Kabailienė, 1999), where it is referred to as the Initial Litorina Sea (Andrén, 1999). In SE Baltic on the coastal area in several sites were distinquished initial layer of Litorina Sea with characteristic brackish diatoms.

To the deposits of Litorina Sea are characteristic: a) very high concentration of diatoms; b) clearly prevailing marine and brackish species and c) changes in dominating diatoms (planktonic forms prevail in deposits of deep zone of the Sea, while periphytic ones are most common in the near-shore zone). Very characteristic marine and brackish species are of genera *Chaetoceros*, *Actinocyclus*, *Coscinodiscus*, *Campylodiscus*, *Nitzschia* and others. There is various difficulties in distinquishing the chronological limit between Litorina and Postlitorina.

The transition between the Litorina and Postlitorina Sea stages is drawn where the siliceous microfossil assemblages that require more marine environment decrease but it is not clearly defined (Hyvärinen, 1988, Kabailienė, 1999b). This transition is time-transgressive, being placed at ca. 4.0 ka BP in Finlan, Estonia and Lithuania (Hyvärinen, 1988, Kabailienė, 1999a, b), at ca. 3.5 ka BP in the Gotland basin and ca. 2.6 ka BP in SW Baltic Sea (Andrén et al, 2000), ca. 2.0 ka BP in Eastern Danmark (Krog, 1979). To Postlitorina deposits of SE Baltic characteristic dominant species is brackish *Actinocyclus octonarius*.

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CORRELATION POTENTIAL OF THE CARBON ISOTOPE TRENDS IN THE UPPER ORDOVICIAN AND SILURIAN SEQUENCES IN THE BALTIC AREA AND OTHER REGIONS

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Stable isotopes, carbon and oxygen in particular, have become popular in different fields of geology, including correlation of sequences and elucidation of environmental parameters. During the last ten years many interesting and useful papers have been published (here we refer only to a few of those discussing also stratigraphical aspects of the topic in the framework of the Ordovician and Silurian: Andrew et al. 1994; Marshall et al. 1997;

Saltzman 2001; Underwood et al. 1997, etc.). The aim of this report is to demonstrate the correlation potential of carbon isotopes, how these have been and can be used in stratigraphy and which conditions should be considered.

We are glad to note that during a few last years huge progress has been made in the Baltic, Germany (papers on Gotland), UK and USA, but also in China, Russia, Spain, South America, etc. Thanks to the new publications we can state now that we know the general pattern of the late Ordovician and Silurian carbon isotope ($\delta^{13}C$) trend.

The following main $\delta^{13}C$ positive pre-Hirnantian excursions have been determined in the late Ordovician: one in the mid-Caradoc (value reaching 2.2‰, Ainsaar et al. 1999), two in the late Caradoc and one in the early Ashgill (Kaljo et al. 1999) in Estonia and several in the Mohawkian sections in North America (Patzkowsky et al. 1997; Ludwigson et al. 2001, Fanton & Holmden 2001; etc.). The Mohawkian excursions are partly only preliminarily published (values 2–3‰), and their exact positions in terms of graptolite biozonation are not known. Nevertheless, the occurrence of these excursions shows that the small $\delta^{13}C$ positive shifts (values close to 2‰), identified in Estonia in the Caradoc and early Ashgill, might have their counterparts in North America. If these are biostratigraphically properly dated in Laurentia as well as in Baltica, then several complicated problems of correlation, sedimentary history and palaeogeography involving both continents would be much better solved through simultaneous application of carbon isotope data.

A good illustration for the above statement is the Hirnantian excursion recognized in many regions of the world. The Baltic material (Kaljo et al. 2001), supported by Dob's Linn, Scotland (Underwood et al. 1997), and Monitor Range, Nevada (Finney et al. 1999), allowed Brenchley et al. (submitted) to propose a composite model for the Hirnantian carbon isotope trend. The $\delta^{13}C$ values began to rise slowly in the uppermost Rawtheyan (*Belonechitina gamachiana* chitinozoan Biozone) but more rapidly in the lower Hirnantian (*Spinachitina taugourdeaui* Biozone) and reached peak values (~7‰) in the *Conochitina scabra* Biozone (= upper part of the *Normalograptus extraordinarius* graptolite Biozone). The *N. persculptus* Biozone shows a slightly falling plateau and then a falling limb until the beginning of the Silurian when the values return to the pre-Hirnantian level. Together with the rising limb of the carbon isotope excursion the *Hirnantia* community appeared in the Baltic area. As shown by Rong et al. (1999), this characteristic relatively cool-water assemblage persisted also during *N. persculptus* time in the conditions of the pre-Silurian sea level rise and partly even at the very beginning of the Silurian.

Comparison with the composite model allows us to analyse different carbon isotope curves based on local sections, correlate them and decide upon their completeness or to identify gaps in the sequences.

The general Silurian carbon isotope trend seems to be more or less well established (Azmy et al. 1998, Kaljo et al. 1998, Saltzman 2001, etc.) even if most of the publications discuss the Wenlock and Ludlow $\delta^{13}C$ positive excursions in Australia, East Baltic, Gotland, Scania and North America where three major excursions (lowermost Wenlock, upper Homerian, lower Ludfordian) are well determined. The Llandovery trend has been noted in a few papers (Azmy et al. 1998) and in more detail discussed by the present authors (Kaljo and Martma 2000).

Considering both the stratigraphical correlation and environmental interpretation of the isotope data, the exact biostratigraphic dating of the excursions is highly crucial. According to Estonian and Latvian data the rather wide lowest Wenlock excursion began with a slow rise of $\delta^{13}C$ values at the very end of the Telychian; a rapid rise followed in the *Cyrtograptus centrifugus – C. murchisoni* biozones and peak values in the *Monograptus riccardonensis* Biozone. These Baltic data are in harmony with Gotland (Bickert et al. 1997; Azmy et al. 1998) and Pete Hanson Creek, Nevada (Saltzman 2001), data. The latter and especially the Highway 77 section, Oklahoma, seem to be much more condensed, making the $\delta^{13}C$ curve rise and fall more steeply.

In the Baltic area the Homerian carbon isotope shift is well represented only in the Ohesaare core (Kaljo et al. 1997), in Gotland and Wales (Corfield & Siveter 1992) the excursion is rather low. The Laurentian Pete Hanson Creek section, Nevada (Saltzman 2001), shows a prominent well-dated (*M. ludensis* Biozone) excursion supporting the global significance of the event. Considering the Baltic data, where lowstand of the $\delta^{13}C$ curve after the Homerian peak occurs in the *Neodiversograptus nilssoni* Biozone, we can suppose that in the Nevada section the lower boundary of the *N. nilssoni* Biozone should be dropped until the bottom of the thick limestone beds occurring above the peak.

The Early Ludfordian shift is extraordinary due to very high $\delta^{13}C$ values in Gotland (Bickert et al. 1997) and Scania (Wigforss-Lange 1999). Quite "normal" values for a peak in Latvia (Kaljo et al. 1997), Nevada and Oklahoma support Saltzman's (2001) explanation that some local reason might be responsible for this anomaly. This is not fully true, but still some connection seems to exist with the pattern we have observed in the early Wenlock where along with the deepening of the sea from Viki to Priekule drill cores $\delta^{13}C$ values decrease from 5.2 to 3.1‰ (Kaljo et al. 1998).

In conclusion, we would like to repeat the statement we made above: carbon isotope trend could serve as a tool for correlation, especially of sections of different facies, but the best results will be achieved when carbon isotope data are applied together with biostratigraphic criteria.

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DISTRIBUTION OF VERTEBRATES IN UPPER ORDOVICIAN – LOWER DEVONIAN OF TUVA (RUSSIA)

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Microremains of vertebrates from different levels of O_3–D_1 sequences of Tuva were collected by T. Moskalenko, L. Aksenova, N. Kulkov, and L. Ratanov (Novosibirsk) in period of 1968–1987 as accessory material during the investigations of conodonts. Special sampling of vertebrate remains in the deposits of Upper Silurian – Lower Devonian (Elegest and Kadvoj sequences) was carried out by V. Karatajūtė-Talimaa as a member of the research expeditions in 1968 and 1971 (leaders E. Vladimirskaya and A. Krivobodrova – St. Petersbourgh).

The descriptions of *Elegestolepis grossi* Kar.-Tal. and thelodonts of Tuva were published in 1973 (Karatajūtė-Talimaa, 1973) and 1978 (Karatajūtė-Talimaa, 1978).

The following six vertebrate assemblages can be established for the O_3–D_1 sequence of Tuva (see Table).

The most ancient assemblage is very poor and is presented only by tesserae of new genus *Propatoraspis*, belonging probably to Tesakoviaspidida, the new order of Agnatha distributed in the O_3–S_1 of Siberian Platform.

II–V assemblages are presented by microremains – scales of thelodonts, chondrichthyans, mongolepids, acanthodians, anaspids, osteostracans and galeaspids (?). The four local thelodont biozones can be distinguished (Talimaa, 2000): Loganellia asiatica (assemlages II and IIa), Loganellia kadvoiensis (assemblage III), Helenolepis obruchevi = Loganellia tuvaensis (assemblage IV) and Helenolepis navicularis (assemblage V).

In the terrigenous deposits of the lower part of Khondergei Regional Stage are distributed only scales of *Elegestolepis grossi* and remains of exoskeleton of Osteostraci (assemblage VI).

The last publications with definition of the age of all formerly recognized biostratigraphical units (Regional Stages or Beds) were carried out by Kulkov et al., 1985 and Vladimirskaya et al., 1986. After the authors of these publications the S/D boundary pass between Tauganteli and Khondergei Regional stages and to Přidolian Stage are attributed the upper part of Pitchi–Shui and Tauganteli Regional stages.

After the opinion of L. Ratanov the deposits of Přidoli Stage are absent in the Silurian sequence of Tuva. In that case the S/D boundary must be draw between the Dashtygoi and Pitchi–Shui Regional stages. Such statement confirm the findings of conodonts in the deposits of Pitchi–Shui Regional Stage: *Pelekysgnathus serratus elatus* Carls et Gandl., *Spathognathodus* cf. *steinhornensis* Ziegler (determination of T. Moskalenko), *Icriodus woschmidti* Ziegler (determination of L. Aksenova).

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Table. Stratigraphic distribution of vertebrates in the Upper Ordovician – Lower Devonian of Tuva.

Age after Vladimirskaya et al., 1986			Age after Ratanov L.S. (new material)	Age after Vladimirskaya et al., 1986								
System	Series	Stage	Stage	Regional Stage or Beds	Nos of assemblages	Vertebrate microremains assemblages	Local thelodont biozonal scheme (Talimaa, 2000)		Localities Nos of samples			
									Elegest	Pitchi-Shui	Tchadan	Kadvoj
Devonian	Lower	L	D_1 Lochkovian	Khondergei	VI	*Elegestolepis grossi* Kar.-Tal., Osteostraci indet.		↑	409, 415, 295, 293, 291, 286-1	68–51-P 68–51 68–52	(Bazhin-Alaak)	675 674
Silurian	Upper	Pr		Tauganteli	V	*Helenolepis navicularis* Kar.-Tal., *H. multicostata*? Kar.-Tal., *Elegestolepis grossi* Kar.-Tal., *Elegestolepis*? gen. nov. Osteostraci indet. Anaspida gen. nov.	*Helenolepis navicularis*		276, 276-1 272-3, 272 271 256		125 R,119 R K-76-11	672 670-1 670
		Pr / Ld		Pitchi-Shui	IV	*Loganellia tuvaensis* (Kar.-Tal.), *Helenolepis obruchevi* Kar.-Tal., *H. multicostata* Kar.-Tal., *L.? kadvoiensis* (Kar.-Tal.), *Elegestolepis grossi* Kar.-Tal., *El.* n.g.? Osteostraci? ind., Anaspida n. gen.	*Helenolepis obruchevi* = = *Loganellia tuvaensis*	*Elegestolepis grossi*	249, 247, 244, 243, 239, 238, 236		29 R, 16 R K-70-87 K-70-84	668 665
		Ld		Dashtygoi U.	III	*Loganellia*? *kadvoiensis* (Kar.-Tal.), *Polymerolepis* sp. nov., *Udalepis forata* g. et sp. nov., Galeaspida??, Acanthodei gen. ind., Osteostraci? ind.	*Loganellia* ? *kadvoiensis*		226–224			664
	Lower	W		Dashtygoi L.	IIa	*Loganellia asiatica* (Kar.-T.), *Polymerolepis* sp. nov., *Elegestolepis grossi*? Kar.-Tal., *El.* sp. nov., Acanthodei g. ind.		↑				663 662
		W		Akchalym			?					
		Ln_3		Angatchi	II	*Loganellia asiatica*? (Kar.-Tal.), *Helenolepis*? sp. ind., *Polymerolepis* sp. nov.	*Loganellia asiatica*					660-4 660-1
		Ln_3	Ln_3	Kyzyl-Tchiraa	II	*Loganellia asiatica* (Kar.-Tal.), *Tchunacanthus* sp. nov.? (cf. *obruchevi*)	*Loganellia asiatica*		K-70-134		Kyzyl-Tchiraa 702 271 R 253 R	
		Ln_{1-2}		Alash	II	*Loganellia asiatica* (Kar.-Tal.)	*Loganellia asiatica*					694
Ordovician	Upper	Ash		Khondelen	I	*Propatoraspis juncta* (n.gen. n. sp.)			Khondelen 68-3v, 68-4-1			

THE DETAILED STRATIGRAPHY OF TURONIAN OF BELARUS ABOUT FORAMINIFERS

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Turonian deposits on the territory of Belarus have a wide distribution. They are present by mudstone – chalky deposits, thickness of witch varies from 30 m in the northwestern part of Pripyat Depression to 122 m in the southwestern of the Podlasie–Brest Depression. According to foraminiferous composition Turonian stage is divided into two substages. Zone Gavelinella nana corresponds to the lower stage. It is subdivided into subzones "Bolivina" kushensis (lower) and Tappannina simplex. General species for both subzones are: Spiroplectammina cuneata, Gaudryina arenosa, G. folium, G. serrata, Globorotalites turonicus, Gavelinella nana.

The lower subzone differs by presence of species Brotzenella aff berthelini, Cibicides lepidus, "Bolivina" kushensis, Tappannina aff eouvigeriniformis extinction. The lower border of the subzone Tappannina simplex is defined by the appearance of species Gaudryina angustata, Eponides turonicus, Gavelinella ammonoides, G. kelleri dorsoconvexa.

Zones Gavelinella ammonoides and Gavelinella moniliformis correspond to the upper substage of the Turonian. In the structure of the lower zone subzones Reussella turonica (lower) and Reussella carinata are present. Subzone R. turonica is characterized by the presence of species Gaudryina arenosa, Globorotalites turonicus, Sitella aff gracilis, Reussella turonica, Eouvigerina aff regularis, Tappannina simplex.

Subzone Reussella carinata is defined on the basis of the appearance of species Gaudryina laevigata, Globorotalites multiseptus, Reussella carinata, Citella angustata and mass occurrence Gavelinella moniliformis.

For the first time the given subzones were established by V. S. Akimez (1981) and traced by us on the most part of the territory of Belarus. It allows to use given biostratigraphical divisions in the stratigraphical scheme of the Cretaceous adjoummen of Belarus.

Characteristic species of zone Gavelinella moniliformis are: Verneuilina muensteri, Heterostomella carinata, Gaudryina variabilis, Globorotalites multiseptus, Gavelinella moniliformis, Reussella kelleri, Globorotalites lapparenti, Tappannina eouvigeriniformis.

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CORRELATION BETWEEN DEEP AND SHALLOW SHELF ON THE BASIS OF O-BENTONITE, EAST BALTIC

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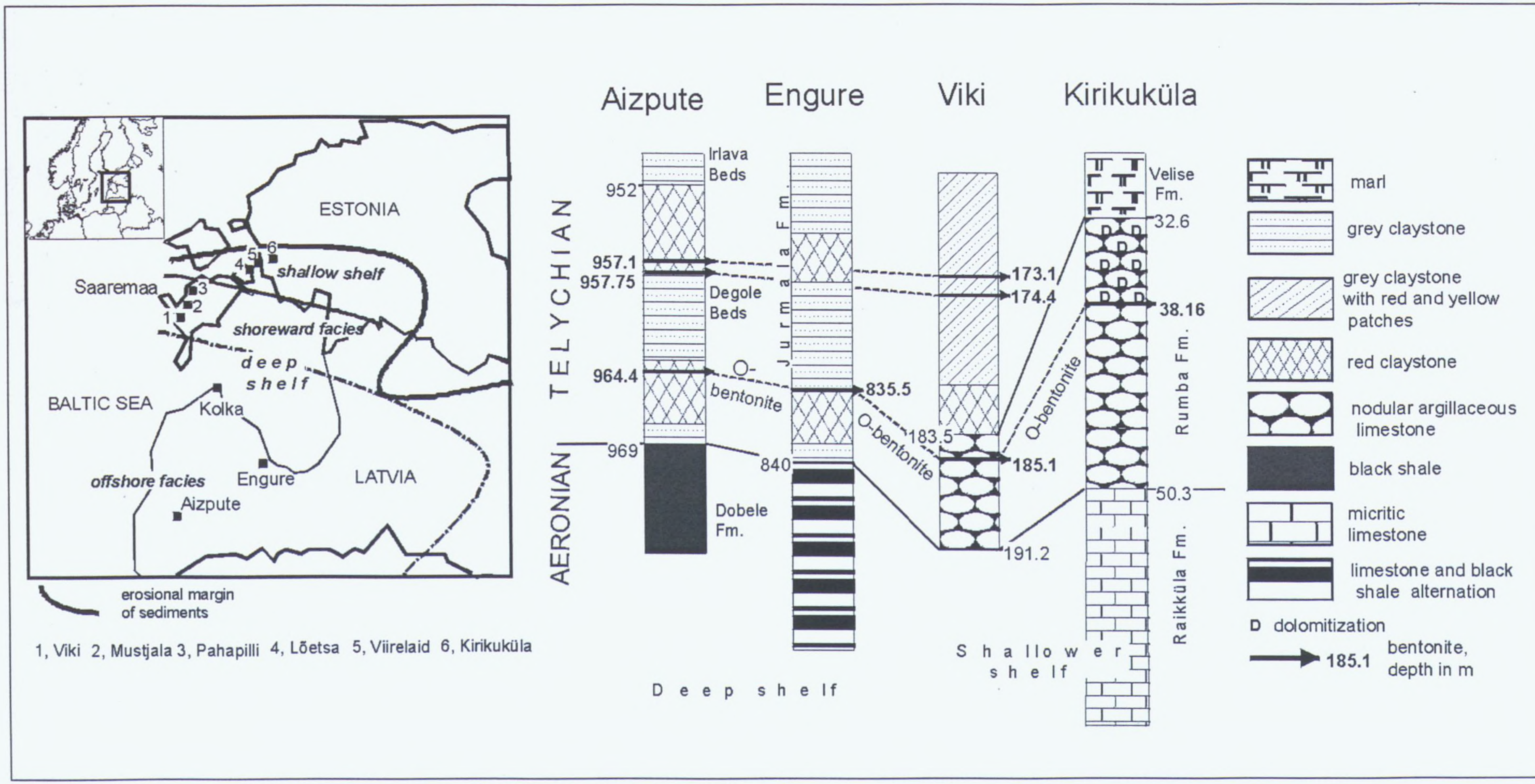
Bentonites (metabentonites, K-bentonites) are layers of altered volcanic ash within sedimentary rocks. In Estonia the thickest Silurian bentonite is O-bentonite ("opornyi", according to Einasto et al., (1972) or Osmundsberg bentonite, according to Bergström et al., (1998)), 5–25 cm thick. Its color is pale, yellowish or grey. In Estonia the O-bentonite is situated in the upper part of Rumba Fm., the lower formation of Telychian, Upper Llandovery, Lower Silurian (Fig. 1). O-bentonite is a good stratigraphic marker in rocks of shallower shelf due to its great thickness. In the rocks of deep shelf of West Latvia the thickness of O-bentonite decreases to 1 cm. Recently a method of Orville (1967) has been applied for correlation of the bentonites by Kallaste and Kiipli (in press).

The method bases on X-ray diffractometry (XRD) measurements of the concentration of $NaAlSi_3O_8$ in pyroclastic sanidine occurring in volcanic ash layers. For O-bentonite (Table) the characteristic content of Na-component in sanidine is 20.7–21.5 mol%. This sanidine composition is unique among Telychian bentonites of Estonia and Latvia. Two bentonite beds higher in the Aizpute and Viki sections (Fig. 1B) confirm the O-bentonite correlation. The lower layer, violet in colour, from depth 957.75 m from Aizpute core, with Na-sanidine mol% 26.7, correlates to the layer from the Viki core, depth 174.4 m, with Na- sanidine mol% 26.5. Upwards in the Aizpute core mottled violet and grey bentonite layer from depth 957.1 m, with $NaAlSi_3O_8$ mol% 39.6 correlates to red bentonite from the Viki core, depth 173.1 m, with $NaAlSi_3O_8$ mol% 40.6, and to the layer from the Kolka core (West Latvia) from depth 596.6 m with $NaAlSi_3O_8$ mol% 39.9. Some differences between values of Na mol% of the O-bentonite of the present abstract and paper by Kiipli, Kiipli and Kallaste (2000) occur. The difference, in maximal 1 mol%, is caused by application of new fitting and calculation technique (Kiipli and Kallaste, in press).

The analysis of trace elements of bentonite bulk material give supplementary data for identification of coeval beds. For O-bentonite low contents of Nb, Zr, Y, Rb, Sr, and Ti are characteristic (Table). Comparing



(A) (B)

Fig. 1. (A) Location of bore-holes; (B) Correlation of Telychian O-bentonite between cores from deep and shallow shelf.

Table. Main and trace elements of O-bentonite.

Core	Depth	$NaAlSi_3O_8$	SiO_2	Al_2O_3	MgO	CaO	K_2O	Fe_2O_3	TiO_2	Na_2O	MnO	LOI_{920deg}	Nb	Zr	Y	Ga	Rb	Sr	Ni
	m	mol%	%	%	%	%	%	%	%	%	%	%	ppm	ppm	ppm	ppm	ppm	ppm	ppm
Kirikuküla[1]	38.16	22.0	nd	19.8	0.6	0.40	15.30	1.5	0.25	nd	0.012	nd	4	167	6	16	66	75	4
Viirelaid	85.0	nd	62.9	18.5	0.5	0.33	14.76	1.9	0.36	nd	nd	0.97	6	207	7	15	60	74	0
Lõetsa[2]	66.05	21.8	65.4	17.6	0.5	0.27	15.33	1.3	0.33	<0.7	0.016	0.57	5.2	140	10.4	21	65.4	64.6	8
Pahapilli[3]	78.35	21.1	63.7	18.2	0.8	0.23	14.81	1.3	0.29	<0.7	0.022	0.84	7	144	9	17	72	60	18
Mustjala	124.8	nd	63.9	18.2	1.3	0.32	14.41	1.1	0.37	<0.7	0.011	0.94	nd	nd	nd	nd	nd	nd	nd
Mustjala[4]	124.8	nd	61.7	18.0	1.2	0.23	14.10	1.3	0.37	0.1	0.020	1.85	14	200	3	nd	nd	76	3
Viki[5]	185.1	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	11	181	4	16	92	111	12
Viki[6]	185.1	21.1	nd	20.9	2.5	0.50	12.10	2.5	0.40	nd	0.018	nd	9	194	7	18	95	106	11
Aizpute[3]	964.4	21.3	50.1	26.0	2.2	0.42	4.48	7.0	0.62	<0.7	0.029	7.32	15	255	10	nd	nd	nd	nd

[1]--main elements from Kiipli and Kallaste, 1996; trace elements -- average of 2 analysis from Kiipli et al., 2001

[2]-- trace elements, average of analysis from 5 different depths, (Kiipli et al., 2001)

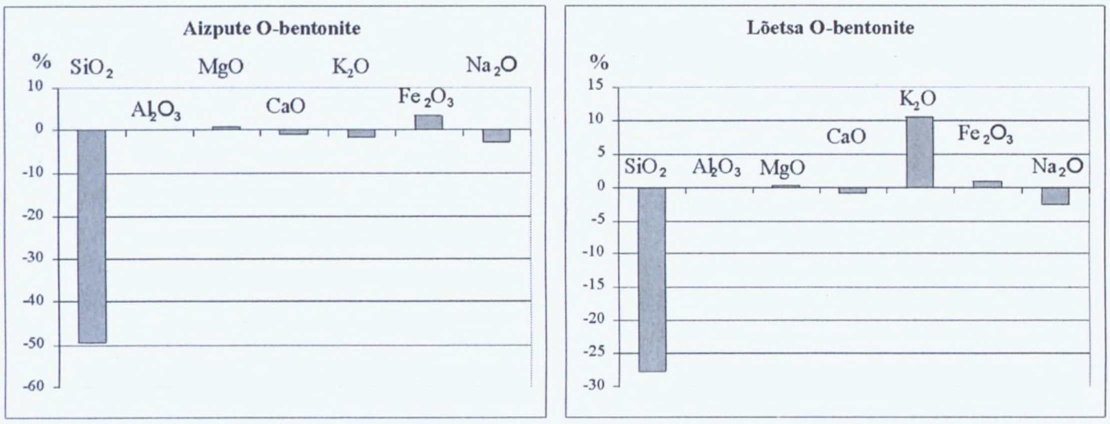
[3]-- main elements from Kiipli et al., 2000, trace elements from Kiipli et al., 2001

[4]-- elements from Huff et al., 1997

[5]-- elements analysed in St Andrews, data from Kiipli et al., 2001

[6]--main elements from Kiipli and Kallaste, 1996; trace elements from Kiipli et al., 2001

the O-bentonite layer with other bentonites from Saaremaa cores low trace element content comes obvious (Kiipli et al., 2001, fig. 6 and table 2). In the O-bentonite of the Aizpute core enrichment of Nb, Zr and Ti compared to O-bentonites from Saaremaa takes place. This can be explained by difference in chemical processes affecting the settling fresh volcanic material in deep and shallow shelf. In shallow areas authigenic K-feldspar in matrix of bentonites prevails (Kirikuküla and Lõetsa cores), with sea deepening the matrix becomes more illite-smectitic (Viki core). In the cores of deep shelf kaolinite is a common component of bentonite matrix. In shallow shelf external uptake of potassium probably promotes authigenic K-feldspar formation. In deep shelf removal of dissolved silica from volcanic material is more important (Fig. 2) and causes the higher residual enrichment in immobile chemical elements Nb, Zr, Y and Ti compared to shallow sea feldspathites.



(A) (B)

Fig. 2. Enrichment-depletion of main elements in O-bentonite from: (A) deep shelf represented by the Aizpute core and (B) shallow shelf, Lõetsa core compared to the composition of melt inclusions from quartz of the Osmundsberg K-bentonite (Huff et al., 1998).

Two different correlations between shallow and deep shelf sections have been published. The first correlates the Rumba Fm. of Adavere stage with *M. sedgwickii* zone of the upper part of the Dobele Fm., Aeronian. O-bentonite occurrence within Telychian *turriculatus* zone (Bergström, 1998) presumes the correlation of the Rumba Fm. with Lower Degole Beds, the Jurmala Fm., Telychian. Our correlation between Viki and Aizpute cores confirms the last one: the Rumba Fm. is of early Telychian age.

O-bentonite correlation between cores from shallower shelf and deep shelf shows spatial and temporal variation in the distribution of red beds (Fig. 1B). Red claystone of the Aizpute core (967.7–964 m), accommodating O-bentonite in the depth 964.4 m, corresponds to the light grey nodular limestone of the

Rumba Fm. in the Viki core and argillaceous, partly dolomitized limestone in the Kirikuküla core. Upwards, grey shale in the Aizpute core (964–958 m) correlates with the red and mottled section in the Viki core (183–176 m). Based on correlation of three bentonite layers (Fig. 1B), the earlier onset of red facies in the West Latvia compared to the Saaremaa is obvious. The emerge of red facies in West Latvia was a period of sea transgression, which took place after late-Aeronian regression. Moving of red facies to Saaremaa coincides with the beginning of Velise time, which marks the next pulse of transgression. The formation of red facies was supposedly initiated by down-welling bringing oxygen from surficial waters to the deeper parts of the sea. Bioproductivity decrease due to poor supply of nutrients by down-welling was the mechanism governing red bed formation.

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CORRELATION OF UPPER LLANDOVERY BENTONITES IN THE BALTIC REGION

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Volcanic ash interbeds in sedimentary sections extending through several facies zones offer uniquely perfect time planes for correlations. In recent sediments tephrochronology is widely used and correlations are proved by the composition of volcanic glass. In the Palaeozoic rocks no glass has preserved, original amorphous volcanic dust was altered completely to the clay minerals and authigenic potassium feldspar. Therefore, bulk composition or glass cannot be used for proving correlations. Fortunately some trace components and minerals have not altered. Consequently, analysing of immobile trace elements and some relatively stable trace magmatic minerals can help substantially to identify beds in different sections belonging to the same eruptive layer. Seven-year experience of our working group in this field is based on the study of about 280 altered volcanic ash samples mostly from the Upper Llandovery sections (less from the Ordovician and Wenlock) of the Baltic region (Kiipli, Kallaste 1996; Kiipli et al. 1997, 2001; Kiipli, Kallaste 2002 in press).

Using of immobile and other trace elements is particularly effective between sections within the same facies zone. Thereby it can occur that correlating between closely locating sections from different facies (Viki–Ohesaare) is more difficult than to far away sections (Estonia–Norway) characterised by similar facies. It points to the facies control over the alteration of volcanic ash. Figure 1 left chart shows changes of two immobile element (Nb and Ti) concentrations from shallow sea sediments in Estonia to deep sea facies on Sõrve peninsula and Latvia. Due to the bigger loss of silica and other main components in the formation

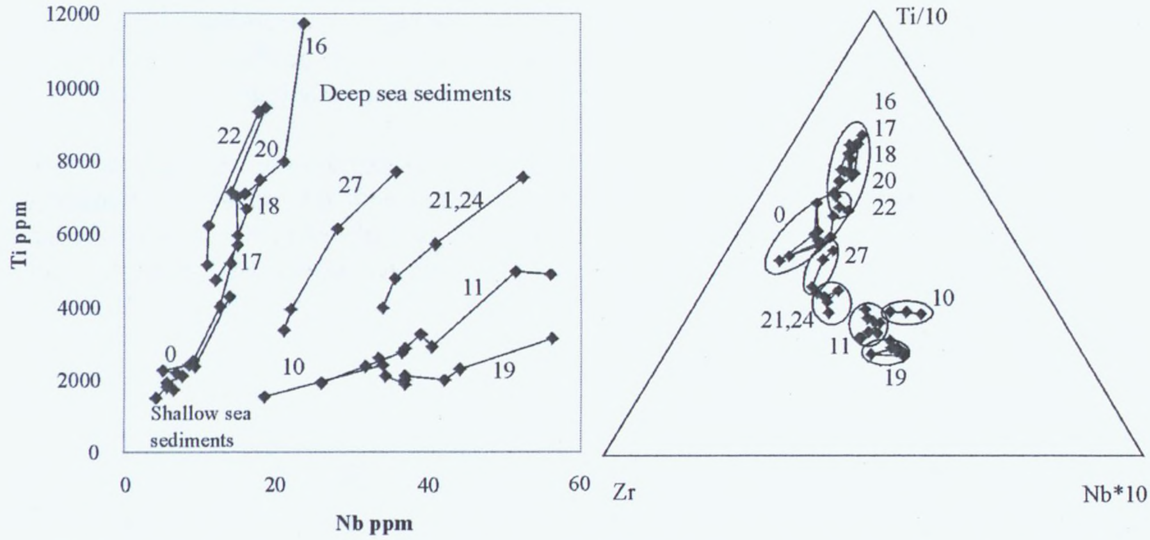

Fig. 1. Ti, Nb and Zr in some Upper Llandovery volcanic beds from Estonia and Latvia. Numbers correspond to the beds in Kiipli et al. 2001. Lines connect correlated beds from different sections.

of kaolinite in deep sea sediments there occurs the residual enrichment (2–3 times compared to the shallow sea facies) with immobile elements. The right triangular chart (Fig. 1) shows clusters of the same beds on the basis of Nb–Zr–Ti ratios. Using of immobile element ratios eliminates the diagenetic trends and enables better identification of beds. Useful elements are Nb, Zr, Ti, Th, Y, Sr and others. These elements can be analysed accurately by the X-ray fluorescence method.

Besides different alteration pathways in various facies some other factors complicate correlations on the basis of trace elements: 1. Late diagenetic changes (mobile elements must be excluded from consideration). 2. Fractionation of ash material during air transport and sedimentation. 3. Change of source magma composition during long-lasting eruption. These compositional trends can be studied by dense sampling or averaged by sampling of full thickness of bed. The first approach is more informative but the second is less expensive.

According to the recent studies (Kiipli, Kallaste 2002 in press) the most fruitful method for correlating of bentonites is analysing of magmatic sanidine composition. $NaAlSi_3O_8$ content in sanidine from the Upper Llandovery volcanic beds varies from 20 to 47 mol % (Fig. 2). The sanidine composition was analysed by X-ray diffraction in coarse fractions (0.04–0.1 mm) after washing out of clay material. The achieved precision (± 1 mol %) is sufficient forclear discrimination of many ash layers. Width of the sanidine reflection is a useful sign as well. In the case of very wide reflections occurring in some beds the sanidine composition is variable and not a good discrimination criterion. For analysing 1–2 g of bentonite is commonly sufficient. In deep sea sediments from Latvia even thin, less than 1 mm, beds can be often well correlated by this method.

In one core section the biotite composition of bentonites was studied by the EDS microanalyser. Biotite is the most common magmatic phenocryst used widely for field identification of the volcanic origin of clay-rich interbeds.

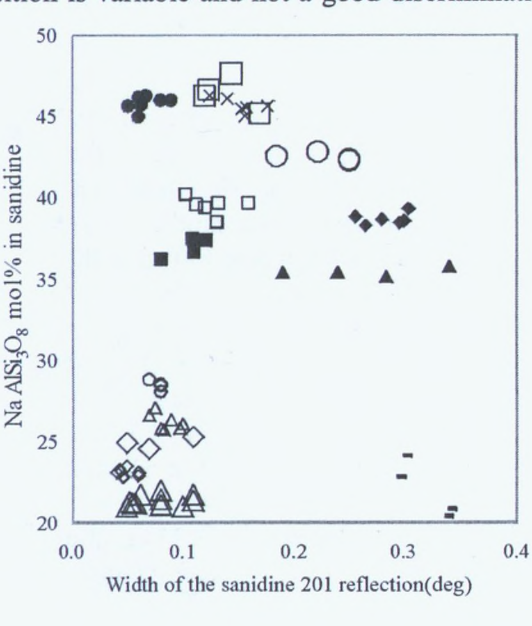

Fig. 2. Composition of magmatic sanidine from studied bentonites. The same symbol marks correlated beds (numbers are from Kiipli et al., 2001): filled circles – 19, large empty quadrangles – 3, crosses – 23, large circles – 21, small empty quadrangles – 11, filled rhombs – 27, filled triangles – 28, small emty circles – 6, emppty triangles – 10, small empty rhombs – 13, large empty triangles – 0 and lines – 7.

The Upper Llandovery bentonites contain biotite with variable composition comparing different beds. The composition of biotite is relatively stable within bed. Studied biotites could be classified as iron-rich and magnesium-rich. Iron-rich varieties are characterised often by the remarkably higher aluminium contents (Fig. 3).

In some cases lithological signs may be indicative. For example, in Saaremaa in the middle of the Upper Llandovery section beds from the andesitic source occur as hard light-coloured feldspathites, alternating with soft bentonites from more acidic source. Uniquely great thickness of particular bed is a parameter which can be measured easily and used for correlation. Well-known examples are Kinnekulle Bed in the Caradoc and Osmundsberg Bed in the Llandovery (Vingisaar 1972; Einasto et al. 1972; Bergström et al. 1995, 1998).

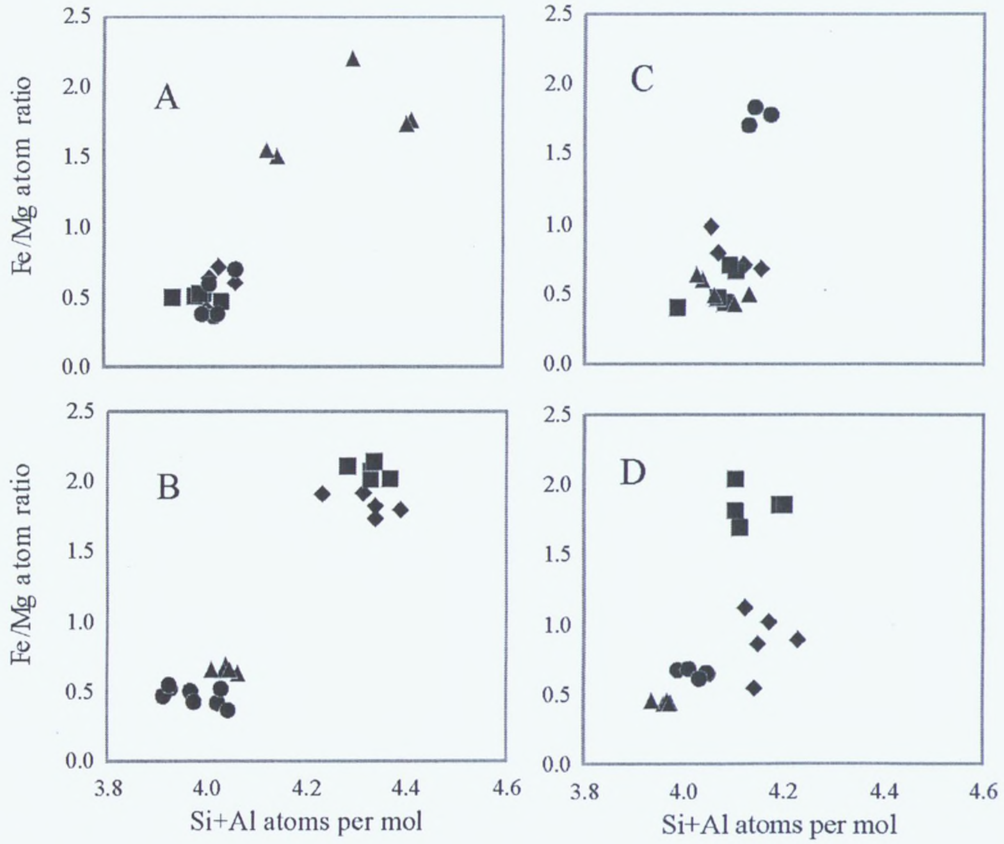

Fig. 3. Biotite composition of bentonites from Viirelaid core section. Five biotite flakes have been analysed from each ash layer. Bed numbers: A rhombs – 23, quadrangles – 19, triangles – 18, circles – 17; B accordingly 14, 13, 12, 11; C 10, 9, 7, 6; D 4, 1, new, 0.

We can conclude that several useful parameters can be analysed for identification of a particular volcanic interbed in sections. But asking philosophically: "How many coinciding signs in two samples from two different sections do we need to consider correlation proved? When can we be sure that there is no third volcanic bed with similar signs?" The only correct answer is "never" until we have not studied all volcanic beds occurring in the region in the stratigraphic unit under consideration. Although possibly all of them can be never studied, the important conclusion is that for increasing the probability of correlations one needs to use the best available biostratigraphic framework and analyse as many as possible reasonable parameters, excluding elements and minerals which distribution is controlled by sedimentary and diagenetic processes. Using compositions of all occurring ash beds we must correlate not a single bed, but a section with another section taking in account succession of number of volcanic beds. Experience in the Upper Llandovery has shown that although volcanic beds with similar compositions occur repeatedly, the unique alternation of beds in vertical section from different source volcanoes allows well establishing coeval ones.

Currently in the Upper Llandovery of Estonia and Latvia volcanic beds from at least 35 eruptions are established. This succession of volcanic beds serves as a good standard for correlating new sections on the basis of bentonites. Bulk sediment trace elements and compositions of magmatic minerals point to at least two different volcanic sources being active at the same time. Correlation between deep sea Ohesaare and relatively shallower Viki sections reveals that boundary of Velise and Jaani formations in Saaremaa (correlated with Llandovery/Wenlock boundary, Nestor 1997) in reality lies considerably lower than Llandovery/Wenlock boundary established in Ohesaare section (345.8 m) on the basis of graptolites (Fig. 4). Ireviken Event (extinction of conodonts) is still higher.

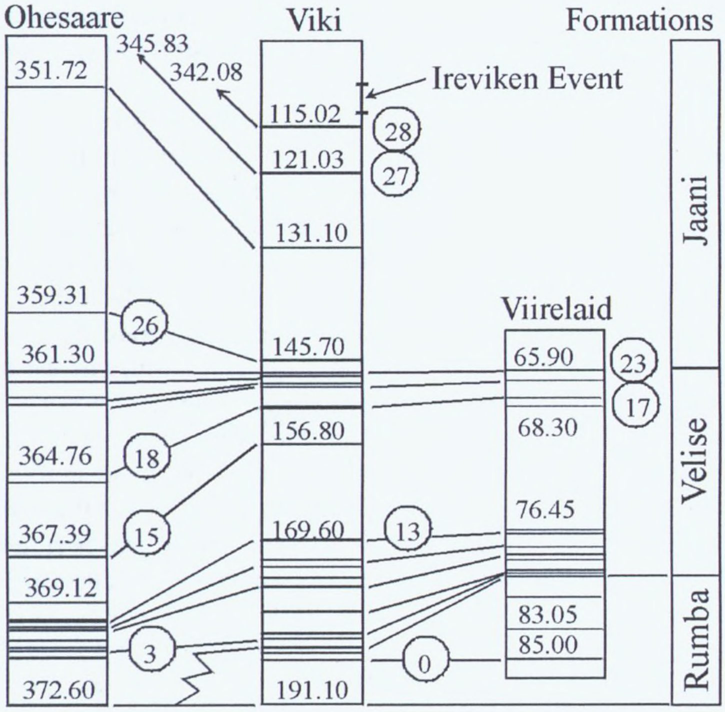

Fig. 4. Correlation of some Upper Llandovery sections from Estonia on the basis of volcanic beds. Depth records are in metres. Numbers of eruptive layers are encircled by ring.

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ILLITE-SMECTITE DIAGENESIS OF ORDOVICIAN K-BENTONITES OF THE BALTIC BASIN: IMPLICATIONS FOR BASIN DEVELOPMENT

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Clay mineral evolution in sedimentary basins is principally controlled by temperature and time. The most common process observed with the advancing diagenesis is the smectite-to-illite transformation through a series of intermediate mixed-layer illite-smectite (I–S) minerals. Progressive smectite-to-illite conversion is frequently used as an empirical geothermometer showing the diagenetic grade of the sediments (Pollastro, 1993) and thus reflecting the tectono-thermal evolution of the sedimentary basins. The Baltic Basin is an old stable intercratonic sedimentary basin of the East-European Platform in which the complete stratigraphic record extends from the latest Precambrian to the Cenozoic Neogene period. Sedimentary column is the most thickest (> 2000 m) and complete in the south-western part of the basin, whereas in the northern part of the basin only sediments of Upper Proterozoic and Lower Palaeozoic age are known. Consequently, we lack for any sedimentary evidence for about last 300–400 Ma of the basin development there.

In this contribution we report a study of mixed-layer I–S from the Ordovician K-bentonites in northern and central part of the Baltic Basin (Fig. 1). Ordovician and Silurian sequence of the basin contains numerous K-bentonite (altered volcanic ash) beds which clay fraction is in most cases rich in mixed-layer I–S. K-bentonites studied are the Kinnekulle, Grimstorp and Grefsen bentonite beds/complexes, which are the most widespread and thickest in the area (Bergström *et al.*, 1995).

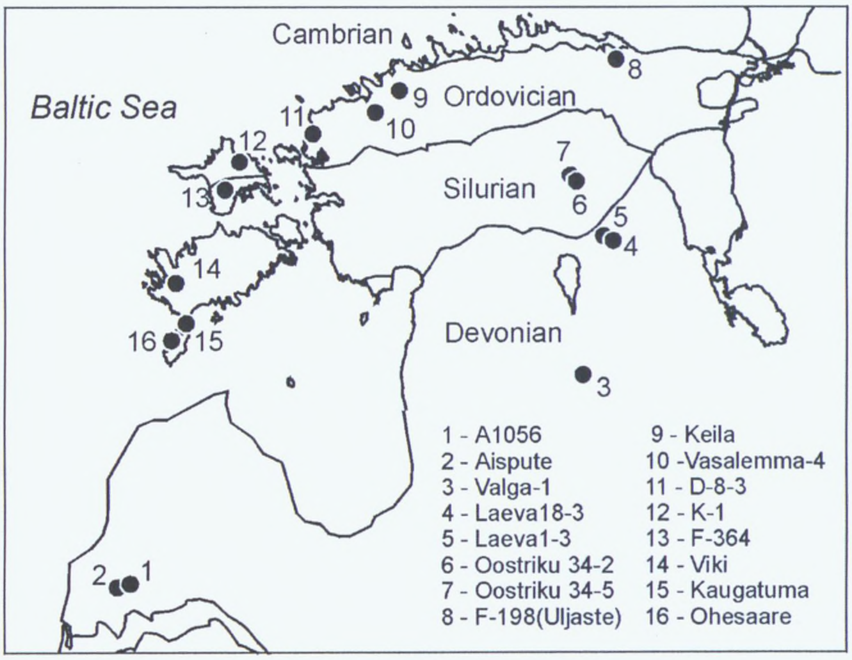

Fig. 1. Location of the drillcores studied.

Clay fraction (< 2 and < 0.2 μm) of the bentonite beds was studied by means of X-ray diffraction and K-Ar dating. The clay fraction of the bentonite is characterized by I–S with R1-1.5 type of ordering. The smectite layers content in illite-smectite (%S) is ~25–28% in near-to-surface position in the northern part of basin, whereas %S decreases gradually with the increasing burial to 16–19% at 1 km depth in the central part of the basin (Fig. 2). The K-Ar ages of the bentonite clay < 0.2 μm are confined between 357 and 384 Ma indicating the most intensive illitization during Middle to Late Devonian.

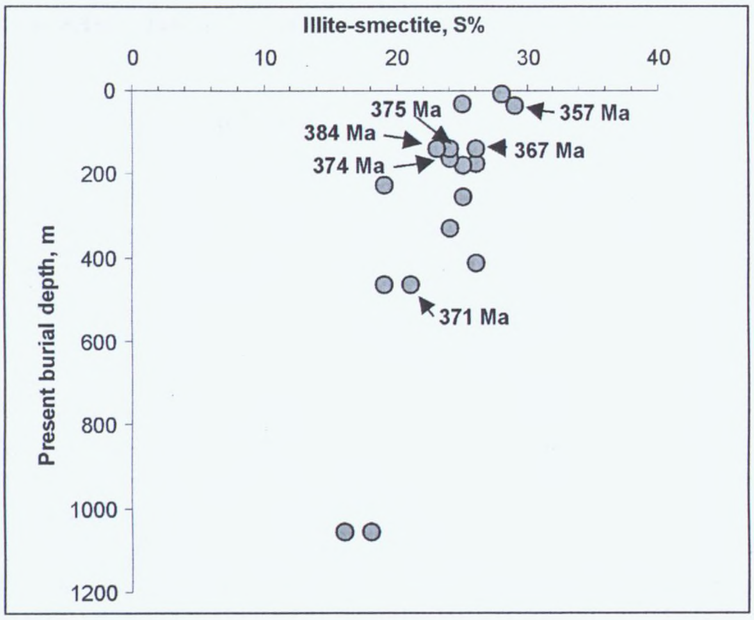

Fig. 2. Illite-smectite expandability of studied samples and K-Ar age of selected samples.

Illitization Aonset@ temperatures are estimated usually between 60–110 °C at burial depths over 2–3 km (Hoffman and Hower, 1979). Highly illitic diagenetic grade (%S) of I–S in studied bentonites suggests temperatures exceeding 100 °C during diagenesis. However, organic material maturation state indices and biomarker parameters as well as low compaction degree suggest that these and also Cambrian sediments below the Ordovician strata in northern part of the basin are thermally very immature and were never deeply buried (< 1.5 km) and/or heated above 50–80 °C (Talyzina *et al.*, 2000; Viira *et al.*, 2001). A similar highly illitic I–S (%S 10–20%) with diagenetic age of 320-380 Ma was found in Lower–Cambrian Blue Clays of Baltic Basin, which was explained by Kirsimäe *et al.,* (1999) and Kirsimäe and Jørgensen (2000) as a result of a low temperature and slow rate illitization during shallow burial. Illitization of both clays coincides with the collision Laurentia and Baltica plates during Silurian–Devonian, which resulted in folding of Caledonides and formation of deep foreland basins. I–S isotope ages suggest that this was also the period of maximum burial of Lower Paleozoic sediments in the northern part of the basin. Thickness of the Late Silurian and Devonian clastic sediments could have reached ~3–6 km in Caledonide foreland basins (Larson and Tullborg, 1998). Nevertheless, the kinetic modelling of the illitization of Cambrian clays and Ordovician bentonites suggest that maximum thickness of eroded sediments in the northern part of the basin was only about 500–800 m. Alternatively, the illitization may have resulted and/or enhanced by the expulsion of the hot K-rich fluids at the same period of basins' history.

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MIDDLE CARADOC TERRIGENOUS SEDIMENTATION IN SOUTHERN ESTONIA

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The beginning of Oandu time tentatively marks the transition from the unification stage to the differentiation stage in the sedimentation history of the eastern part of the Baltoscandian Palaeobasin. On the other hand, it also reflects transition from humid, moderate climate to arid subtropical-tropical climatic conditions (Nestor & Einasto, 1997). The general distribution of the lithofacies determines the formation of three different sedimentation areas in the Baltic part of the basin. The Caradoc facies and sedimentation problems have been studied by Ainsaar and Meidla.

The Central Baltoscandian Confacies Belt, embracing the most seaward facies in southern Estonia and western Latvia, was characterized by the clay-dominated sedimentation during late Keila, Oandu and Rakvere time (considered here as middle Caradoc; Fig. 1). The belt is represented by claystones with a very low carbonate content (Mossen Formation) which in some intervals can be enriched with silt (e.g. in the lowermost beds of the Priekule Member in the Valga section the content of the coarse silt fraction (16–63 μm) reaches almost 30% of the total insoluble residue). The Mossen Formation occurs in maximum thickness (9–10 m) in South-West Estonia and West Latvia. The sedimentation of organic-rich mud (Plunge Member) in late Keila time can be explained as an episode of flooding, and stratification of the sea after the regression during mid-Keila time (Ainsaar & Meidla 2001).

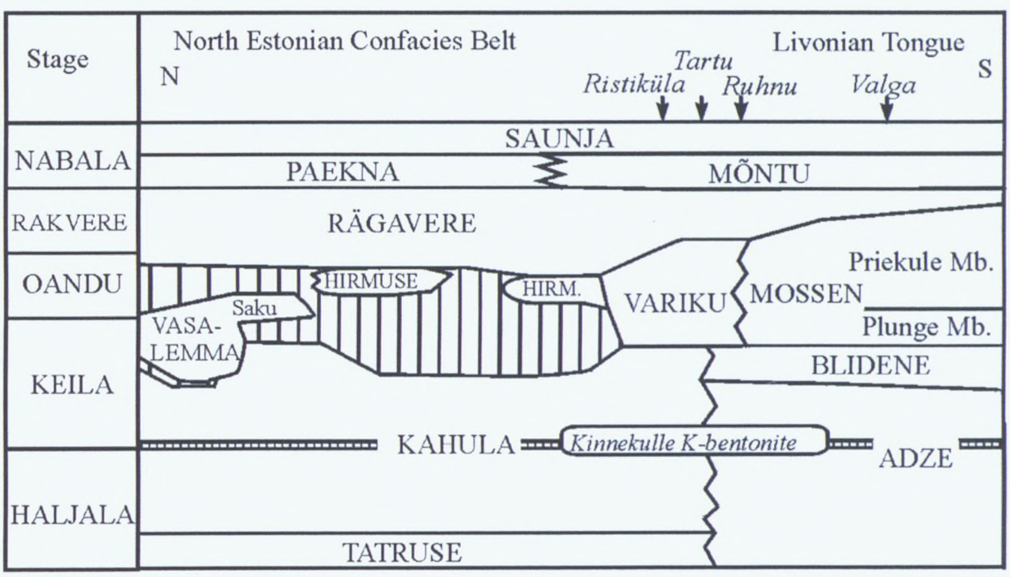

Fig. 1. Stratigraphy of the Caradoc deposits in Estonia (modified from Ainsaar and Meidla 2001). Vertical ruling shows a hiatus, arrows point at the relative position of the studied sections.

The Transitional facies belt, located between the Central Baltoscandian Confacies Belt and the North Estonian Confacies Belt, was represented by a 30–50 km wide belt of fine-siliciclastic silty sediments (Variku Formation) in southern Estonia during late Keila to early Rakvere time. The thicknesses of the Variku (maximum 15 m) and Mossen formations are rather similar (Fig. 2). Lithology of the Variku Formation shows certain variations in different parts of the basin. In the Ruhnu core the Formation is represented by 6.7 m thick claystone with only a few thin argillaceous limestone interlayers. In this respect these claystones are more similar to the rocks of the Mossen Formation of the Central Baltoscandian Confacies Belt than to the typical Variku silt- and claystone beds. We keep the Ruhnu section in the Transitional facies belt due to the lack of organic material in the claystones. In the Tartu and Ristiküla sections the Variku Formation usually consists of alternating of thin silt- and claystone beds. A thicker claybed (1–2 m) overlies the Kahula Formation in the Tartu core (the content of < 8 μm fraction reaches 60% of total insoluble residue). This bed could be a tongue of the Plunge Member extending northwards and representing the mud facies very poor in organic matter. In the Ristiküla section the silt- and claystone beds are thicker and the alternation is not so frequent. The uppermost part of the section consists of a thick marlstone bed (~ 3.5 m). These differences in the lithology of the Variku siliciclastic beds in the Ristiküla and Tartu cores could be caused by the hydrodynamic fractionation of terrigenous material during recurrent sea level changes which could be induced by storms. Rather good preservation of ichnofossils in the siltstones of the Variku Formation refers to the possibility that we can have here a process called event sedimentation. Highly changeable hydrodynamic conditions, dominating below the fair-weather wave base in the Transitional facies belt, resulted during the Keila–Rakvere interval in alternating silt- and claystone beds 10–50 cm in thickness. The terrigenous material forming the mid-Caradoc mixed rocks of the study area could be eroded from the older Vendian–Cambrian and/or Early Ordovician sedimentary rocks during several sea level changes and might have been repeatedly reworked. It has been suggested (Assallay et al. 1998) that silty material formed in large amounts during glacial erosion and was distributed by wind activity. High silt concentration could indicate the beginning of intensive orogenetic processes or cooling of neighbouring areas.

In the North Estonian Confacies Belt, representing mainly the onshore facies in North Estonian mainland, Hiiumaa and partly Saaremaa islands, the late Keila age sediments are locally missing (Fig. 2). This could be the result of both sea level fall or rapid sea level rise. The erosion or nonsedimentation rather frequently follows the sea level rise in the nearshore area (Van Wagoner et al. 1988). Grainstones of the Vasalemma Formation formed in the high energy belt during Keila–Oandu time. The mixed carbonate-argillaceous sediments (marls) of the Hirmuse Formation formed in the North Estonian area during Oandu time. The Hirmuse Formation occurs in two separate distribution areas (Ainsaar & Meidla 2001). This particular distribution pattern may suggest the existence of submarine nonsedimentation belts or be a result of postsedimentational removal of the material. The thickness of the Hirmuse Formation is almost 4 m in North Estonia, but only 0.5 m in the areas close to the hiatus (Fig. 2).

Pure calcareous mud covered transgressively the older sediments both in North and South Estonia in the Rakvere time, during a new sea level high stand in the Ordovician basin history.

Three sedimentation areas are distinguished in Estonia during the middle Caradoc. The Tartu, Ristiküla and Ruhnu sections describe the sediments of the Transitional facies belt formed during different sea level stands. Fine terrigenous material was cyclically transported to the deeper part of the basin during the regressions.

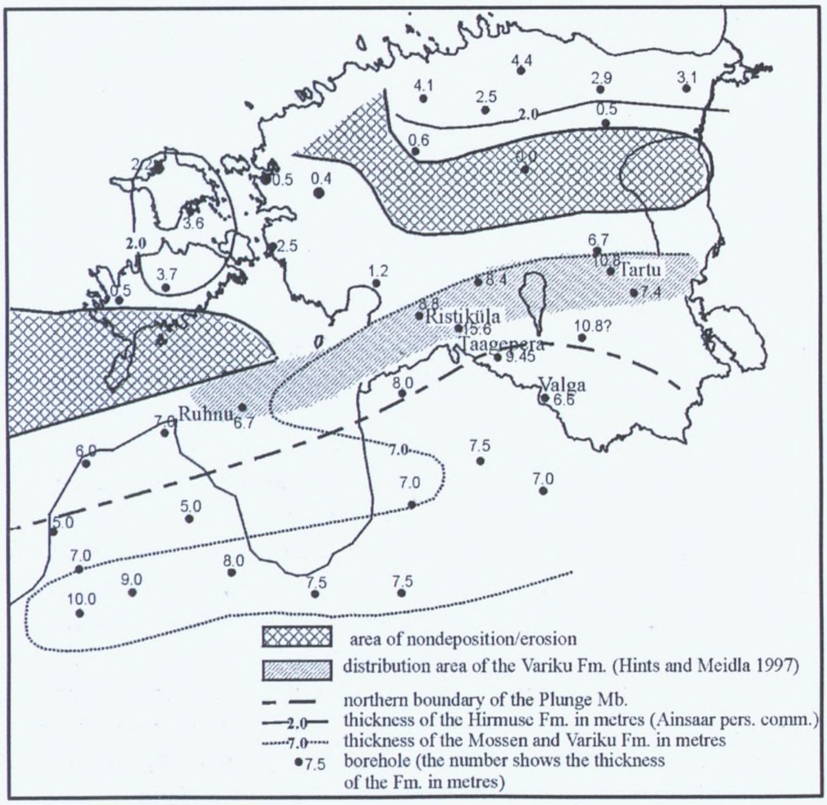

Fig. 2. Distribution and thickness of the Hirmuse, Mossen and Variku formations (Ainsaar and Meidla, 2001; partly Ulst et al., 1982).

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SEISMOSTRATIGRAPHY OF THE ORSHA DEPRESSION: NEW DATA AND GEOLOGICAL IMPLICATIONS

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The Orsha Depression is situated in the northeastern part of Belarus. It is a negative tectonic structure that appeared in the Riphean time as a result of the Earth's crust extension. The lowered position of this structure evidently shown by the crystalline basement and Middle Riphean–Lower Vendian deposits. It is bounded by the faults that show the present-day vertical amplitude of 800 to 1,700 meters. These bordering faults have an angle of dip of about 50–60 degrees. The Orsha Depression (OD) consists of the Vitebsk and Moguilev Minor Depressions and Central Orsha Horst. The Vitebsk Minor Depression (VMD) is situated in the northern part of the Orsha Depression. The crystalline basement within the VMD was subsided to 1,7 km. The Moguilev Minor Depression (MMD) is situated in the southern part of the Orsha Depression. The crystalline basement within the MMD is subsided till 1,4 km. The Central Orsha Horst is situated in the central part of the Orsha Depression. It is bounded by subregional faults, shows a SW–NE orientation and separate the Vitebsk and Moguilev Minor Depressions. The crystalline basement occurs at a depth of 1,200 m within the Central Orsha Horst.

The crystalline basement within the Orsha Depression is characterized by seismic wave velocities from 5,700 to 6,000 m/s. It is formed by granulates, amphibolite-gneissic complex, granites and granodiorites that represented on a seismic profile as a clear image of a linear border underlain by numerous randomly arranged subhorizontal small reflectors.

Riphean deposits cover the crystalline basement within the Orsha Depression that was included in the Volyn–Orsha paleodepression during Riphean time. Lower Riphean deposits overlie the basement and are represented by quartzitic sandstones. A thickness of this layer is about 250 m. Middle Riphean deposits overlie it and are formed by of siltstones and quartzitic sandstones which thickness is sometimes more than 600 m. The Upper Riphean bed is the next layer exposed in the Orsha Depression. It consists of sandstones, dolomites, clay, and siltstones. The total thickness of this deposits is no more than 80 m. Seismic wave velocities within Riphean deposits range from 3,870 to 4,080 m/s. The seismic pattern of these deposits is characterized by numerous subhorizontal small reflectors more clearly visible than in the upper part of the basement. The Upper border of these deposits is an extended line which represent the erosion surface.

The Upper Riphean bed is covered by *Vendian deposits* that consist of the lower and upper parts. Lower Vendian deposits are built up sandstones, tillites, mudstones and tuffs. A thickness of this layer is sometimes more than 500 m. Upper Vendian deposits are formed by siltstones, clays and sandstones. A thickness of this layer is about 300 m. Seismic wave velocities within Vendian deposits range from 2,500 to 2,680 m/s. The seismic picture of these deposits is characterized by numerous subhorizontal long reflectors. The upper border of these deposits is not so clearly extended line as that of the Riphean surface of erosion.

Cambrian deposits are represented in the Orsha Depression only within its lower part (the Baltic Series) that is formed by siltstones, clays and sandstones with glauconite. These deposits are limited above and below by erosion surfaces. The surfaces are shown in seismic profiles as clearly defined reflectors downward and double reflectors upward. Seismic wave velocities within Cambrian deposits are the same as those measured Vendian deposits.

Devonian deposits overlie Cambrian deposits and include three parts. There are Lower Devonian, Middle Devonian and Upper Devonian deposits. Lower Devonian deposits are built by sandstones, marls, dolomites, clays. Their thickness is about 50 m. Seismic wave velocities within these deposits are range from 2,500 to 2,680 m/s. Middle Devonian deposits are formed by dolomites with thin layers of clay, marl and gypsum, sandstone. Their thickness is about 200 m. Seismic wave velocities within these deposits vary between 2,000 to 2,260 m/s. Upper Devonian deposit are formed by dolomites with thin layers of clay and marl. Their thickness is about 200 m. Seismic wave velocities within these deposits range from 2,000 to 2,260 m/s. The seismic picture of these deposits is characterized by numerous horizontal long reflectors almost without breaks.

The uppermost part of the platform cover in the Orsha Depression above Devonian deposits is represented by Cenozoic deposits that include mostly morainic clay and sand.

The evolution of the Orsha Depression as a whole consists of five stages. These are Gothian Quasi-Platform Stage (Early Riphean), Early Baikalian Kataplatform Stage (Middle Riphean – Late Vendian), Late Baikalian Orthoplatform Stage (Late Vendian – Early Cambrian) and Hercynian Orthoplatform Stage (Early – partly Late Devonian). These stages are clearly reflected in the seismic picture of the platform cover.

The Gothian Quasi-Platform Stage (Early Riphean) was the beginning of the rift structure formation. The depression appeared in Early Riphean time on the Orsha Depression's place. It was filled in with arcosic greywacke and send-quartz phosphate-bearing formations.

During the Early Baikalian Kataplatform Stage (Middle Riphean – Late Vendian) few formations appered in the Orsha Depression. These are a red-coloured siltstone-sandstone formation, a red-coloured quartzitic sandstone formation, a carbonate-terrigenous formation, a continental glacial formation. These formations were developed under rifting conditions that took place in the Orsha Depression as a part of the linear Volyn–Central Russian Zone. The Early Baikalian Kataplatform Stage was a period of the maximum subsidence under the tension stress regime.

The Late Baikalian Orthoplatform Stage (Late Vendian – Early Cambrian) was characterized by the deposition of the tuff-sedimentary terrigenous formation that was created within the post-rift depression. After this stage during the Middle and Upper Cambrian, Early Devonian time the territory of the Orsha Depression was an uplifted area and the erosion processes took place there.

During the Hercynian Orthoplatform Stage (Early – partly Late Devonian) three formations were created in the Orsha Depression. These are a sulfate-terrigenous-carbonate formation, variegated terrigenous formation, carbonate formation. These formations were developed under marine conditions associated with a sea transgression.

After this stage the territory of the Orsha Depression was uplifted again and the erosion processes took place in this area during Carboniferous, Permian and Mesozoic time. Only in Cenozoic time sediments appeared in the depression. These were morainic deposits that created a clay-terrigenous-glacial formation.

Thus, five stages could be distinguished in the Orsha Depression evolution. These stages are evident in different formations that show special seismic pictures in the two- or tree dimensional models. Sedimentation brakes existed between these stages. These brakes are represented in the seismic profiles as clearly defined long reflectors.

STRATIGRAPHY AND PALEOGEOGRAPHY OF DEVONIAN DEPOSITS IN THE PRIPYAT AND LVOV–LUBLIN TROUGHS

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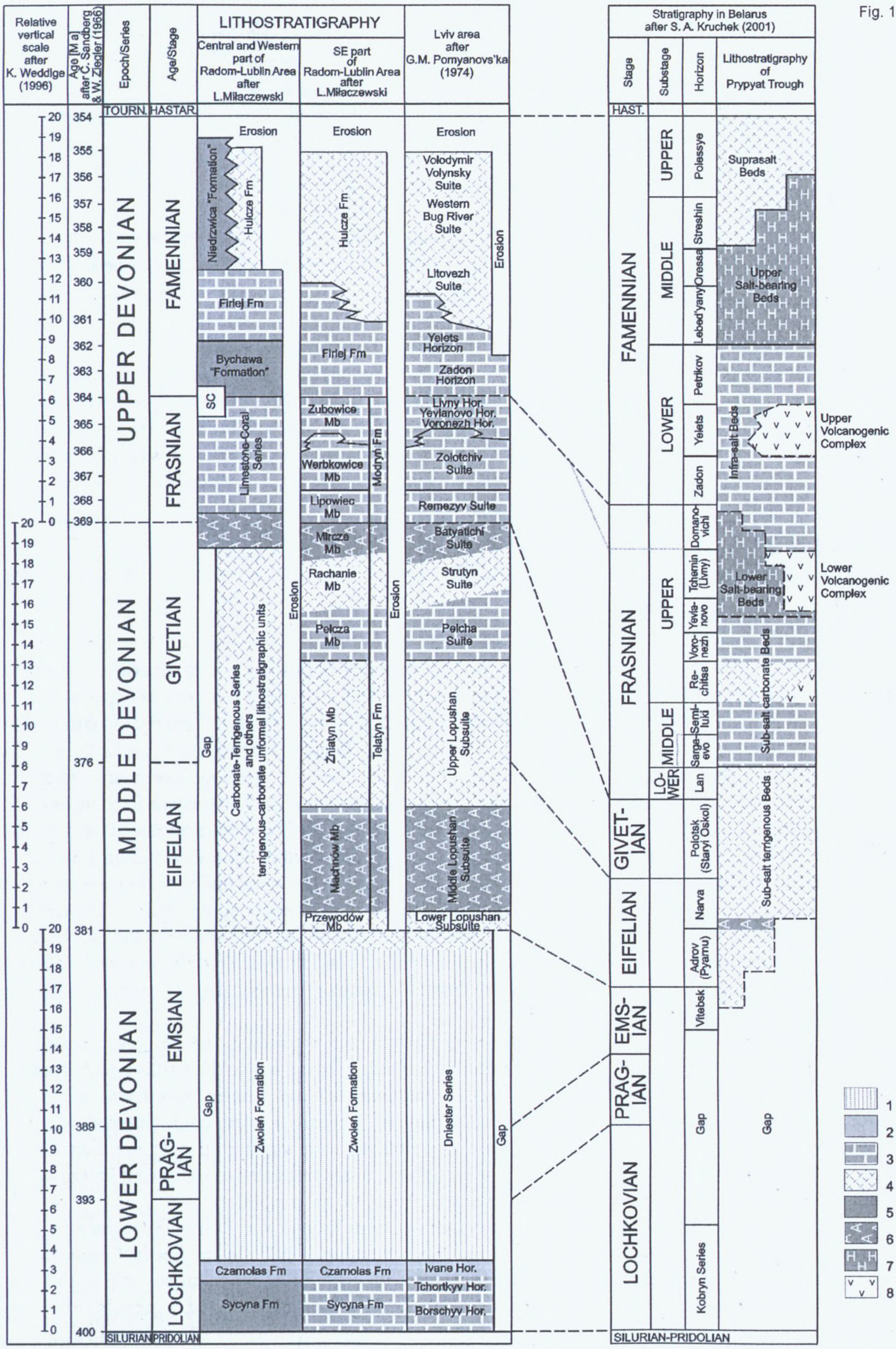
The Pripyat Trough was developed inside the old East European Platform (EEP). The Lvov–Lublin Trough was developed in a pericratonical depression situated along the SW border of the EEP. The sedimentary basins were not confined to only these two troughs, but extended also to the more elevated sites. The connections between both troughs during the Devonian were not permanent. Devonian sediments were deposited in the regressive phase of the Silurian Transgressive-Regressive Cycle and in the Middle Devonian – Early Carboniferous T–R Cycle. The T–R cyclicity of minor ranks is also evident even in lithostratigraphic units **(Fig. 1)**.

The regressive tendency in sedimentary basins during the **Early Devonian Epoch** is clearly visible. The extension of the Early Lochkovian sedimentary basin **(Fig. 2a)** was inherited from the Silurian ones. The Lvov–Lublin and Pripyat Troughs did not exist at that time. Most of deposits were accumulated in the fairly deep shelf sea. This sea regressed gradually and the sediments accumulated changed from the pelagic to neritic, littoral, up to red alluvial ones. The supposed primary extension of this basin to the NE was relatively large, so the deposition took place also on some elevated slopes of the EEP. (Please, note that the geographical directions mentioned in this text are used only for convenience, because during the Devonian these directions were different). On the contary, the extension of the sea southwards, to the territory of the supposed "Caledonian" terranes was relatively narrow (Fig. 2a). The most complete section of Lower Devonian sediments occurs in the Lvov–Lublin (or Lublin–Lvov) Trough. The thick (up to 1000 m) series of marine deposits (Sycyna Fm, Czarnolas Fm, Borschyv, Tchortkyv and Ivane Horizons – see **Fig. 1**) rest concordantly, without break on Silurian graptolithic shales. Marine deposits changed gradually upwards into red continental ones of the Old Red Sandstone type (Milaczewski, 1981, Golubtsov et al., 1981). The primary thickness of these deposits was as high as ca 1500 m (Zwolen Fm., Dniester Series – see **Fig. 1**). In the Podlasie–Brest Depression near Brest (Fig. 2a) Lower Devonian deposits occur as a thin (up to 40 m) series of marls and shales interbedded with the brachiopod-bryozoan and crinoidal-algal packstones and boundstones of the Early Lochkovian age (Kobryn Series – see Fig. 1). The Pragian and Lower Emsian deposits have not been known there (Makhnach et al. [Eds.], 2001). Most probably this is the primary lack of sediments similar to that observed in the Polish part of the Depression or in Ukraine on the elevated slopes of the Ukrainian Shield, as well as in the most part of the Belarus territory.

During the **Middle Devonian Epoch** the structural plan of the area studied changed gradually, but the Lvov–Lublin and Pripyat Troughs still did not exist. These changes begun during the Late Emsian with the lowering of the EEP elevated parts and with the marine transgression to the east and northeast. This transgression was probably induced by an eustatic event of the rise of the World Ocean level. The sea reached the Belarussian territory on the east from the side of the Moscow Syneclise and on the southeast, from the Dnieper–Donets Depression. During the same time the other marine transgression came from the west. This fact is proved by the occurrence of Upper Emsian marine deposits with abundant fauna remains in the northern part of the Holy Cross Mountains nearby Kielce. Probably, the Late Emsian as well as the Eifelian transgressions came gradually and were controlled topographically. Upper Emsian terrigenous and carbonate-terrigenous deposits in the Lublin area, as well as in the Pripyat Trough are still poorly palaeontologically investigated. The Late Emsian sedimentary basins in Poland and Belarus were probably separated during that time.

The sea in the territory of Belarus was initially united for short period with the sea in the west during the **Eifelian.** This fact is proved by the occurrence of the typically marine fauna in the Eifelian carbonate deposits in the Lublin area and in Belarus (Milaczewski, 1981, Makhnach et al. [Eds.], 2001, K. Narkiewicz – personal communication, 2002).

Fig. 1



1 – alluvial plains, 2 – tidal zones, 3 – shallow shelf sea with carbonate sedimentation, 4 – shallow shelf sea with terrigenous sedimentation, 5 – deeper shelf sea, 6 – evaporitic lagoons, 7 – salt basin, 8 – zone of subsea and island volcanism, SC – stratigraphic condensation

Fig. 2a – EARLY LOCHKOVIAN

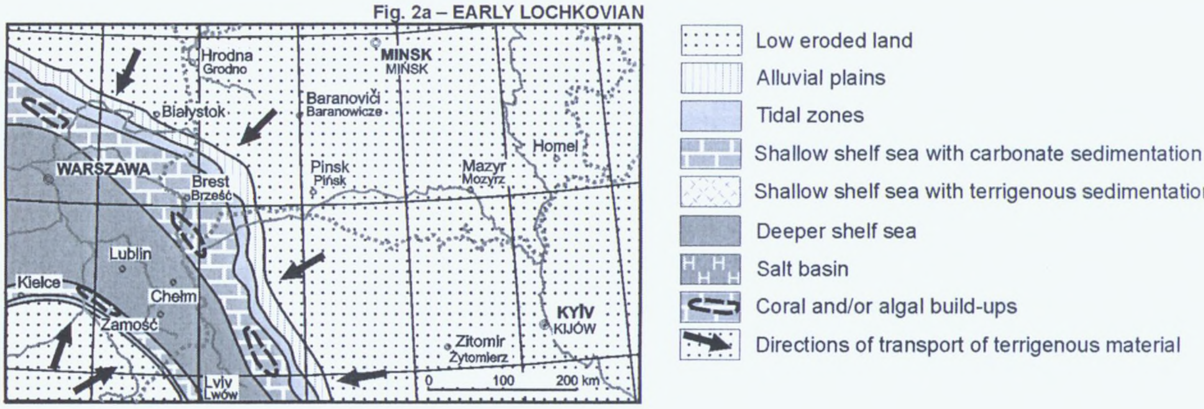
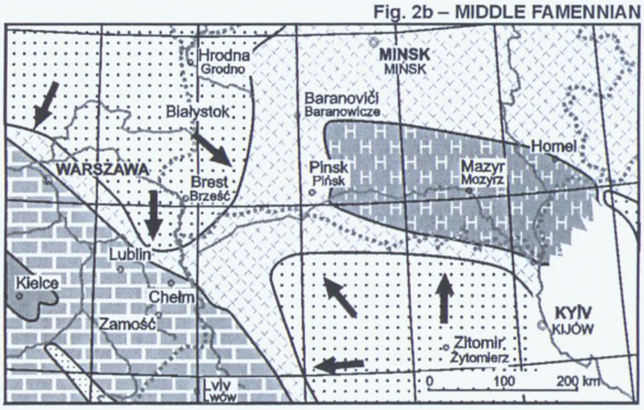


Fig. 2b – MIDDLE FAMENNIAN



During the **Givetian** the connection of the marine sedimentary basins in the area described were limited. In the Pripyat Trough the Givetian profile consists mainly of red and spotty terrigenous deposits with fish and plant remains. In the SE part of the Lublin area and in the Lvov Trough the lower (but not lowermost) part of the Givetian profile has a distincly transgressive character. Very abundant marine fossils occur there in limestones, as well as conodonts of the Givetian *varcus* Zone (K. Narkiewicz – personal communication, 2001). The upper part of the profile has a regressive character and is similar to the Belarussian sequences (Fig. 1).

The total thickness of Eifelian and Givetian deposits is as high as 200–300 m in the Lvov–Lublin and Pripyat Troughs. Small gas fields were discovered in these deposits within the Lvov–Lublin Trough.

During the **Late Devonian Epoch** marine sedimentary basins in the areas under study were widespread due to new transgressive events. Because of the peneplanation of the former alimentation territories and/or an eustatic rise of the sea level the silicic-clastic basin fill was replaced by the carbonate and evaporite ones.

During the **Frasnian** in the Lublin area as well as in the Lvov and Pripyat Troughs and on the Volyn and Podolsk slopes of the Ukrainian Shield, mainly carbonate, in places extremely shallow marine deposits with coral and algal build-ups were accumulated.

In the second half of the **Late Frasnian** the very distinct tectonic activity in the areas of the Pripyat Trough and Dnieper–Donets Depression took place. It resulted in rifting and in their filling with thick strata (up to 1000 m) of unique salt and volcanic deposits. At that time the Lvov–Lublin Trough was originated. In this trough and in its environs, as well as in the other parts of Belarus, a warm epicontinental sea existed. The seawater was of normal salinity. The organic assemblages were similar all over the area. The total thickness of Frasnian deposits in the Lvov–Lublin Trough increases from east to west from 350 m to700 m. Athickness of the Frasnian in the Pripyat Trough ranges from 300 m to 1,500 m.

During the Famennian the sedimentation in the Lvov–Lublin Trough and Pripyat Troughs was, in general, transgressive and regressive (Golubtsov et al., 1983, Pushkin, 1997). In the **Early Famennian,** the marine sedimentary basins on both areas were connected with each other due to an eustatic rise of the sea level. This worldwide event at the Frasnian–Famennian boundary caused sinking of organic build-ups and a change of organic assemblages. In the Pripyat Trough, the bulk (up to 400 m) of carbonate and volcanogenic deposits with interbeds of peculiar brachiopod and oncolithic limestoned and dolomites was accumulated. These interbeds have good reservoir properties for oil and gas accumulation (Sinichka, 1997). In the central and western part of the Lublin Trough, a series of clayey and marly deep marine sediments was deposited (Bychawa "Formation" – see Fig. 1) [Milaczewski et al., 1983]. In the eastern part of the Lublin Trough and in the Lvov Trough, the lower part of the Famennian profile consists of the characteristic marly, nodular, (knobby) limestones with very abundant brachiopod fauna remains (up to 300 m in thickness) (Milaczewski, 1981).

Beginning from the Middle Famennian and especially during the **Late Famennian** the connections between the seas in Belarus, Ukraine and Poland became gradually limited. In the Pripyat Trough during the **Middle Famennian** (see Figs. 1 and 2a) an enormous series (up to 3,500 m in thickness) of salt with 30 beds of potassium salt was deposited. In the Suprasalt Beds (Fig. 1) developed as terrigenous-clayey-carbonate deposits (up to 300 m in thickness) there are interbeds of oil and fuel shale.

During the Late Famennian to the Middle Tournaisian the accumulation of deep marine clayey-marly deposits of the Niedrzwica "Formation" (up to 400 m in thickness) took place in the western and central part of the Lublin Trough (Milaczewski et al., 1983). On the contrary, in the northern and eastern part of the Lublin Trough and in the Lvov Trough the Middle and Upper Famennian are represented by a thick (up to 700 m) series of different deposits. Sandstones occur in the lower part of the profile. Dolomites occur in the middle part, whereas the red or spotty marls and claystones lie in the upper part (Hulcze Formation in the Lublin Trough and Litovezh, Western Bug River and Volodymir Volynsky Suites in the Lvov Trough – see Fig. 1). A thickness of this series increases near the feet of the regional (Volodymir Volynsky and Kock) synsedimentary faults. Probably, in the Middle Tournaisian, the accumulation of sediments was interrupted in the areas under study.

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CHITINOZOAN REFLECTANCE AND THE THERMAL HISTORY OF THE LOWER PALAEOZOIC SEDIMENTS OF THE BALTIC BASIN

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The Baltic basin is a Phanerozoic sedimentary basin, situated on the western margin of the East European Craton (EEC). It was initiated during the Latest Vendian–Early Cambrian times and then underwent its main episode of subsidence during Silurian–Early Devonian times (Poprawa et al., 1999). The sedimentary succession in the basin is a cratonic sequence, largely of Early and Middle Palaeozoic age, with a thin cover of younger Mesozoic and Tertiary sediments. The most complete succession is found at the SW margin of the EEC – up to 2500 m, whilst adjacent to Teisseyre–Tornquist Zone (TTZ) the sediment thickness locally exceeds 5000 m. Levels in the Baltic basin (Lithuania) rich in organic matter include the Cambrian, Late Ordovician and Early Silurian black shales and the Late Silurian clay-rich succession. The Ordovician organic enriched shales and carbonates accumulated in the passive continental margin setting. In contrast the Silurian shales were deposited within a foreland basin, the system being driven by Baltica, Eastern Avalonia and Laurentia plates collision along the SW margin of the EEC. The complex interaction of both

Caledonian orogenic and dynamic mantle-flows loading drive high-rate subsidence (exceeding 100 m/Ma; Poprawa et al., 1999; Lazauskiene et al., 2002) in the south-western part of the basin such that Silurian sediments can increase by up to 10 times their thickness in comparison to the coeval cratonic succession. Hence the primary control on the distribution, thickness and thermal maturation of these rocks is burial during this Caledonian deformation (essentially Silurian) along the TTZ. Intense intracratonic subsidence during Devonian time (McCann et al., 1997) resulted in deposition of terrigenous and clay-carbonaceous sediments, organic-rich layers being much less abundant. The basin flanks were considerably uplifted during Carboniferous times. Subsequently, the succeeding Permian–Cenozoic burial has been of much less magnitude.

Although, the geodynamic evolution of the Baltic is well studied, the thermal history of the basin within both the context of the active Caledonian tectonic situation and the prolonged post-Devonian burial is not understood. Fundamentally this is because the majority of the succession is of Early and Mid Palaeozoic age and generally does not contain either vitrinite or spores. The reflectance of vitrinite and the colour of the spores being the standard optical thermal maturity indices. Therefore in order to evaluate the thermal evolution of the basin, Palaeozoic-specific optical thermal maturity indices, such as chitinozoan reflectivity (Tricker et al., 1992), were utilised together with available vitrinite and sporinite reflectivity measurements in the younger sections. Chitinozoan reflectivity have been successfully applied within a series of diverse tectonic systems (Tricker 1992; Marshall, 1995).

A series of 12 wells, representing the major structural-facial zones of the Baltic basin, have been selected along an E–W oriented profile running through Lithuania (Fig. 1).

The complete Palaeozoic succession was sampled for chitinozoan with additonal samples taken to constarin post-Palaeozoic burial. To concentrate the chitinozoans the samples were demineralized and where required the amorphous organic matter removed using an ultrasonic probe. The chitinozoan concentrates were then made into polished thin sections and reflectivity determinations made on 70 suitable samples. This gave a series of thermal maturity profiles for all the wells (Figs. 1, 2). Vitrinite and sporinite reflectivities were determined for the Devonian and younger samples. Chitinozoan reflectance values were converted to vitrinte reflectance equivalent values (Tricker, 1992) and the paleotemperatures estimated using the temperature/vitrinite relationship of Barker and Goldstein (1998).

A striking feature of all the well profiles is the presence of reversals in the thermal maturity gradients. These reversals form trends and are a real feature of the reflectivity profiles. They clearly show evidence for a thermal overprint on the normal thermal maturity gradient. To remove these effects the 'normal' sections of gradient were compared (Fig. 3) between the wells using a non-depth constrained scale. This permits identification of the original gradient and restoration of the original burial depths of each well (Fig. 1B). The reverse trend in geothermal gradients shows that the major highs in reflectivity are concentrated along the stratigraphic boundaries of the geological units (Fig. 1A). They are attributed to episodes of high-temperature fluid flushes during the development of the Baltic basin.

Reflectivity values (Rch) of just over 1% are recorded from the uppermost part of Silurian shales (Jūra Fm.) in the wells of western Lithuania (Fig. 1). These reflectivities imply temperatures in a range of 130–150 °C. Increased paleotemperatures effecting Late Silurian intervals are also indicated by high reflectance values in the other wells, where Rv equivalent highs varies in a range of 0.62–0.83. Their gives paleotemperatures as high as 94–130 °C in the central and eastern part of Lithuania. The temperature highs effecting Middle Devonian intervals were estimated as high as 70–85 °C. The reverse trend of chitinozoan reflectivity was recorded in the uppermost part of the Silurian section (Fig. 1). Anomalous high values of reflectivity were defined along the Silurian–Devonian boundary (Fig. 1) that points to the migration of the hot fluids in the overlying Lower Devonian succession. These evidences of thermal maturity highs in the Silurian Foreland basin implies early Palaeozoic maturation events during Caledonian deformation.

Vitrinite reflectance (Rv) was measured in Permian and Jurassic clays of Stoniskiai-1 well (Figs. 1, 2). The Middle–Upper Jurassic samples indicate low values (0.24–0.28 Rv), that are consistent with minimal (palaeo)burial depth of the interval. Anomalous vitrinite reflectance values as high as 0.61 were identified in Upper Permian (Suosa Fm.) strata (Fig. 2). The shift of the geothermal gradients towards the high values could be interpreted as a further hydrothermal event with the temperature of migrating fluids exceeding 90 °C. Such a paleothermal activity is also supported by late Variscan diabase intrusions identified in the Baltic Sea. (Sliaupa et al., 2001).

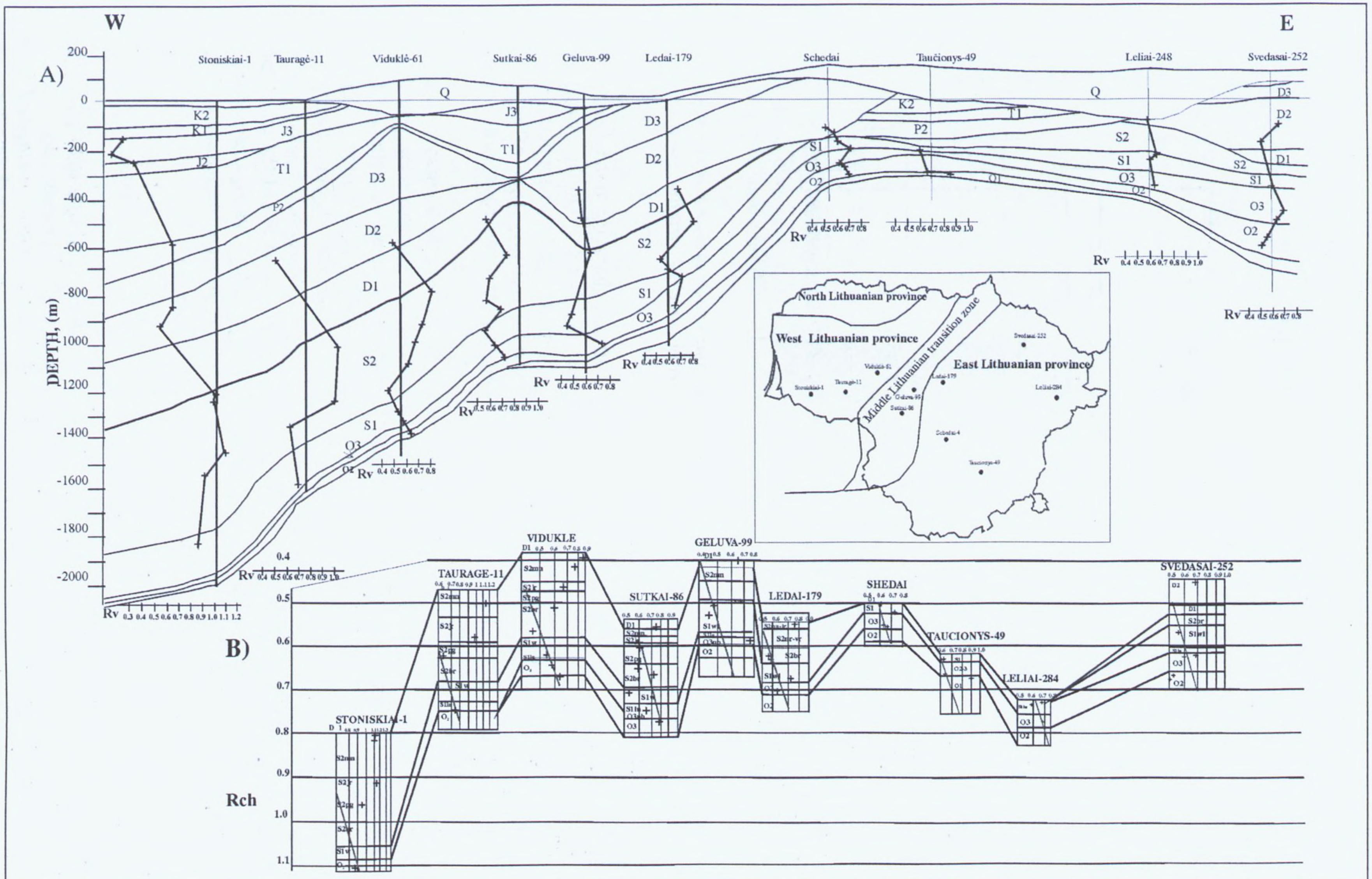

***Fig. 1.** Distribution of the thermal maturity indices in the wells of Lithuania: a) distribution of the values of vitrinite reflectivity equivalent, Rv (marked by crosses) plotted on E–W trending geological cross-section of Lithuania; b) distribution of the measured values of chitinozoan reflectivity Rch (marked by crosses); general geothermal gradient marked by solid line; locations of studied wells and geothermal provinces presented in the insertion; boundaries of the geothermal provinces adopted from (Sliaupa et al., 1999).*

The maturity profile for the most complete reference well (Stoniskiai-1) was remodelled using BasinMod 1D in order to determine the most likely thermal history (Fig. 2).

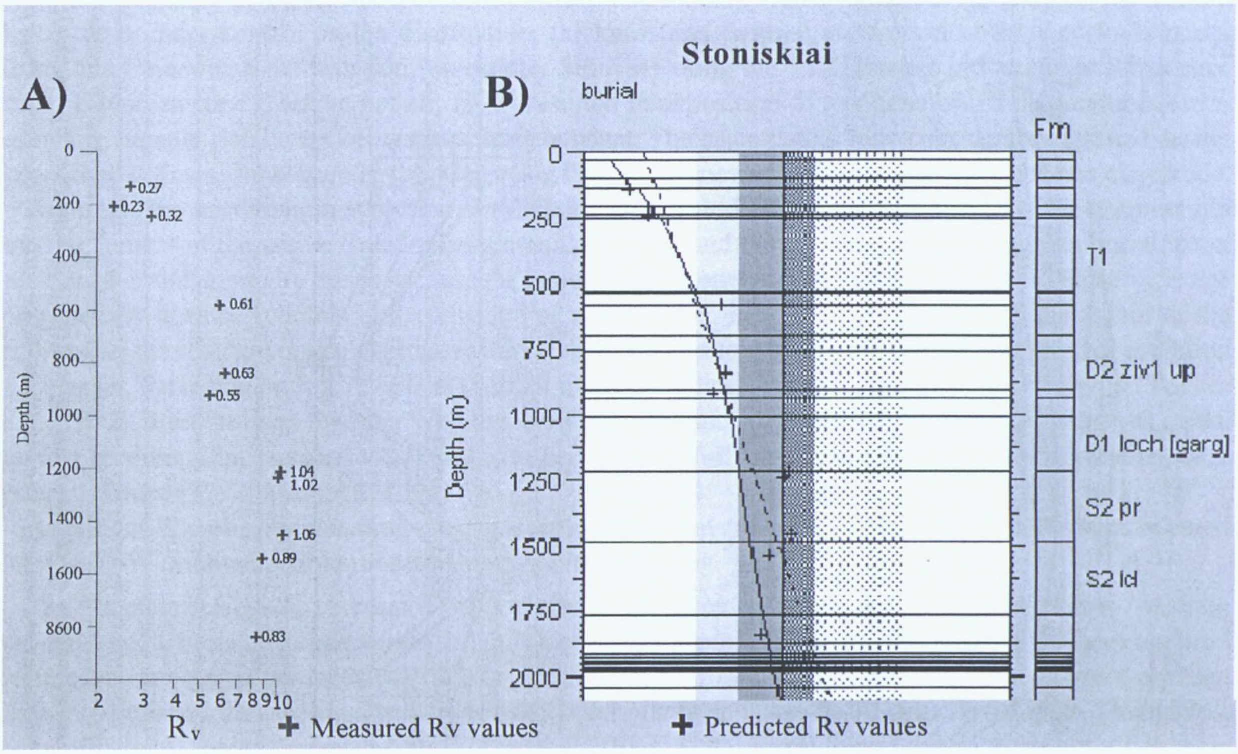

Fig. 2. Distribution of the thermal maturity indices in the reference well Stoniskiai-1: a) distribution of the measured values of chitinozoan reflectivity converted to vitrinite reflectance; b) modelled vitrinite reflectivity gradient.

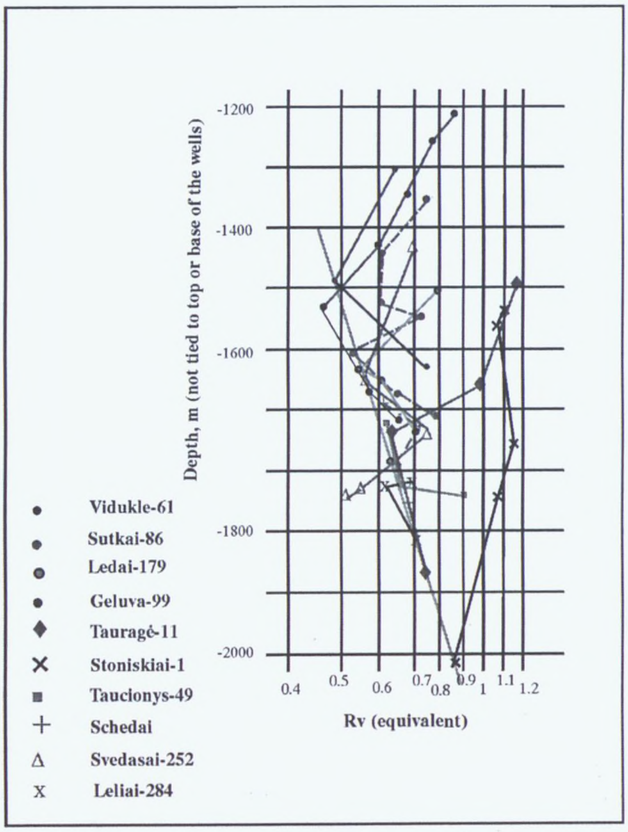

Fig. 3. The chitinozoan reflectivity gradients superimposed on a common depth scale. This shows intervals where the thermal maturity gradient has not been thermally perturbed.

The scattered character of the thermal maturity indices, closely related to the changes of the thermal regime of the basin supports extensive palaeothermal activity that affected both Palaeozoic and Mesozoic rocks in an apparently tectonic stable region.

An aproximation of the highest and the lowest chitinozoan reflectance values in the wells throught the basin shows three zones with the different paleothermal regime in the territory of Lithuania that are compatable to the major geothermal provinces of the basin. However, the data obtained do not yet allow determination of the exact timing of the thermal episodes. This requires more detailed studies including direct identification and typing of fluid events.

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SEQUENCE STRATIGRAPHY OF THE BALTIC SILURIAN SUCCESSION: TECTONIC CONTROL ON THE SEDIMENTATION IN THE FORELAND BASIN

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The Baltic basin overlaps the SW margin of the East European Craton. During the Silurian its subsidence was governed by the flexural bending of the EEC margin due to collision of Eastern Avalonia and Baltica. Two mechanisms, referred to as orogenic and dynamic loading, are accounted for the flexural subsidence of the basin (Lazauskiene et al., 2002)

The tectono-sedimentary evolution of the Silurian Baltic basin (BSB) is recorded in ~3000 m thick sedimentary succession deposited in the foreland setting. Deep-marine shales predominate in the west grading to carbonates in the periphery of the basin. Lithofacies, sequence- and cyclo- stratigraphic and geochemical analysis (Šliaupa, 1999), combined with numerical modelling (Lazauskiene et al., 2000), were applied to study the evolutionary trend of sedimentation, focusing on the imprint of geodynamic processes in adjacent orogens to sedimentary architecture.

Adopting sequence stratigraphic aproach (Karagodin, Armentrout, 1994), ten depositional sequences, corresponding to the Early Llandovery, Middle–Late Llandovery, Early Wenlock, Late Wenlock, Early Ludlow, Middle Ludlow, early Late Ludlow, late Late Ludlow, earliest Přidoli and Late Přidoli were recognized in the Silurian basin. Respectively, ten major super-cycles composed of the lower rank units were identified in the basin on a basis of lithofacies and log correlation (Fig. 1).

The architecture of sequences in the BSB was strongly dependent on tectonic processes in the evolving foreland system that was also shown by the quantitative sedimentary modelling. 2D basin infill modelling allowed the evaluation of different factors influencing the sedimentation in the BSB. Geochemical studies indicated the onset of Caledonian sourcing in the Asghill already. Two major provenances – cratonic and orogenic – competed in terrigenous sediment supply to the basin (Lapinskas, 2000; Nestor, Einasto, 1997). The cratonic provenance was accounted for the sedimentation only in the periphery of the basin, while western source increasingly supplied mainly clayey material to the basin.

Llandovery sequences correspond to the initial stage of flexuring, showing the transformation of the Baltic basin from the passive margin to convergent setting (Fig. 2). The sedimentary architecture of sequences represents an undercompensated stage of the foreland basin infilling. Small thicknesses of sediments in the

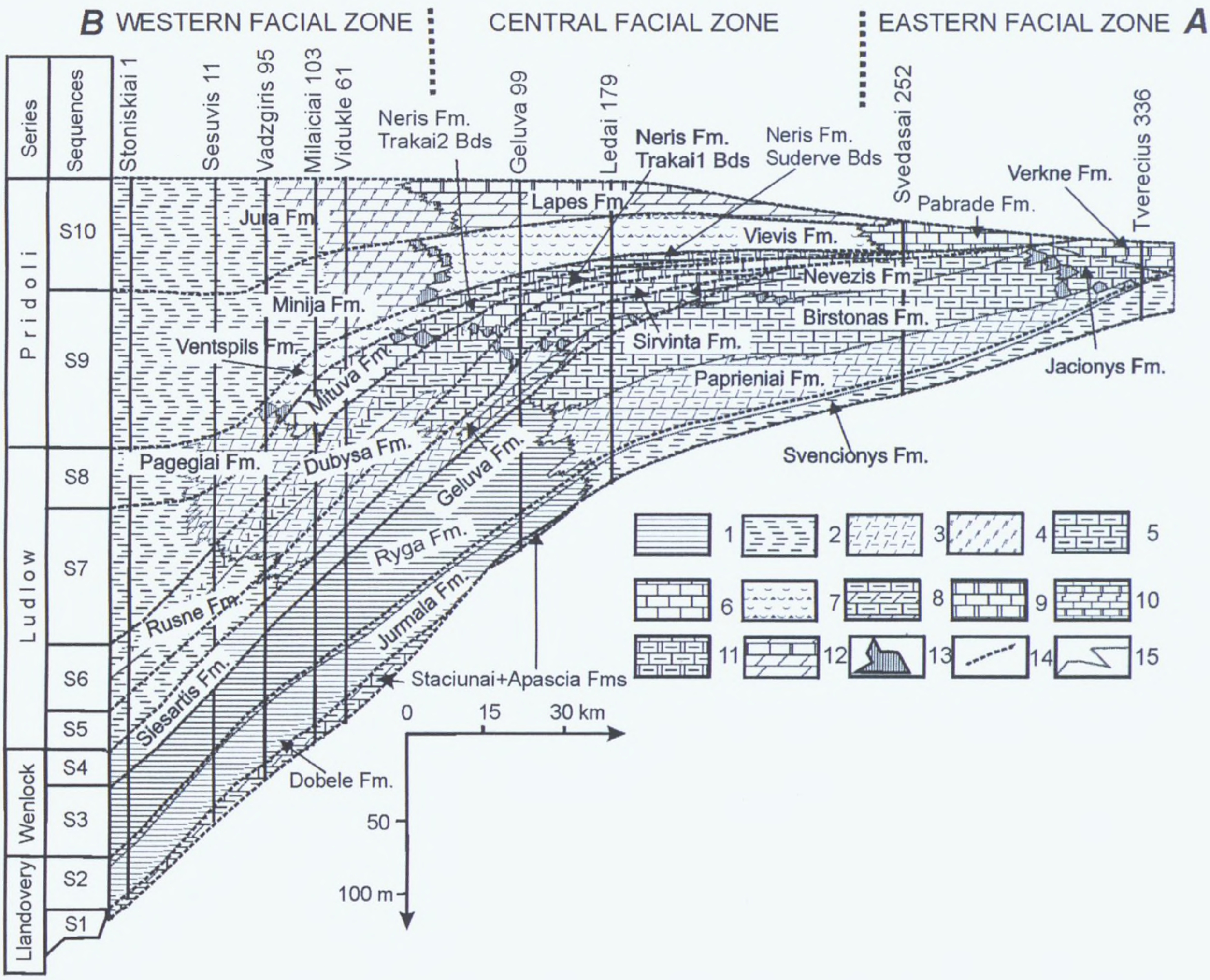

Fig. 1. Lithofacial cross-section throughout the central and eastern parts of the Silurian Baltic basin, distribution of depositional sequences and main tectono-sedimentary stages of the foreland basin developmen. 1 – shales, 2 – carbonaceous claystones, 3 – clayey marlstones, 4 – carbonaceous marlstones, 5 – clayey limestones, 6 – limestones, 7 – organogenic-detrital limestones, 8 – clayey limestones and marlstones, 9 –dolomites, 10 – dolomites and gypsum, 11 – clayey dolomites, 12 – dolomites interbedded with marlstones, 13 – reefs, 14 – boundaries of sequences, 15 – boundaries of formations.

periphery of the basin show rather low terrigenous influx that is explained by low topography of the fold belt and far distance from orogenic front. Gradually accelerating to the west flexural subsidence resulted in the deepening of the basin westward, that associated with progressing widening of the basin.

Wenlock–Early Ludlow sequences reflect intensive flexuring (Fig. 2). The sedimentary geometry of sequences reflects the compensated sedimentation, characterised by accelerating flexural subsidence, the gradual increase of sediment supply and associated narrowing of the basin. As suggested by 2D basin infill modelling, influx of terrigenous material from the cratonic source started to decline from the Late Wenlock, whereas the supply from the orogenic source considerably increased. An increase in orogenic sourced terrigenous material shows the advancement of the Caledonian orogen.

The Late Ludlow–Přidoli sequences compose the final stage of flexuring and basin infilling (Fig. 2). Considerable differentiation of subsidence over the basin indicates evolving flexural bending of Baltica plate margin. An increase of Caledonian sourced sediments, outpacing still accelerating flexural subsidence, resulted in shallowing of the depositional environments over the Baltic basin. The depositional architecture of sequences S7 and S8 shows increasing extent of the basin during the Late Ludlow, related to the progressing subduction-induced flexure (Fig. 2). Succeeding narrowing of the basin during Přidoli could not be explained in terms of eustatic sea level fall only – the mechanism of the ceasing with time dynamic loading was also involved.

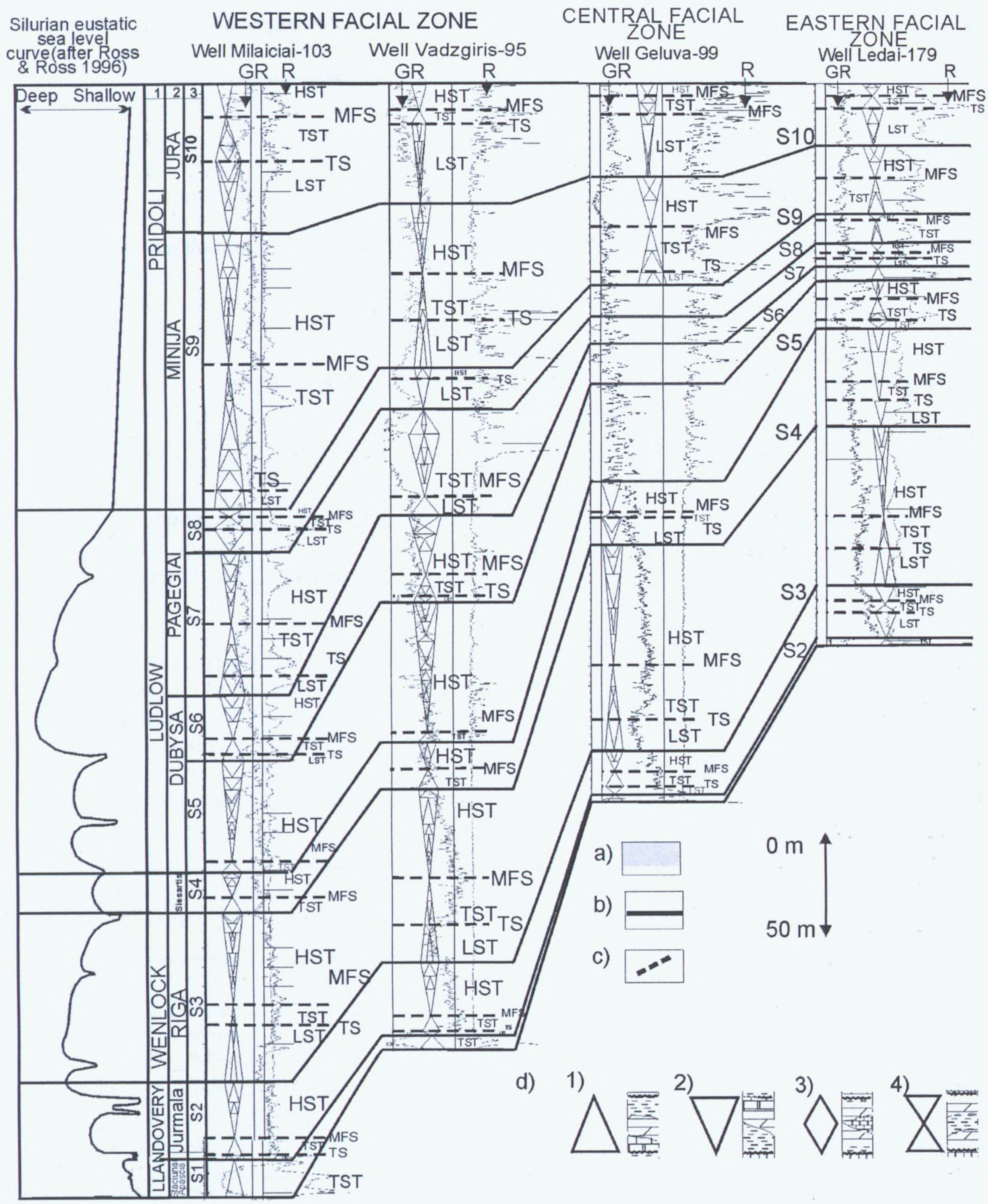

Fig. 2. Sequence and cyclo- stratigraphy of the Silurian succession: 1 – System; 2 – Formation; 3 – Depositional sequences; (a) blanked – reef growth zone, (b) correlative boundaries of sequences and major super-cycles; (c) boundaries of system tracts; (d) different types of cycles: 1) transgressive; 2) regressive; 3) regressive-transgressive; 4) transgressive-regressive (modified from Karadogin & Armanauter 1996); Abbreviations: HS – highstand system tract; LS – lowstand system tract; TS – transgressive sytem tract; SB – sequence boundary; TS – transgressive surface; MS – maximal flooding surface; GR – gamma ray log; R – resistivity log.

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PROBLEMS OF THE WEICHSELIAN STRATIGRAPHY IN ESTONIA

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Estonia is situated in an area of alternation of intensive glacial exaration and accumulation. Because of that most of the interglacial deposits, as important stratigratigraphical levels, were destroyed, displaced or redeposited into deposits of following glaciations. As a result, subdivision and limitation of deposits of different glaciations is highly aggrevated.

Early finds of the interglacial deposits in the farm wells of S Estonia have served as a basis for stratigraphic reconstructions. Having a close bedding to surface at a depth of some metres only, they could not determine the whole bulk of deposits of the last glaciation. Their erratic origin was not yet known either. The age of the interglacial deposits as Eemian has been determined at the Rõngu site only (Thomson 1941). The interglacial deposits at Karuküla were related to the upper part of the Eemian Interglacial (Orviku 1958; Fig. 1). Despite the Weichselian record was incomplete the till stratigraphy was not disturbed yet. Two till beds – the reddish-brown and violet-grey – were separated by the Late Glacial Kammeri interstadial deposits.

	Orviku 1958	Punning *et al.* 1967	Kajak *et al.* 1976	Raukas & Kajak 1995	
Weichselian	Reddish-brown till			Võrtsjärv till	Weichselian
	Kammeri interstadial deposits	Karuküla interglacial deposits	Middle Weichselian interstadial deposits	Tõravere interstadial deposits	
	Violet-grey till			Valgjärv till	

E E M I A N

Fig. 1. Development of the Weichselian stratigraphic reconstructions in South Estonia.

Based on the finite radiocarbon datings, the Karuküla interglacial deposits were plased between the above mentioned tills and considered to represent a new Middle Weihselian Interglacial (Punning *et al.* 1967, a. o.; Fig. 1). This has disturbed the Weichselian stratigraphy in a great deal. After the right Holsteinian age of the Karuküla deposits had been determined which reffered them to the Middle Pleistocene, the Middle Weichselian interstadial deposits were placed instead of them which was also not correct (Kajak *et al.* 1976). In the present local stratigraphic scheme (Raukas & Kajak 1995) the Middle Weichselian interstadial deposits, e. g. those at the Tõravere site, continue to exist between the same reddish-brown Võrtsjärv and violet (or purplish)-grey Valgjärv till units (Fig. 2). Succession of the Weichselian stadial and interstadial units has been destroyed.

The Middle Weichselian interstadial deposits do not lie above but below the Valgjärv till at the Tõravere type site in S Estonia (Liivrand 1991, 1998; Fig. 2). Consequently, the Valgjärv till does not belong to the Lower or Middle but to the Upper Weichselian. It occurs in great thickness in S Estonia reaching 20 m at the Valgjärv stratotype site. This till was formed during the main Weichselian stadials – Brandenburgian and Pommerian – when glaciers reached the southern Baltic area. The Kammeri interstadial deposits lie above this Valgjärv till.

The glacier retreat during the Late Glacial Kammeri interstadial did not extend to northern, but southern Estonia only. Because of that the corresponding tills in N Estonia form one united Upper Weichselian till bed. Its grey colour is due to glacial erosion of the grey Palaeozoic bedrocks. The younger Late Glacial stadial and interstadial deposits are not discussed in the present paper.

The interstadial deposits at the Savala site occur in the limit of the Purtse buried valley in NE Estonia where the tills are represented very poorly. Two till beds with related deposits overlie the interstadial deposits, but it is difficult to say whether they are also underlain by till or by cobbles only which occur upon the Ordovician sediments. Thus, it is not excluded that the interstadial deposits at the Savala site may belong to the Lower Weichselian.

The first Middle Weichselian till bed lies below the Middle Weichselian interstadial deposits at the Vääna-Jõesuu site (Fig. 2). This till separates the Lower- and Middle Weichselian interstadial deposits at the Prangli, Tõravere, Mägiste, Harimäe and Valgjärv sites. It forms a small thickness of about 1–7 m at a depth of about 20–107 m a.s.l. in S Estonia. At the Valgjärv (Kitse) 19 site a part of the displaced Eemian deposits, corresponding to the pollen zone M3, is mashed together with this till (Liivrand 1991). For type site of this till is proposed the Mägiste borehole 266 where it is of violet-brown colour forming thickness of 7 m (Liivrand 1991, 1998). The colour of the Mägiste till unit is not constant, e.g. being grey in the Valgjärv borehole. Borehole 266 discovers also thick (about 95 m) Middle Pleistocene tills and related deposits up to the 35 m b.s.l. (Kajak *et al.* 1963).

The Weichselian Stage starts with the Early Weichselian periglacial deposits having the greatest thickness (70 m) in the Harimäe borehole 323 which is proposed for a type site (Liivrand 1991, 1998). The bottom of these deposits lies about 2–97 m a.s.l. and top of the Weichselian reaches 48–170 m a.s.l. in S Estonia (Fig. 2).

The age of the Eemian deposits has been estimated as 115–128 ka years (BALTEEM Program) and the existence of the first Middle-Weichselian glaciation about 60 ka years ago (QEEN Program). Accordingly, there existed quite a long ice-free periglacial conditions during the Early-Weichselian and also during the Middle Weichselian. The main Weichselian glaciation started about 20 ka years ago and lasted nearly till 10 ka years.

Present official Weichselian stratigraphy in Estonia has been disturbed by the former wrong Middle Weichselian Karuküla Interglacial unit greatly. Corrections are needed to make right correlations of the tills and interstadial deposits in Estonia and neighbouring areas.

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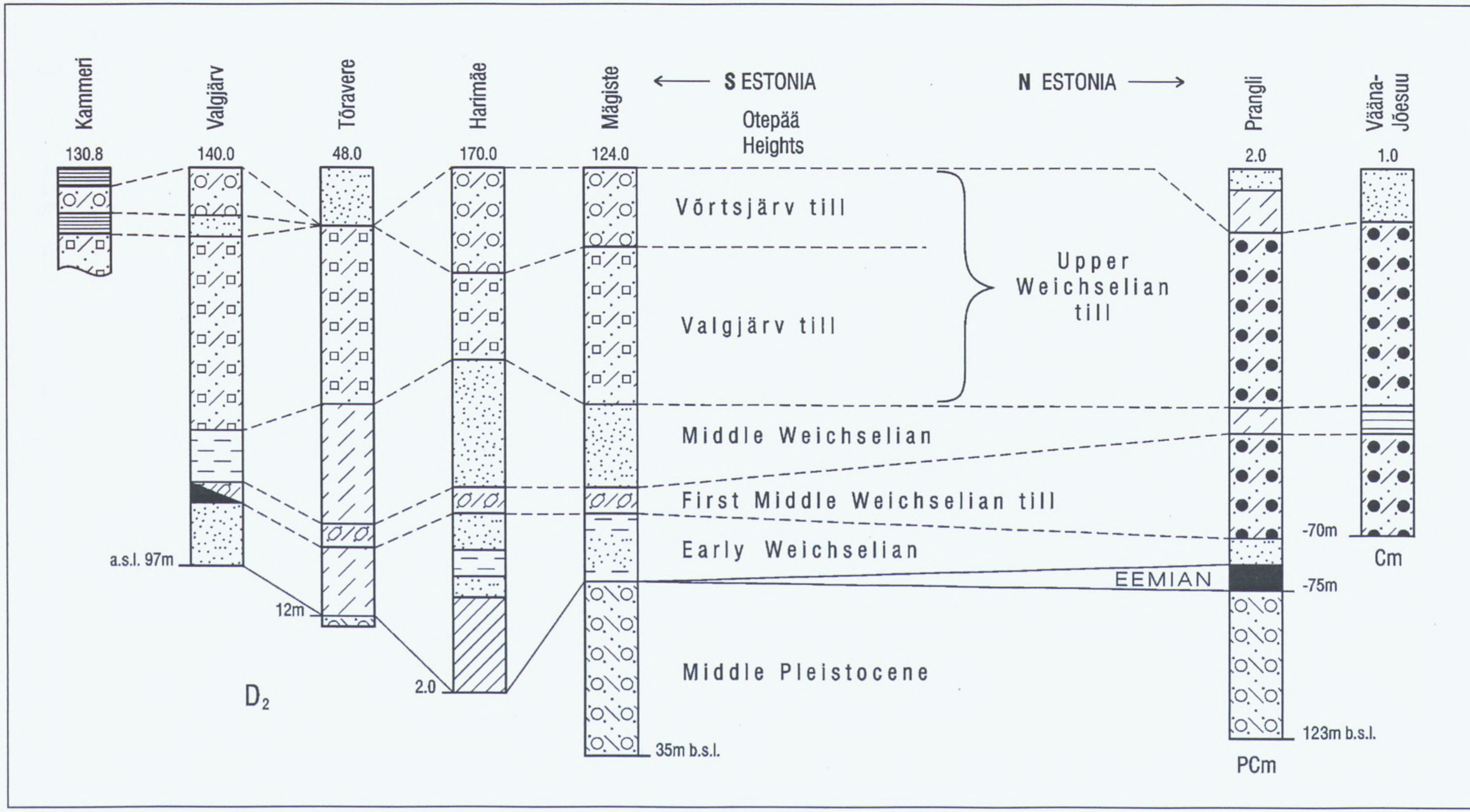

Fig. 2. Simplified correlation of Weichselian deposits in Estonia (after Liivrand 1991, 1998).

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LIMITS OF THE PLEISTOCENE GLACIATIONS AGAINST EXTENTS OF TILLS IN MID-EASTERN POLAND

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The study area comprises the Vistula drainage basin and fragments of the Neman and Pregel drainage basins in the Polish territory, and fragments of drainage basins of smaller rivers directly flowing into the Baltic Sea. The study area is approximately 190 thousands km^2. The Pleistocene complex in this region is the greatest in the European Lowland occupied by the Scandinavian ice sheet. It is generally over 200 m thick (locally more than 300 m) in the northern part of the study area. Localities of the all well-known interglacials in Poland are located in the Vistula drainage basin. There are presumably tills of the all glaciations and ice sheet advances in this drainage basin. This area is situated to the south from the basin of the Baltic Sea. Such location of the Vistula drainage basin have facilitated and directed ice sheet advances. There are not so many tills further to the west, in the Oder and Elbe drainage basins.

Geological structure and lithology of the Quaternary bedrock have influenced the final lithological composition and distribution of the Pleistocene sediments on the area. It concerns lithology of the Quaternary bedrock, first of all in the Baltic Sea Basin and in Scandinavia, from where the ice sheets have advanced.

Northwards from the Vistula, Neman and Pregel drainage basins, and southwards from the Fennoscandia crystalline basement there is a belt of outcrops, mostly of carbonate Palaeozoic rocks under the Quaternary sequence. These rocks are exposed from under the Cretaceous rocks to the east of Bornholm. Outcrops of the Palaeozoic rocks occupy a wider area to the east and in Lithuania, Latvia, Estonia and Belarus, reaching over 700 km. Crystalline rocks outcropped locally from under cover the Pleistocene sediments during the interglacials. The carbonate Palaeozoic rocks were covered mostly by sediments of the last ice sheet advance. These rocks, their weathering wastes and fluvial sediments were the main source of terrigenic material derived by all ice sheets into the Vistula drainage basin.

The analysed geological material comes from 346 localities (173 with petrographic examination, 108 with only palynologic examination and 65 with both). Petrography for 224 cartographic boreholes and 14 exposures (over 4000 samples, mostly from tills) was examined. Among them about 3200 samples formed the basis for determination of local, regional and stratigraphical lithotypes of tills in the Vistula drainage basin and the neighbouring area. Palynologic data from 173 localities (65 cartographic boreholes and exposures, from which till petrography was examined) formed a leading net of basic sites.

Geological material was analysed from 18 sub-regions of smaller drainage basins or their fragments. It introduced petrographic characteristics of the till regional lithotypes, separated by interglacial and interstadial, locally interphasial sediments.

Petrographic characteristics of the till regional lithotypes are very similar in every sub-region of the Vistula drainage basin. Palynology of the examined sediments enabled to define stratigraphical setting of the lithotypes. 16 stratigraphic lithotypes were distinguished. Their lithostratigraphic position seems univocally defined for each distinguished sub-region and for the Vistula drainage basin, also for fragments of the Neman, Pregel and other smaller drainage basins. It enabled presentation of extents of particular till lithotypes and approximate limits of the Scandinavian ice sheet advances during the following glaciations and stadials.

It is possible to distinguish tills of 16 ice sheet advances, which occurred during 8 glaciations, as well as interglacial and warm interstadial sediments in the analysed area. They are: older (lithotype A_1) and younger (lithotype A_2) stadials of the Narevian (Menapian?) Glaciation, Augustovian (Bavelian or Cromerian) Interglacial, older (N_1) and younger (N_2) stadials of the Nidanian (Elsterian?) Glaciation, Malopolanian (Cromerian) Interglacial (?), older (S_1) stadial, Domuratovian (Cromerian) Warm Interstadial(?) and younger (S_2) stadial of the Sanian 1 (Elsterian?) Glaciation, Ferdynandovian (Cromerian) Interglacial, older (G_1) stadial, Mrongovian (Cromerian?) Warm Interglacial and younger (G_2) stadial of the Sanian 2 (Elsterian), Mazovian (Holsteinian) Interglacial, Liwiecian (Fuhne?) Glaciation (C), Zbojnian (Dömnitz?) Interglacial, older (O_1) and younger (O_2) stadials of the Krznanian (Saalian = Drenthe) Glaciation, Lubavian (Rügen) Interglacjal(?), older (W_1) and younger (W_2) stadials of the Wartanian (Saalian = Warthe) Glaciation, Eemian Interglacial, the oldest (B_1), older (B_2) and younger (B_3) stadials of the Vistulian (Weichselian) Glaciation.

Ice sheets of the younger stadial of the Nidanian Glaciation and of the older stadial of the Sanian 1 Glaciation reached the South Polish Uplands. Ice sheets of the younger stadial of the Narevian Glaciation, the older stadial of the Nidanian Glaciation, the younger stadial of the Sanian 1 Glaciation and partly the older stadial of the Wartanian Glaciation were stopped in this area. Remaining limits of the ice sheets were delimited to the north from this upland.

Till of 16 stratigraphical lithotypes of the Vistula drainage basin were subdivided into three petrographical groups on the basis of analysis of gravels (diameter 5–10 mm) of the Scandinavian rocks in tills and their petrographical composition. The first group contained more crystalline than carbonate rocks or similar contents of both. They are tills of the lithotypes: A_2, S_1 and S_2 in this group, with the petrographical coefficient K/W = 1.7–1.0 (K – total of Scandinavian crystalline rocks and quartz, W – total of northern carbonate rocks). The second group was characterised by slightly more carbonate than crystalline rocks. There are tills of the lithotypes: A_1, N_2, G_1, G_2, W_1 and W_2 (K/W = 0.9–0.6). The third group was characterised by more carbonate than crystalline rocks. There are tills of the lithotypes: N_1, C, O_1, O_2, B_1, B_2 and B_3 (K/W = 0.6–0.3).

Three centres of the Pleistocene glaciations occurred. The material from the Swedish petrographical province was carried by the Scandinavian ice sheet advancing into the Vistula drainage basin from the Scandinavian Mountains. Tills of the first group of lithotypes are typical for this province. Gravel of the Bothnian petrographical province was carried by the ice sheet advancing from the Bothnian Bay and tills of the second group of lithotypes is typical for this province. The Finnish petrographical province was transported by the ice sheet advancing from Finland and tills of the third group of lithotypes are characteristic for this province.

Major differences in gravel petrography composition of the Scandinavian rocks were not observed in the separate genetic types of the till. The lodgement tills mostly contain a big number of glacial rafts of older tills. Flow tills in northeastern Poland contain more northern dolomites than northern limestones.

There is only one major difference in gravel composition of the northern rocks in tills, deposited by separate tongues (lobes) of the same ice sheet. This difference refers to the increased content of the northern dolomites, mostly in tills of the Sanian 2 (Wilgian) Glaciation in northeastern Poland.

INTEGRATION OF GRAPTOLITE, CONODONT AND CHITINOZOAN BIOZONATIONS IN THE SILURIAN OF THE AIZPUTE-41 CORE, LATVIA

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Integrated biostratigraphical studies are being undertaken in many regions in the world, but only a few of them have been published in detail (e.g. Loydell et al. 1998; Mullins & Loydell 2001). A comparative study of the distribution of graptolites, conodonts and chitinozoans in the Llandovery and lowermost Wenlock strata in the Aizpute-41 core section was completed recently (Loydell et al. in press). The co-occurrences of graptolites, conodonts and chitinozoans resulted in detail correlation of biozonations based on these three groups.

GRAPTOLITE SUCCESSION

More than 150 graptolite taxa have been identified from the Aizpute-41 core The lowermost graptolites in the section, found in thin interbeds of dark shale in the Remte Formation, indicate the uppermost Rhuddanian *cyphus* Biozone. The base of the *triangulatus* Biozone lies somewhat above the base of the Dobele Formation. The condensed dark shales of the Dobele Formation are rich in graptolites enabling recognition of most of the Aeronian biozones. The lowermost beds of the Jūrmala Formation are barren of graptolites, preventing identification of the uppermost Aeronian *halli* Biozone. Both, the upper Rhuddanian and the Aeronian graptolite assemblages are similar to those described elsewhere, including Bohemia (Štorch, 1998).

The lowermost graptolitic horizon in the Telychian part of the Jūrmala Formation yield graptolites of the *turriculatus* Biozone. Above this level all graptolite biozones, up to the lower part of the *lapworthi* Biozone, are represented. Just above the assemblage of the lower *lapworthi* Biozone, *Cyrtograptus* was identified (*centrifugus* or *murchisoni*). It means that the upper *lapworthi* and *insectus* biozones must be strongly condensed or correspond to a gap. Also, the *centrifugus* Biozone may be missing.

The Sheinwoodian *murchisoni* Biozone is more than 15 m thick in the Aizpute-41 core, and is succeeded by the *firmus* Biozone. The uppermost part of the core available corresponds to the basal part of the *riccartonensis* Biozone. The Sheinwoodian graptolite assemblages are typical for the stage.

CONODONT SUCCESSION AND ITS CORRELATION WITH THE GRAPTOLITE BIOZONATION

The Remte Formation is characterized by mainly long ranging coniform species. Just below the upper boundary of the formation, *A. expansa* Armstrong appears. This level defines the lower boundary of the *A. expansa* Biozone.The strata below this level are assigned to the *D. kentuckyensis* Biozone, and the strata above, between the level of appearance of *A. fluegeli* up to the level of appearance of *D. staurognathoides* (Walliser) are correlated with the *A. fluegeli* Biozone. The lower boundaries of the *P. eopennatus* ssp. n. 1, *P. eopennatus* ssp. n. 2 and *P. a. angulatus* biozones can be easily identified based on the appearance of the nominal taxa. Due to the lack of characteristic taxa, the lower boundaries of the *P. a. lennarti* and *P. a. lithuanicus* biozones cannot be identified. The upper boundary of the *P. a. angulatus* Biozone is tentatively drawn above the uppermost occurrence of *P. a. angulatus*. The interval above, up to the level of appearance of *P. a. amorphognathoides*, correlates with the *P. a. lennarti* and *P. a. lithuanicus* biozones. The lower boundary of the *P. a. amorphognathoides* Biozone is marked by the appearance of the nominal taxon.

Jeppsson (1997) proposed a very detail zonation for the uppermost Telychian and Sheinwoodian strata. The boundaries of his zones correspond to the datums of the Ireviken Event. Unfortunately, only some of these can be identified unambiguously in the Aizpute-41 core. The lowest possible level for the boundary between the Lower and Upper *Ps. bicornis* biozones (= Datum 2 of the Ireviken Event) lies above the uppermost occurrence of *Apsidognathus* sp. and for the boundary between the Upper *Ps. bicornis* and the Lower *P. p. procerus* biozones above the level of disappearance of *P. a. amorphognathoides* (as discussed in Loydell et al. in press). In that paper, the boundary between the Lower and Upper *P. p. procerus* biozones is tentatively drawn above the uppermost sample with abundant *P. langkawiensis* Igo et Koike, and the lower boundary of the *Ps. bicornis* Superzone was considered impossible to identify. However, there exist data which allow an alternative interpretation of the datums of the Ireviken Event in the Aizpute-41 core (discussed further in Männik et al. 2002). The disappearance of *P. p. procerus* indicates the lower boundary of the *K. ranuliformis* Superzone.

The *A. expansa* conodont Biozone correlates with the upper *cyphus* through to the upper *leptotheca* graptolite biozones, and the *A. fluegeli* conodont Biozone with the uppermost *leptotheca* and lower part of the *convolutus* graptolite biozones. The base of the *D. staurognathoides* conodont Biozone lies within the *convolutus* graptolite Biozone and its upper boundary in the *turriculatus* Biozone (*proteus* Subzone). The base of the *P. eopennatus* conodont Superzone correlates approximately with the base of the *crispus* graptolite Biozone, the base of the *P. eopennatus* ssp. 2 Biozone is close to the *crispus*/*sartorius* graptolite Biozone boundary. The base of the *P. a. angulatus* conodont Biozone (and also of the *P. celloni* Superzone) occurs within the *crenulata* graptolite Biozone, and the base of the *P. a. amorphognathoides* Biozone close to the base of the *lapworthi* graptolite Biozone. The boundary between the Lower and Upper *Ps. bicornis* conodont biozones, as interpreted in Loydell et al. (in press), lies not below the lower *murchisoni* graptolite Biozone. The last is particularly significant because the Llandovery–Wenlock boundary at the stratotype, Hughley Brook, is at Datum 2 (Jeppsson & Männik 1993). Thus, assuming no diachronism of this Ireviken datum point, the base of the Wenlock, as defined by the 'golden spike' at Hughley Brook, lies not at the base of the *centrifugus* graptolite Biozone, but at least one graptolite biozone higher.

The base of the *P. p. procerus* Superzone lies within the *murchisoni* Biozone, as it does also in the Ohesaare core (Loydell *et al.* 1998). However, the base of the succeeding *K. ranuliformis* Superzone lies one graptolite biozone higher (near the base of the *riccartonensis* graptolite Biozone) in the Aizpute-41 core than it did in the Ohesaare core (Loydell et al. 1998).

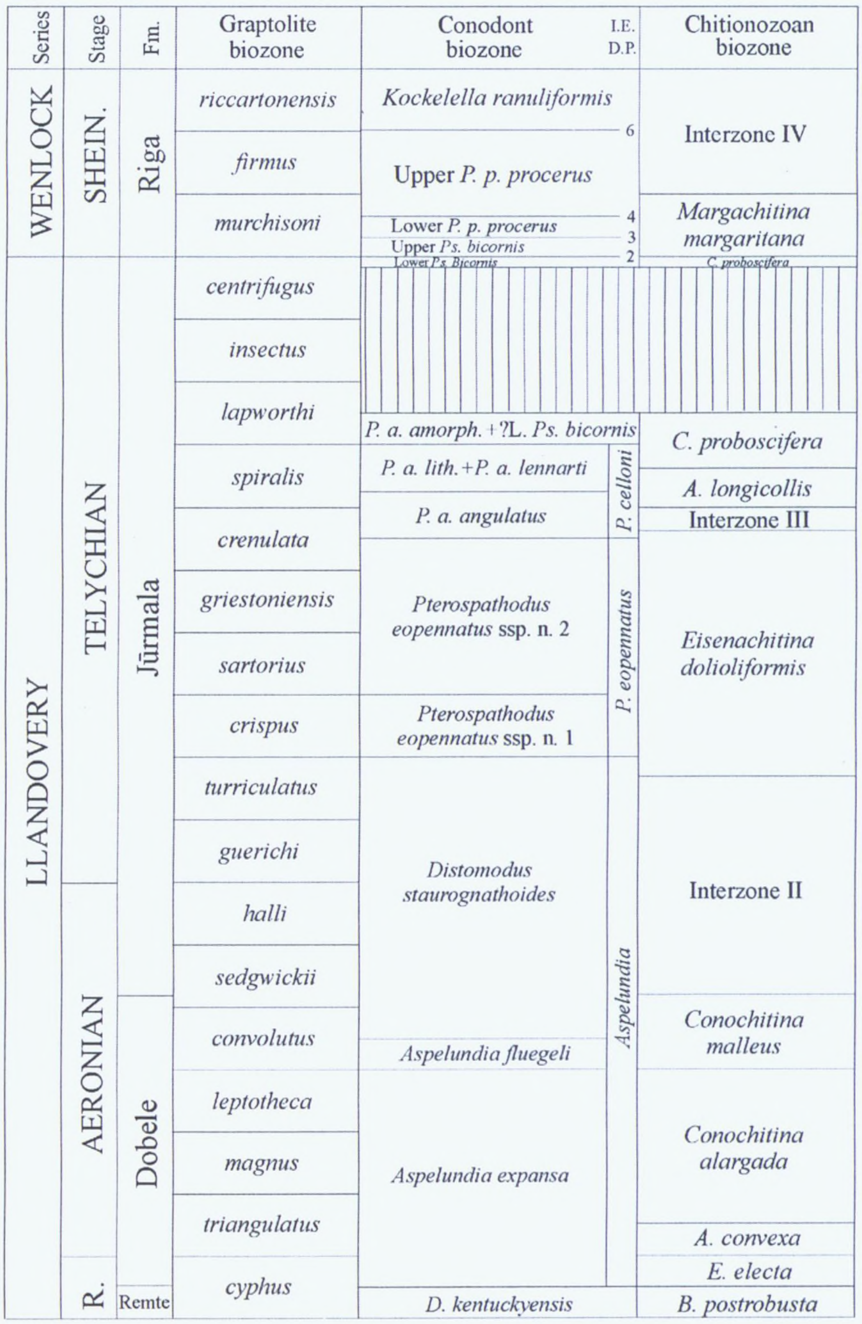

Figure. Correlation of graptolite, conodont and chitinozoan biozones as demonstrated by the Aizpute-41 core. R. = Rhuddanian; SHEIN. = Sheinwoodian; I.E.D.P. = Ireviken Event datums. Datums 2, 3, and 4 may in reality lie higher (see Männik et al. 2002).

CHITINOZOAN SUCCESSION AND ITS CORRELATION WITH THE GRAPTOLITE BIOZONES

The whole succession of the East Baltic Llandovery chitinozoan biozones (Nestor 1994) is represented in the Aizpute-41 core, with several processed samples yielding no specimens (indicated as Interzones I–III).

The lower part of the Remte Formation, up to 977.94 m, does not contain graptolites. Higher in the section, in the upper Remte and lowermost Dobele formations, graptolites of the *cyphus* Biozone were established. It means that the *B. postrobusta* chitinozoan Biozone correlates with the lower and the *E. electa* Biozone with the upper part of the *cyphus* Biozone. This agrees with the data from the Ohesaare core (Loydell et al. 1998). The *A. convexa* chtinozoan Biozone is very condenced in Aizpute, correlating approximately with the lower part of the *triangulatus* graptolite Biozone. The base of the next, *C. alargada* chitinozoan Biozone lies in the middle part of the *triangulatus* graptolite Biozone. The upper boundary of the *C. alargada* Biozone coincides with the boundary between the *leptotheca* and *convolutus* graptolite biozones. *C. alargada* Cramer was originally used as a zonal species by Verniers et al. (1995) in their global Silurian chitinozoan biozonation. In the East Baltic sections the *C. alargada* Biozone is now separated from the former *A. convexa* Biozone s. l. (Nestor 1994), corresponding to its upper part. The index species of the next chitinozoan biozone, *C. malleus* Van Grootel (nomen nudum) was described in the unpublished Ph. D. thesis (1990) and was firstly introduced as a zonal species by Dufka in Bohemia (zone C in Dufka, 1992). In the East Baltic, the *C. malleus* Biozone corresponds to the strata previously assigned to the *C.* cf. *protracta* Biozone (Nestor 1994). The lower boundary of the *C. malleus* Biozone is defined by the first appearance of the index species (the zonal assemblage is described in Nestor et al. in press). The absence of chitinozoans in the strata above (indicated as Interzone II) means that the exact position of the upper boundary of the *C. malleus* biozone in the Aizpute-41 core, and its correlation with the graptolite sequence, remains uncertain. Tentatively, the *C. malleus* Biozone is correlated with the *convolutus* graptolite Biozone. Due to the lack of chitinozoans in the Interzone II the position of the base of the *E. dolioliformis* chitinozoan Biozone in the graptolite sequence remains also uncertain. The *E. dolioliformis* Biozone, as identified in the Aizpute-41 core, corresponds to at least four graptolite biozones, from the uppermost *turriculatus* up to the *crenulata* biozones. In the Aizpute-41 core, *A. longicollis* appears first at a level close to the base of the *spiralis* graptolite Biozone, above strata barren of chitinozoans (identified as Interzone III which correlates with the uppermost part of the *crenulata* graptolite Biozone).

The base of the *C. proboscifera* chitinozoan Biozone correlates with a level in the middle-upper part of the *spiralis* graptolite Biozone in the Aizpute-41, but with somewhat higher level, probably with the lowermost *lapworthi* Biozone in the Ohesaare core (Loydell et al. 1998). The strata above the probable gap recognized in the graptolite sequence, up to the top of the *murchisoni* graptolite Biozone, correspond to the *M. margaritana* chitinozoan Biozone. The base of Interzone IV, identified by the disappearance of *A. longicollis*, correlates approximately with the base of the *firmus* graptolite Biozone in both the Aizpute-41 and in the Ohesaare core (Loydell et al. 1998).

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EVOLUTION OF UPPER SILURIAN–LOWER DEVONIAN TABULATE CORALS IN THE NORTHERN URALS

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In the Northern Ural region, there occur many sections allowing to study the changes in biota (including the tabulate corals) across the Silurian and Devonian boundary. The best of these sections are located on the western slope of the Subpolar and Northern Urals, and in the Chernov and Chernyshev Ridges.

The changes in the associations of the tabulate corals in the Silurian–Devonian boundary interval occured in the conditions of the end-Silurian regression followed by a transgression at the beginning of Devonian. Four main tabulate coral associations were recognized in Pridoli (Silurian) and two in Lockhov (Devonian). No distinct changes in the taxonomic composition of the tabulate coral faunas, as also in faunas in general, were recognized in the Silurian–Devonian boundary interval. Instead, the disappearance and appearance of tabulate coral taxa seem to be gradual. Devonian associations are closely related to their Silurian ancestors. In Early Devonian, the representatives of genera Favosites, Syringopora, Squameofavosites, Parastriatopora, but also of Tiverina and Caliopora, all known already from Silurian, are well represented.

Comparison of coeval tabulate coral associations from different geographical areas and recognition in them of common or closely related taxa allow us to perform more reliable correlations between sections on the western and eastern slopes of Urals, and also of Ural sections with those in Kuzbass and Northern Altaj.

MOLLUSK FAUNA OF THE CALLOVIAN (MIDDLE JURASSIC) FROM LATVIA

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The remains of mollusks are the most abundant macrofossils from the Jurassic deposits in the Baltic region. Representatives of all main groups of Jurassic mollusks, including bivalves, gastropods, cephalopods, and scaphopods, have been found in carbonate concretions, containing clay, sand particles, partly pyritised goethite ooliths, and abundant fossil remains, that were collected from the banks and bed of rivers Zaņa and Lose. Six concretions were disintegrated with the aim of determining and counting all macrofossils. More than 5600 macroremains of mollusks, articulated brachiopod shells, fragments of sea urchins, and polychaete tubes were detected in the material, as well as numerous pieces of carbonised wood, remains of vertebrates (fish scales?), and several microscopic tests of foraminiferans. The fossil remains show good preservation, more than a half of bivalve valves are complete, and 10–25 per cent of shells are preserved with joined valves.

The remains of bivalves, including 32 species identified in the material gathered by the author as well as 20 other species which have been reported elsewhere (e.g. Ishcreyt, 1936), are the most diverse and abundant among Mollusca. Bivalves compose about 2/3 to 3/4 (73 per cent on average) of all determined mollusk fossils. Among them *Opis* (*Trigonopis*) *curvirostra, Goniomya ornata, Bositra buchii, Astarte sauvagei, Quenstedtia gibbosa,* and *Corbulomima macneillii* have been recognised for the first time in the Callovian of Baltic, and *Mesosaccella morrisi, Gervillella aviculoides, Entolium demissum, Tancredia planata* for the first time in the Jurassic deposits of Latvia.

Gastropod remains are more rare, composing on average 18 per cent of all mollusk fossils. 20 gastropod taxa have been found in total, the most part of them, including *Conotomaria* cf. *conoidea*, *Proconulus* cf. *intactum*, *Metriomphalus segregatus, Pseudomelania laubei, Cryptaulax tortileoides, C. pseudoechinatum, Dicroloma athulia, D. pellati, Colostracon pellati, Actaeonina commoda, A.* cf. *temperata*, and *Sulcoactaeon* aff. *tenuistriatus*, being recorded in the Jurassic of Latvia for the first time. The rarest remains belong to representatives of Scaphopoda (6.8 per cent of all mollusk remains) and the cephalopods *Kosmoceras* and *Erymnoceras*, as well as undeterminated belemnites (2.2 per cent). Only five species of ammonites are represented in the author's material, whereas 18 species were previously reported from the whole section of the Callovian of Latvia.

In comparison with assemblages from Lithuania and Poland, ammonites are much less diverse, gastropods show greater diversity, but assemblages of bivalves are almost of the same composition. These differences can be probably explained by better preservation of fossils in concretions from Latvia, which in many cases contain well preserved tiny shells of small gastropods, and by the origin of the deposits closer to the shore line of the Callovian basin, probably in conditions of slightly lowered salinity.

Morphometric data show that a largest part of the mollusk shells from the Jurassic section of Latvia are smaller then those belonging to the same species represented in rocks from Lithuania, Poland and England. The composition of the concretions, and the morphometric data, probably provide evidence of insufficient oxygen content in the bottom layer of water at the time of deposition.

The results of bionomic analysis show the dominance of groups of benthic, infaunal filtrators within the assemblages of the Callovian basin. Infaunal shallow burrowers, including the bivalves *Protocardia, Anisocardia, Codakia, Mesosacella* and some others, as well as scaphopods, and epifaunal components, which were attached by byssus (*Entolium, Meleagrinella, Modiolus, Oxytoma, Pinna, Radulopecten*), dominate the benthic assemblage. Representatives of mobile benthos are also diverse and abundant, but infaunal deep burrowers (the suspension-feeders *Goniomya, Pleuromya* and *Quenstedtia*, and the deposit-feeder *Paleonucula*), unattached recumbents (*Gryphaea*), as well as pseudoplanktonic (*Bositra*) and benthonektonic forms (amonites) are the rarest components of the fauna. Assemblages dominated by infaunal shallow burrowers are characteristic of the sublittoral zone of marine basins, on soft sandy substrates containing a high proportion of clay particles (Fürsich, 1976). The assemblage under discussion probably existed in a shallow warm Callovian Sea, on a moderately soft sandy-clayey substrate within the sublittoral zone in a changeable sedimentation environment. Good preservation of fossils, lack of evidence of transportation, and lack of preferred orientation lead to the suggestion that the remains of organisms were quite rapidly buried without any long transportation. Possibly, carbonate concretions originated as a result of catastrophic storms, when clastic particles from various environments, living animals, empty shells, their fragments and other skeletal parts of animals and plant remains were mixed together. Calcite precipitation around the accumulations of organism remains was favoured by rapid burial in organic-rich sediments, at high salinity level.

The results of the study support the opinion of Liepiņš (1961), who claimed that it is possible to subdivide the Callovian sequence of Latvia in detail. By ammonite findings, the Callovian deposits overlying the Papile Formation correspond to the Middle–Upper Callovian ammonite *Erymnoceras coronatum, Kosmoceras ornatum* and *Quenstedtoceras lamberti* zones. This interval could be correlated with the upper part of the Papartine Formation and Skinija Formation of Lithuania. Unfortunately, due to glacial disturbance of the Jurassic deposits in Latvia, the boundaries between these zones cannot yet be traced clearly in the outcrops.

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CURRENT STATE AND PROBLEMS OF BEDROCK STRATIGRAPHY IN LATVIA

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Despite the fact that the volume of studies in the bedrock stratigraphy is strongly declined in Latvia during the last decade mainly due to political and economical reasons, some achievements could be evaluated positively. One of the most valuable results of stratigraphical research is the computer-based database of all boreholes reaching bedrock in the territory of Latvia and more accurate definition of stratigraphic boundaries of units, based on the re-interpretation of core logging, seismostratigraphical and other data (Pomerantseva, Mūrnieks, Savvaitova *et al.*).

The progress in studies of the Cambrian reservoir rocks of Baltic is achieved participating in a joint project of Danish and Baltic geologists (Latvian participant A. Zabele). Sedimentological, palaeontological and stratigraphical study of the Ordovician sequence is made in collaboration between State geological survey of Latvia, Tartu University and University of Latvia.

Recently a new, re-evaluated version of the bedrock stratigraphy was provided as the result of a joint project involving the most part of Latvian stratigraphers (Gailīte *et al.* 2000). Stratigraphical schemes were not changed dramatically, but the elaboration of the project highlighted clearly problems of bedrock stratigraphy, some of which, concerning mostly the Devonian, are mentioned below. The correlation and age of the Rēzekne Formation is still under discussion mainly due to uncertainties establishing and correlating this unit. Using of the International Stratigraphical Guide as a base document for stratigraphical procedures turns our attention to re-evaluation of the relationships between lithostratigraphical and chronostratigraphical units, and some changes were made in a stratigraphical scheme of the Frasnian sequence in Latvia, particularly regarding lithostratigraphical units within the Pļaviņas and Daugava formations. Further studies of the Famennian deposits by bio-, litho- and cyclostratigraphic methods strengthened the point of view supporting the fourfold division of the Famennian. Attempted application of the so called DCT (Devonian Correla Tables) technique for the Devonian sequence of Latvia had shown that the Famennian biozones embrace much more long time intervals if compared with the Frasnian ones, what could be explained by the presence of significant stratigraphical gaps within the Famennian sequence.

The age of the siliciclastic and carbonate deposits currently attributed to the Lower Carboniferous is still unclear, and an important task of future studies could be to deepen the micropalaeontological research. The Permian limestone sequence has been a little studied in the last decade, but in 2000–2002 a new impulse in study of the Late Permian palaeoecology and palaeogeography in the Baltic area is given by Polish geologists (P. Raczyński *et al.*) in collaboration with specialists from the Baltic States. The Triassic clayey deposits in Latvia have not been studied quite a long time, except the clay exploration in early 1990-ties. A new palaeontological data recently obtained from the Jurassic sequence allow to proceed in stratification and correlation of the Callovian deposits (Lukševiča, this volume).

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SEDIMENTOLOGY AND VERTEBRATE FAUNA OF THE GAUJA REGIONAL STAGE (LATE GIVETIAN) OF MAIN DEVONIAN FIELD

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The Gauja Regional Stage as a chronostratigraphical unit was formally established by Liepiņš in 1951. In different periods it has been treated as a separate stage or as a lower part of the Šventoji Regional Stage (Kuršs, Vijding, Mark-Kurik 1981), corresponding to the *Asterolepis ornata* placoderm Zone. In central, western and south-eastern Latvia and north-western Lithuania the Gauja Regional Stage is represented by one lithostratigraphical unit – the Gauja Formation. It is dominated by sandstone in its lower part and by clayey deposits (siltstone and clay) in the upper part. Its thickness increases from south (59 m) to north (120 m) – towards the main provenances. Measurements of cross-stratification (Kuršs 1975, 1992) show that throughout the basin the current systems most commonly were directed from north to south – straight from the source area.

Sandstones of the Gauja Formation are yellowish and reddish, fine- to medium-grained, usually cross-stratified and often contain pebbles of siltstone, clay and quartz, together with the fish bones. The light minerals are represented by quartz (average content 85.2 %), feldspars (8.3%) and micas (4.3%), but the heavy mineral assemblage is dominated by ilmenite, magnetite, garnets, apatite, zircon and tourmaline (Kuršs 1975). Clay deposits are mostly bright-red and made of illite with admixture of chlorite up to 5–10%. Both the sandy and clayey deposits of the Gauja Formation often contain carbonate aggregates – cement, concretions and veins.

Sandstones of the Gauja Formation show various types of sedimentary structures, but the trough cross-stratification is the most common, indicating influence of quite strong currents. Clayey deposits usually have planar bedding, and have been formed in low-agitation sedimentary environment.

The Gauja Formation is rich in fossil vertebrates known mainly from sandstones. The ichthyofauna includes psammosteid heterostracans *Ganosteus stellatus*, *Psammolepis alata*, *P. heteraster*, *P. paradoxa*, *P. undulata*, *P. venyukovi*, *Psammosteus* sp., undetermined osteostracan, placoderms *Livosteus grandis, Plourdosteus livonicus, Eastmanosteus* cf. *pustulosus*, *Asterolepis ornata*, probable placoderm *Hybosteus mirabilis,* acanthodians *Nostolepis gaujensis, Homacanthus gracilis, Devononchus concinnus, Haplacanthus ehrmanensis*, as well as another 11 taxa mentioned by Valiukevičius (1998), possible chondrichthyan *Nodocosta*, sarcopterygians *Miguashaia grossi, Glyptolepis baltica, Laccognathus panderi, Grossipterus crassus*, Dipteridae gen. indet., *Megadonichthys kurikae*, Osteolepididae gen. indet., *Panderichthys rhombolepis,* and *Livoniana multidentata*. Besides, rare stromatoporoids, conchostracans, lingulates and articulate brachiopods, as well as a large fragments of stems of *Prototaxites* are recorded in the Gauja Formation (Kuršs, Vijding, Mark-Kurik 1981).

In north-eastern Latvia the Gauja Regional Stage due to distinct lithological features, specific sedimentary structures and mineral composition have been divided in two formations (Kuršs 1975; Gailīte *et al.* 2000): Sietiņi Fm. (lower part of the Gauja RS) and Lode Fm. (upper part of the Gauja RS). These two formations approximately correspond to the lower sandy part and the upper clayey part of the Gauja Formation, respectively.

The Sietiņi Formation is composed of light-grey, almost snowy sandstone, a little more coarsely grained than the sandy beds of the Gauja Formation. The mineral assemblage of the light-grey sandstone is dominated by minerals very resistant to weathering: if compared to the sandstone of the Gauja Formation, the content of quartz is higher in the light fraction, and the amount of zircon and tourmaline is larger in the heavy fraction. Garnets and apatite are almost absent in the sandstone of the Sietiņi Formation. In exposures of the Sietiņi Formation in Latvia – Sietiņiezis, Baltā klints, Bāle Quarry – many slump structures are present likely showing unstable bottom of the sedimentary basin. Another structures like "double cross-stratification" and cross-stratification with increased thickness of the beds probably illustrates presence of local depressions or micro-deltas within the basin. The Sietiņi Formation almost lacks fossils, containing a few vertebrate remains; only some large fragments or stems of silicified ferriferous "wood" of *Prototaxites* have been found.

The Lode Formation overlies the irregular, eroded upper surface of the Sietiņi Formation. It has been suggested (Kuršs 1992) that the erosion took place due to the slump processes. The Lode Formation is made of clayey deposits visually similar to those of the Gauja Formation. Large lenses of light-grey, very fine-grained clay are specific features of the Lode Formation, which allow to recognise this stratigraphical unit during the geological mapping. The mineral composition of clays of the Lode Formation differs from those of the Gauja Formation as well. Besides the dominant illite, clays of the Lode Formation in Latvia contain also kaolinite (15–25%), and its content still increases eastwards. The lenses of fine-grained clays have formed within large (more than 100 m wide and 20 m deep) slump depressions documented in details by V. Kuršs (1992) in Lode Quarry (Latvia) and age-equivalent deposits in Pechory Quarry (Russia). Both the Sietiņi and Lode formations are almost carbonate-free.

A rich vertebrate assemblage is known in the Lode Formation, as well the plant macroremains, parasitic flatworms, arthropods (conchostracans, ostracodes, mysidaceans, eurypterids) are present. The agnathan and fish assemblage is represented by heterostracans *Psammolepis paradoxa, P. alata* and *P. undulata*, antiarch fish *Asterolepis ornata*, acanthodian *Lodeacanthus gaujicus*, sarcopterygians *Miguashaia grossi, Strunius* sp. nov. (Upeniece 1995), *Glyptolepis baltica, Laccognathus panderi, Grossipterus crassus, Latvius* sp. nov., *Eusthenopteron* sp. nov., *Panderichthys rhombolepis*, and actinopterygian *Cheirolepis* sp.

Many lithological similarities between the deposits of the Gauja Regional Stage in north-eastern Latvia and the western areas of Russia have been noted by V. Kuršs (1992), as all these siliciclastics are suggested to be accumulated in currents directed from the eastern, strongly weathered blocks of the Fennoscandia. The above-mentioned trends of changes (content of quartz increases in sandstone, and kaolinite/illite ratio in clay) continue eastwards from Baltic area. In the Lava Quarry (Leningrad district) sandstone of the Gauja RS in light fraction contains almost only quartz, but in the Pechory Quarry (Pskov district) the same age clayey deposits besides the dominant illite have 40% kaolinite. Still higher content of kaolinite has been recognised along the Oredezh and Luga rivers (Leningrad district) in the clay pebbles found in sandstone (Kuršs 1992).

A new data on sedimentology and fossils of the age-equivalents of the Gauja Regional Stage have been obtained during the field trip of abstract authors in Russia, Pskov and Leningrad districts, in summer of 2001.

Age-equivalent of the Lode Formation are exposed in the Pechory Quarry. The lower part of the geological section here is represented by the "Lode-type" grey clay, with exposed thickness up to 10 m. In one interval black and bright-blue sooty accumulations, very similar to the uranium-molybdenum concentrates documented in the Lode quarry, are present in the grey clay bed. These accumulations occur at base of one sand-rich interval, 40 cm thick, and likely have formed due to redeposition and enrichment under influence of groundwater, which more easily penetrate through the sandy bed than the grey clay sequence. The grey clay is covered by clayey siltstone, up to 5 m thick, red, with violet and yellow stains and belts. The clayey siltstone is similar to one of clayey deposit types present in the Lode Quarry. The clayey deposits in the Pechory Quarry are overlied by the light-grey, cross-stratified fine-grained sandstone of the Amata Formation. This geological section is similar to that in older part of the Pechory Quarry, where it has been documented that the grey clays infill large slump depressions (Kuršs 1992), like in the Lode Quarry.

A new material of vertebrate fossils collected in 2001 comprises both macro- and microremains of psammosteid heterostracans *Psammolepis alata*, *P. venyukovi*, partly articulated skeletons and separate plates of antiarch *Asterolepis ornata*, scales of probably another placoderm, isolated scales and spine fragment of mesacanthid acanthodian cf. *Lodeacanthus*, disarticulated bones of coelacanth *Miguashaia grossi* and porolepiform *Laccognathus panderi*, teeth and scale fragments of struniiform and osteolepiform sarcopterygians, and possible actinopterygian remains, showing close resemblance to the assemblage from the Lode Quarry. The vertebrate remains have been collected from the upper part of greenish-grey very fine "Lode-type" clay layer (Figure).

In several places along the Luga and Oredezh rivers light-grey sandstone is present, which most possibly belong to the lower part of the Gauja Regional Stage, and could be the age-equivalent of the Sietiņi Formation. Grain size of sandstone varies from fine to coarse, in places with admixture of gravel and clay pebbles. Several erosional surfaces are present in the sandstone, and local slump depressions filled with deformed clay and sand material are related to some of the surfaces. Trough cross-stratification is dominant sedimentary structure of the sandstone, and its orientation show that the main current system has been directed to south. In places cross-stratification is deformed. Unusually thick (1.2–1.5 m) cross-beds are documented in some cases and could correspond to point bar, mid-channel bar or to microdelta. Fossils except silicified "wood" of *Prototaxites* have not been found in the sandstone. In the outcrop situated in the Tolmachovo Village, a fragment of *Prototaxites*, 55 cm in diameter, has been noticed. In the same outcrop, the deposits are abound of globular calcite cement associated with the sandstone bed rich in clay pebbles. By lithological features (light-grey colour, relatively coarse-grained texture, slump depressions, cross-stratification with slump(?) deformations and many of clay pebbles) the above-mentioned deposits could be attributed to the Sietiņi Formation.

Lateral facies changes, sedimentary cycles with synchronous boundaries traced over 500 km and presence of phosphate inclusions suggest that the deposits of the Gauja RS most possibly have been accumulated in a wide epeiric basin, which covered large areas in the north-western part of the East-European Craton. Its southwards directed current system indicate that the main provenances have been situated northwards from the basin – in the Fennoscandia. Along the southern slope of the Baltic Shield, in north-eastern Latvia, south-eastern Estonia and western areas of Russia the deposits of the Gauja RS (Sietiņi and Lode formations) likely correspond to the submarine part of the deltaic zone. This conclusion is based on presence of various-scale slump depressions, specific types of cross-stratification and increased thickness of the whole Gauja RS, up to 120 m (Kuršs 1992).

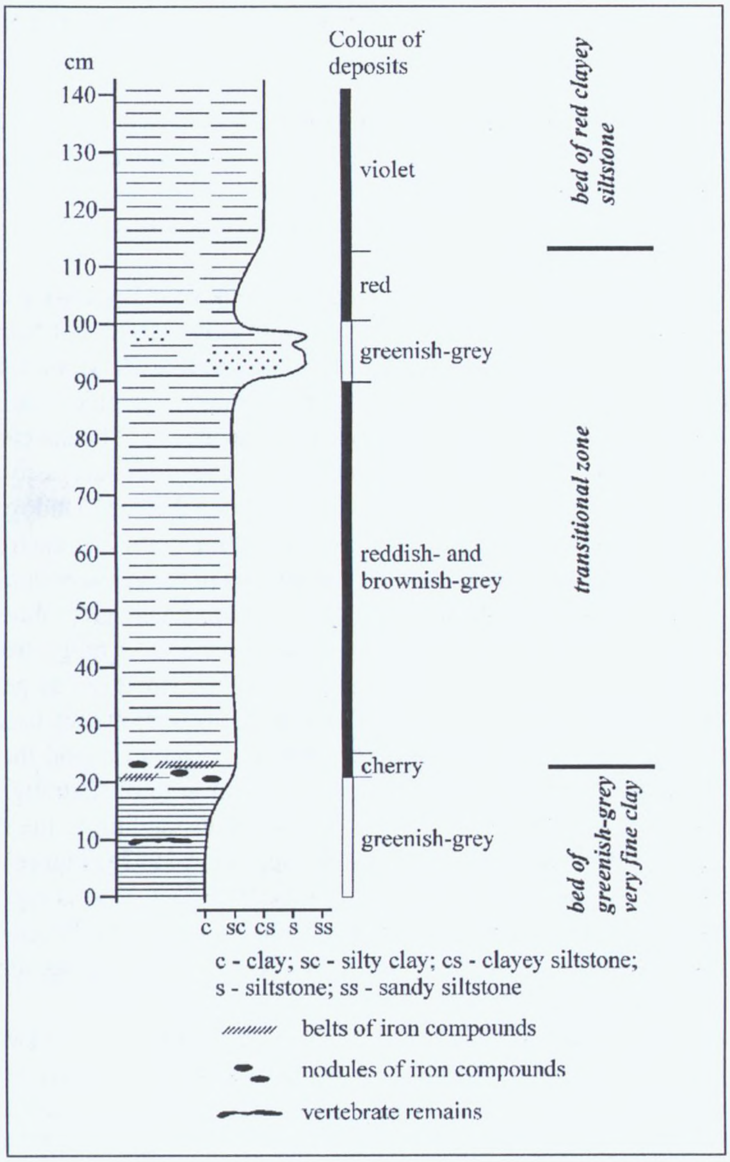

Figure. Geological section of a fragment of wall, the Pechory Quarry, the Gauja RS, Lode Fm. The upper part of the lowermost greenish-grey clay layer is particularly rich in vertebrate remains.

The lithological and palaeontological data collected during the field trip 2001, give the evidence that the deposits of the Gauja Regional Stage in the Pskov and Leningrad districts are similar to their age-equivalents in the north-eastern Latvia and south-eastern Estonia, and correspond to the same sedimentary facies – likely the submarine slope of the deltaic zone.

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DATING OF THE IREVIKEN EVENT AND THE PROBLEM OF THE LLANDOVERY–WENLOCK BOUNDARY, SOME POSSIBLE DEVELOPMENTS

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The Llandovery–Wenlock boundary was defined in Hughly Brook near Leasows Farm, Welsh Borderland. Although no graptolites were recorded from strata between 10 m below and 3.0–4.5 m above the boundary in its type section, it has been argued that the base of the *centrifugus* graptolite Biozone coincides with the base of the Wenlock Series (Holland 1980). Detailed palaeontological and biostratigraphical studies across the boundary at Leasows indicate that it does not coincide with any level in the sequence of faunal changes recognized in the section (Mabillard and Aldridge 1985). Comparative analysis of conodont successions through the Ireviken Event interval from Gotland (Sweden) and from the Leasows revealed that the Llandovery–Wenlock boundary in its type section most probably lies very close to (or coincide with) Datum 2 of the Ireviken Event (Aldridge et al. 1993; Jeppsson 1997a). As this level can be easily recognized in the calcareous facies, it gives us a good opportunity to identify the Llandovery–Wenlock boundary in sections where graptolites are rare or missing. In terms of the graptolite biozonation, the Ireviken Event was considered to range from within the *spiralis* Biozone up to the uppermost *murchisoni* Biozone (Jeppsson 1997a). However, as graptolites are extremely rare in the Ireviken Event interval on Gotland, Jeppsson's correlations were mainly based on interpretations of various data from different sections and, in reality, the position of Datum 2 (and the Llandovery–Wenlock boundary as defined at Leasows) is still reliably undated in terms of graptolite stratigraphy.

Detailed conodont biozonation with the zonal boundaries corresponding to the datums of the Ireviken Event have been worked out for the interval of this event (Jeppsson 1997a; Figure). This zonation can be easily applied (i.e. the datums of the Ireviken Event identified) in vast regions represented by calcareous rocks, but several problems, caused by ecological changes in the composition of faunas, appear when moving further offshore, into the graptolite-bearing environments. There, unambigous identification of several datums may become complicated (e.g. Loydell et al. 1998).

Recent comparative studies of faunal successions in the Aizpute-41 drill core (western Latvia) resulted in detailed correlation of graptolite, conodont and chitinozoan sequences (Loydell et al. 2002, in press). As discussed in these papers, the distribution of *Apsidognathus* sp., *Aulacognathus* sp. and *Pterospathodus amorphognathoides amorphognathoides* Walliser indicate that datums 2 and 3 of the Ireviken Event in the Aizpute-41 core lie not below the lower part of the *murchisoni* graptolite Biozone (Figure). However, rare occurrences and sporadic distribution of these taxa in the section makes it unreliable that the uppermost specimens observed in the section really represent the range ends (i.e. the extinction levels) of these taxa.

Analysis of the magmatic sanidine composition of several altered volcanic ash beds (bentonites) in many drill cores suggest that the Ireviken Event might have started even later than early *murchisoni* time (Figure). The new data suggest that the bentonite occurring at 115.0 m in the Viki core (about 1 m below Datum 1 of the Ireviken Event in this section, Jeppsson and Männik 1993) most probably correlates with the bentonite at 342.8 m in the Ohesaare core (a level in the *murchisoni* graptolite Biozone – Loydell et al. 1998) and with the bentonite at 917.4 m in the Aizpute-41 core (Kiipli and Kallaste in press; Figure). The distribution of some conodonts (mainly simple-cone taxa usually discarded as too poorly studied to be used reliably in high-resolution stratigraphy) seem to support the geochemical correlations. According to Jeppsson (1997b), *Panderodus* sp. n. N reaches Datum 3 of the Ireviken Event, and probably becomes extinct at this level, and *P. langkawiensis* Igo et Koike disappeared at (or just below) Datum 4 of the event. Both taxa reach well above the bentonites at 342.8 m (Ohesaare core) and at 917.4 m (Aizpute-41 core), suggesting that datums 3 and 4 in both sections may in reality occur high in the *murchisoni* graptolite Biozone (Figure). Moreover, if the conclusion in Jeppsson (1997b) that *Kockelella ranuliformis* (Walliser) appears (or re-appears) at Datum 1 of the Ireviken Event is correct, then also the position of that datum can be located quite precisely in all discussed here sections (Figure). The possible level of Datum 1 in all sections fits well with the correlations suggested by the bentonites. Unfortunately, at the moment, there is no data to locate the position of

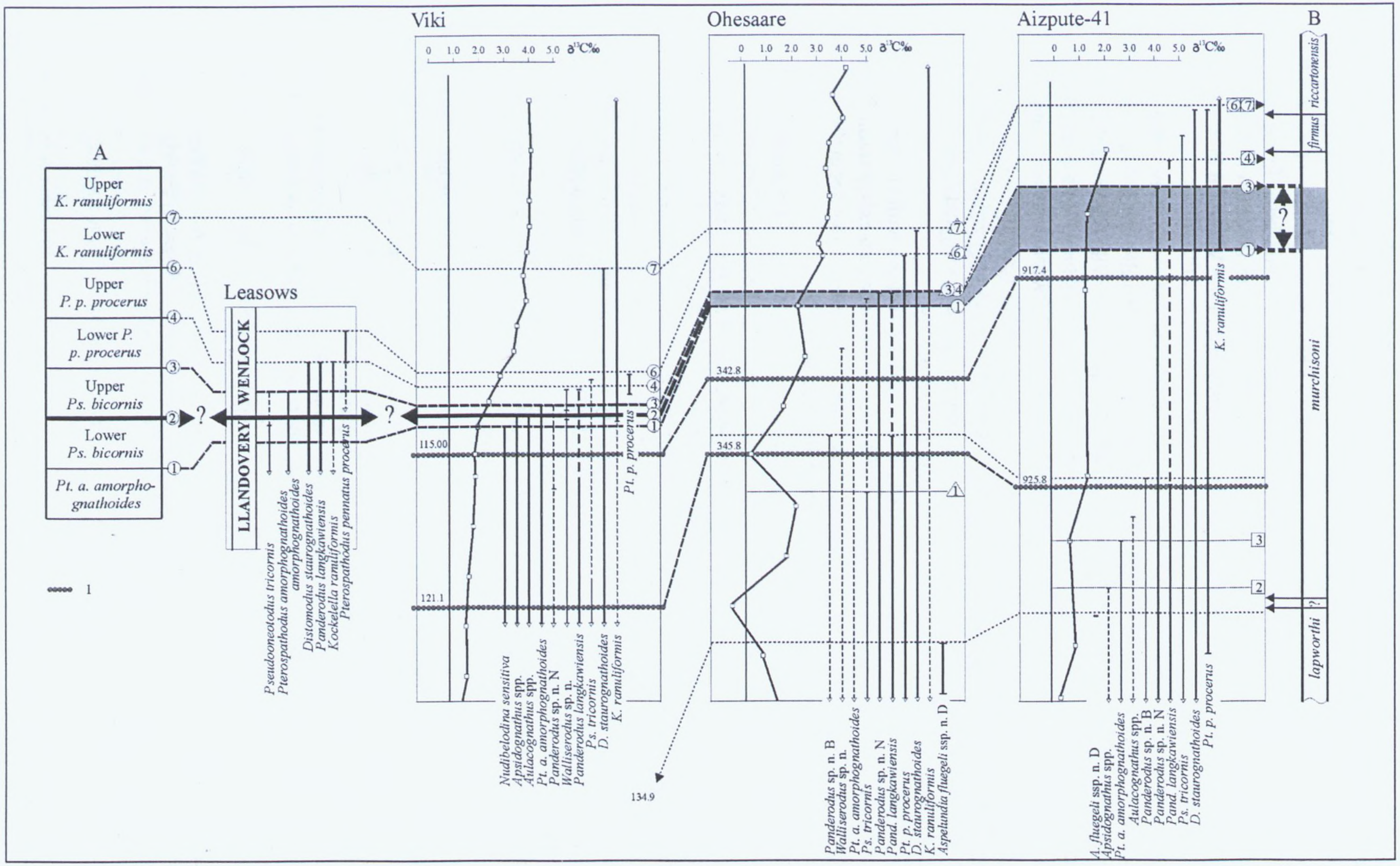

Figure. Correlation of the Llandovery–Wenlock boundary beds at Leasows and in the Viki, Ohesaare and Aizpute-41 core sections based on conodont biostratigraphy and geochemical data. Distribution of only some selected conodont taxa are shown. A – conodont zonation based on Jeppsson (1997a). B – graptolite biozonation based on Loydell et al (in press). Numbers in circles, triangles and squares – datums of the Ireviken Event (in circles – as used in this paper; in triangles – as interpreted in Loydell et al. (1998); in squares – as discussed in Loydell et al. (2002, in press)). Dashed lines are correlation lines. Light-grey belts in the Ohesaare and Aizpute-41 sections indicate the interval where the Llandovery–Wenlock boundary as defined at Leasows most probably lies. Arrows at the ends of the distribution lines of taxa indicate that these taxa occur also below (or above) the interval illustrated. Solid distribution line – taxon is common and occurs in most of the samples from the interval. Dashed distribution line – taxon is rare and occurs sporadically. 1 – bentonites.

Datum 2, and accordingly the level of the Llandovery–Wenlock boundary as defined at Leasows, in the Ohesaare and Aizpute-41 cores. Instead, the interval where the boundary should be looked for can be identified (Figure).

Also the distribution of *Aspelundia fluegeli* ssp. n. B in the studied sections indicates the possibility that the Ireviken Event most probably starts in the upper *murchisoni* graptolite Biozone and not in its lower part. This taxon has been found in many sections in Estonia. Everywhere it has a short range in the lower part of the *Pt. a. amorphognathoides* conodont Biozone, evidently marking a brief re-appearance of the *Aspelundia* lineage before its final extinction. In the Aizpute-41 core, *A. fluegeli* ssp. n. B has been found only in one sample from the uppermost *lapworthi* graptolite Biozone, in the Ohesaare core it occurs in the interval of 353.5–355.7 m (also in the *lapworthi* Biozone – Loydell et al. 1998) and in the Viki core in the interval of 134.9–136.3 m (Figure).

The studies of the changes in the $\delta^{13}C$ curve might provide some additional criteria for correlation of sections from different environments. The early Sheinwoodian $\delta^{13}C$ positive excursion is well known from many regions all over the world. Detailed studies of faunal successions and $\delta^{13}C$ changes in the Viki core revealed that here a distinct continuous increase in the $\delta^{13}C$ values started very close to (or at?) Datum 1 of the Ireviken Event (Figure). This seems to agree with the data from the Aizpute-41 core (interval between possible datums 1 and 3). However, in the Ohesaare core the $\delta^{13}C$ curve is more variable and the distinct increase in the $\delta^{13}C$ values seems to start already just above the lower bentonite at 345.8 m. Further $\delta^{13}C$ studies are needed to work out more detailed curves for the Ohesaare and Aizpute-41 cores to compare them precisely with that from the Viki core.

To sum up the data available at the moment it seems that, most probably:

- the Ireviken Event is much younger than considered till now and it correlates with an interval from the upper part of the *murchisoni* graptolite Biozone up to the lowermost *riccartonensis* graptolite Biozone;
- the base of the Wenlock, as defined by the "golden spike" at Hughley Brook, lies not at the base of the *centrifugus* graptolite Biozone, but at least one graptolite biozone higher;
- two different versions of the Llandovery–Wenlock boundary are in use: one of them corresponds to the base of the *centrifugus* graptolite Biozone (or in some cases to the base of preceding *insectus* graptolite Biozone) and another one is defined in the type section of the boundary at Hughley Brook and evidently lies in the upper *murchisoni* graptolite Biozone.

Even if the dating of the Ireviken Event above is not correct (further studies of other sections are needed to prove this), also based on the data discussed in Loydell et al. (2002, in press; the lowest possible level of Datum 2 and, accordingly, of the Llandovery–Wenlock boundary in its type section in terms of graptolite stratigraphy) it is evident that the "golden spike" at Hughley Brook does not coincide with the base of the *centrifugus* graptolite Biozone but lies considerably higher.

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KARTALASPIS AND OTHER EARLY DEVONIAN ARTHRODIRES AND THEIR STRATIGRAPHICAL SIGNIFICANCE

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Macroremains of fossil fishes, including arthrodires, even small ones, such as actinolepids and phlyctaeniids, are fairly rare in cores of boreholes drilled through the Devonian rocks. The macroremains are often fragmentary and it is hard to identify single isolated skeletal elements at the generic or specific level. In the table the best-preserved arthrodires from the Baltic area (including East Prussia) and Belarus are listed. Most of these forms are still undescribed, among others also *Kartalaspis belarussica*, the index fossil of the Rezekne Formation (Upper Emsian). An exception is *Actinolepis spinosa* Mark-Kurik, 1985. This species is an actinolepid; all the other arthrodires in the list are phlyctaeniids. The material of *Actinolepis, Kartalaspis* and a new phlyctaeniid genus from East Prussia (Kulikovskaya core) is more or less complete. Of the remaining three arthrodires, nothing but single exoskeletal plates have been found.

Table. Actinolepid and phlyctaeniid arthrodires in the Lower Devonian (D_1) and/or in the ?basal Middle Devonian (D_2) of the Baltic area (including East Prussia = Kaliningrad District, Russia) and Belarus. Abbreviations: dex – right plate, sin – left plate.

Taxon	Skeletal elements	Borehole and depth	Region	Stratigraphical unit
Actinolepis spinosa Mark-Kurik	Head + trunk-shield + scales	Ventspils-D_3 237 m	Latvia	D_1 Kemeri or Rezekne? Fm
Phlyctaeniid n.gen.?	AL sin	Rujiena 10-K 262.5 m	Latvia	D_1 Rezekne? Fm
Kartalaspis belarussica Mark-Kurik, nomen nudum	Head + trunk shield + endo-cranium	Vil'chitsy-1, 327.6 & 344.5 m; Raigla-425, 272.2 m	Belarus Estonia	D_1 Vitebsk Fm Lepel' & Obol' Beds; Rezekne Fm, lowermost part
Phlyctaeniid n.g. et sp.	Trunk shield	Kulikovskaya-1, 1158-1154 m	East Prussia	D_1
?*Phlyctaenius pusilla* (Gross)	AVL dex	Nemanskaya-9 890.5 m	East Prussia	D_1 Viešvile Gr.
Diadsomaspis cf. *elongata* (Gross)	PNu dex	Liepkalnis-137 631.2 m	Lithuania	D_2 Pärnu or D_1 Rezekne? Fm

The age of the arthrodires under consideration is in some cases controversal (or poorly known), and has changed in the course of time due to reinterpretation of the age dating of fossil-bearing strata. *Actinolepis spinosa* was described by Mark-Kurik (1985) as one of the fossil fishes coming from the Lower Devonian (Pragian–lower Emsian) Kemeri Formation (Fm) or Regional Stage (R.S.) in western Latvia. Karatajute-Talimaa (1997) mentioned that the uppermost part of the Kemeri R.S. in this region could correspond to the Rezekne R.S. in eastern Latvia. Thus this arthrodire can have a younger age.

Although recognized as one of the index fossils for the Rezekne R.S. (Rzhonsnitskaya & Kulikova, 1990), *Kartalaspis belarussica* has existed in literature as a nomen nudum for many years and the species name has been spelt differently (*belorussica, byelorussica*). This arthrodire is remarkable from two aspects. Firstly, it is the only phlyctaeniid from the Baltic area + Belarus with the endocrania preserved, even in two specimens. Secondly, it has been found in two distant localities: the Vil'chitsy borehole, Mogilev District, eastern Belarus, and the Raigla borehole, SE Estonia. In the Vil'chitsy core, *Kartalaspis* remains occurred on two levels of the Vitebsk Fm, a unit, largely coeval to the Rezekne Fm (Golubtsev, 1997; Mark-Kurik, 2000). The upper level (depth 327.6 m) belongs to the upper part the Lepel' Beds and the lower one (depth

344.5 m) to the Obol' Beds (S. Kruchek, pers. comm.). The Raigla core sample (depth 272.2 m) with a *Kartalaspis* skull comes from the lowermost part of the Rezekne Fm (Kleesment et al., 1980).

The East Prussian new phlyctaeniid from the Kulikovskaya core (1158–1154 m) has a complete trunk shield resembling that of *Heintzosteus* from the Kapp Kjeldsen division, Wood Bay Fm (Pragian) of Spitsbergen (Blieck et al., 1987). Its almost straight spinal plates are shorter than those of *Heintzosteus*. Unfortunately, more exact age of this Early Devonian arthrodire is not known.

An interesting phlyctaeniid anterior lateral (AL) plate, coming probably from the Rezekne Fm, has been found at a depth of 262.5 m in the Rujiena borehole in northern Latvia. The high plate has a long straight lower margin and a very short upper portion. The latter can be considered as an advanced character for phlyctaeniids. The plate shows some similarity to the equivalent plate of such a phlyctaeniid as *Kolpaspis* from the Battery Point Fm, Gaspé Peninsula, eastern Canada. Numerous phlyctaeniids have been found in the Battery Point Fm. This unit is now considered to be of the Emsian age (Blieck & Cloutier, 2000).

A tiny anterior ventrolateral (AVL) plate was discovered in the East Prussian Nemanskaya-9 core at a depth of 890.5 m, at the same level with the specific heterostracan *Skalviaspis* (Karatajute-Talimaa, 1989), i.e. in the Viešvile Group, Pragian-Emsian (Paškevicius, 1997). The plate belongs probably to a particularly small phlyctaeniid *Phlyctaenius* (*Phlyctaenaspis*) *pusilla* (Gross, 1937), known from the lower Emsian Klerf Beds of the Rhineland, Germany. Earlier the same fish remain was mentioned under the name *Aggeraspis*? (Karatajute-Talimaa, 1981).

One more arthrodire, *Diadsomaspis* cf. *elongata* (Gross), 1933, described by Gross (1937) from the Upper Emsian (Upper Koblenz) of the Rhineland, can be reported from the Baltic area. This comparatively large form is represented by a single paranuchal (PNu) plate, ornamented with concentrically arranged ridges. Karatajute-Talimaa (1989) mentioned this phlyctaeniid (under the name *Diadsomaspis* sp.) as coming from the basal conglomerate of the Middle Devonian Pärnu Fm from a depth of 631.2 m in the Liepkalnis core. But according to Lyarskaya (1974, Table 1), the same form (*Diadsomaspis* sp.) was found in the upper part of the Rezekne Fm.

It can be concluded that at least two phlyctaeniids, similar to the arthrodires of the same group and known from the Rhineland, occur in the Lower Devonian (or close to the upper boundary of it?) of the Baltic area. Their occurrence shows the contacts of the Baltic and Rhineland basins during the Early Devonian. *Kartalaspis* and probably the phlyctaeniid from the Rujiena core confirm the Early Devonian age of the Rezekne Fm, as phlyctaeniids are especially characteristic of the Lower Devonian of different regions, such as Spitsbergen, Rhineland and eastern Canada.

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ULTRASCULPTURE ON THE EXOSKELETON OF EARLY AGNATHANS AND FISHES

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Early vertebrate microremains, such as scales, tesserae, fin spines, fragments of platelets, etc., so-called microvertebrates, or ichthyoliths, are found in the residues of carbonate rocks treated with acetic acid. Microvertebrates are very useful for the subdivision of geological sections and correlation of beds, and they served as a basis for compiling the Silurian Biozonal Standard Scheme. The characteristic features of scales, tesserae and platelets are their morphology, sculpture and microstructure. The sculpture pattern and its measurements differ, depending on the systematic position of the taxon, the developmental level of the species, the position of the scales or platelets on the body, and also on the habitat of the particular agnathan or fish. The width, length, and height of sculptural elements differ very strongly; for example, the width of ridges and grooves on thelodont scales is 20–200 microns, and the length may coincide with the length of the scale. The sculpture (ridges, furrows, notches, spines, tubercles, etc.) on the agnathan and fish exoskeleton is known to have the hydrodynamic function, reducing the frictional drag.

Smaller sculptural elements, measured in only tens of microns, have been described and illustrated in the literature during last decades after electron raster microscope was taken into use. Herein these elements are referred to as ultrasculpture. The ultrasculpture is known in the early gnathostomes, and is explained as the imprints of epidermal cells. Schultze (1977, Pl. 13, figs. 1, 2), Richter (1995, Pl. 1, figs. 1, 3) and Smith (1977, Pl. 10, figs. 81, 82) showed nodules on the ganoid scales of actinopterygians and polygonal pattern on the cosmoid scales of sarcopterygians. Deryck and Chancogne-Weber (1995, Pl. 1) described elongate nodules on the acanthodian scales. A peculiar fine pattern of polygones and stripes was discovered on the scales that were identified as early chondrichthyans on the basis of scale morphology and microstructure (Märss and Gagnier, 2001, Fig. 3G-I). Somewhat different polygones were found on other possible chondrichthyan scales (Märss et al., 2002). These polygones consist of hexagonal fields like sarcopterygians have but each arises backwards having a scale-like appearance. In both taxa the polygonal fields are separated by grooves (intercellular channels).

Much less data can be found on the agnathan groups. Gross (1967) was the first to mention the fine sculpture ("netzartiges Muster") on the surface of thelodont scales but it was not photographed for technical reasons. Later on also the finest ornamentation on thelodont scale surfaces was illustrated (Märss, 1986; Pl. 17, fig. 7; Long, Turner, Young, 1988, Fig. 3; Märss, 1999, Pl. 2, figs 11–15). The ultrasculpture has also been shown on the ridges of some heterostracan dermal plates (Blieck, 1982, Pl. 3, figs. 3, 4; Pl. 4, fig. 1). Thus, the ultrasculpture, as known so far, is rather different consisting of elongate nodules, roundish or elongate tubercles, longitudinal stripes, irregular polygones with remains of thin walls, very regular or scale-like polygones separated from each other with grooves. The diameter of polygonal fields and width of stripes (5–20 microns) fit with measurements of smaller epithelial cells.

The morphology, sculpture and microstructure of vertebrate microremains are not always sufficient to identify a taxon, for example, to differentiate some thelodonts from some early chondrichthyans, or some chondrichthyans from some acanthodians. At the same time the microstructure and the upper layer of the exoskeletal elements (enamel, enameloid, etc.) differ in higher taxa and therefore the ultrasculpture on the surface of scales cannot be exactly the same. It is not known how much it changes among higher taxonomic groups, how constant it is on the scales or platelets from different parts of the body of a specimen, how much the ultrasculpture of thelodonts differs on the denticles from the bucco-pharyngeal region and from that on the scales, etc.? The study of ultrasculpture might give an additional characteristic for the identification of agnathan and fish microremains.

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THELODONTS AND DISTRIBUTION OF ASSOCIATED CONODONTS FROM THE LLANDOVERY–LOWERMOST LOCHKOVIAN OF THE WELSH BORDERLAND

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The distribution of thelodonts and their co-occurrence with conodonts from the Middle Llandovery–lowermost Lochkovian of Britain has been studied. Two new taxa, *Loganellia*? sp. nov. and a new paraloganiid have been established, and two subspecies of *Paralogania kummerowi* (Gross) suggested (Märss and Miller in prep.). Thelodonts are rare in the Llandovery and more so in the Wenlock Series where predominantly *Loganellia* sp. and *Thelodus* sp. are accompanied by simple coniform conodonts such as *Panderodus* Ethington 1959, *Decoriconus* Cooper, 1975 and *Dapsilodus obliquicostatus* (Branson and Mehl, 1933). The majority of the material detailed here originates from the Ludlow and Přidoli Series. Like the Wenlock, the basal part of the Ludlow Series is mostly barren of thelodonts but important species such as *Paralogania martinssoni* (Gross) and *Thelodus laevis* (Pander) have been recovered in small numbers from the Middle Elton Formation. In the uppermost part of the Upper Bringewood Formation, *Paralogania kaarmisensis* Märss (in prep.) and *Phlebolepis elegans* Pander have been recovered together with the zonal conodont *Polygnathoides siluricus* Branson and Mehl, 1933 (P. C. J. Donoghue and R. E. Elliott *pers comm.*). These thelodont taxa have also been discovered from the overlying Leintwardine Formation in the lowermost part of the Ludfordian Stage. The Whitcliffe Formation, uppermost Ludfordian, provides the most diverse and well preserved material. *Thelodus parvidens* Agassiz and *Thelodus trilobatus* Hoppe, considered here as separate species, often dominate in the late Ludfordian in association with the rarer zonal conodonts *Ozarkodina snajdri* (Walliser, 1964), *Ozarkodina crispa* (Walliser, 1964) and *Ozarkodina remscheidensis eosteinhornensis* (Walliser, 1964). At the base of the Přidoli Series bone beds produce abraded specimens but at Linley, near Much Wenlock and at Netherton and Lye, W. Midlands, well preserved specimens indicate a change to a thelodont fauna dominanted by *Paralogania ludlowiensis* (Gross) and including the new paraloganiid. Higher in the Přidoli

Series, exposures are scattered and small but a succession in faunas including *Katoporodus ?timanicus* (Karatajūtė-Talimaa), *Goniporus alatus* (Gross), *Paralogania kummerowi* (Gross) and *Loganellia*? sp. nov. can be traced before the incoming of the lowermost Devonian taxon *Turinia pagei* (Powrie) along with *Nikolivia toombsi* (Turner). Thelodont faunas offer great potential for correlation between the Welsh Borderland and the Baltic, particularly in the lower part of the Ludlow Series and uppermost Přidoli Series, where zonal conodont species are extremely rare and often absent in Britain.

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STRATIGRAPHY OF THE SEDIMENTS IN THE NEUGRUND IMPACT STRUCTURE INFILLING

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The Neugrund impact structure is one of the most important geological discoveries in Estonia. It was first discovered in 1996 by marine seismoacoustic and magnetometric profiling, which was carried out after proposing the hypothetical buried impact structure east of the Osmussaar Island (Fig. 1). During subsequent years seismic and sidescan sonar profiling, video recordings and sampling of the crater structures were performed (Suuroja & Suuroja, 2000).

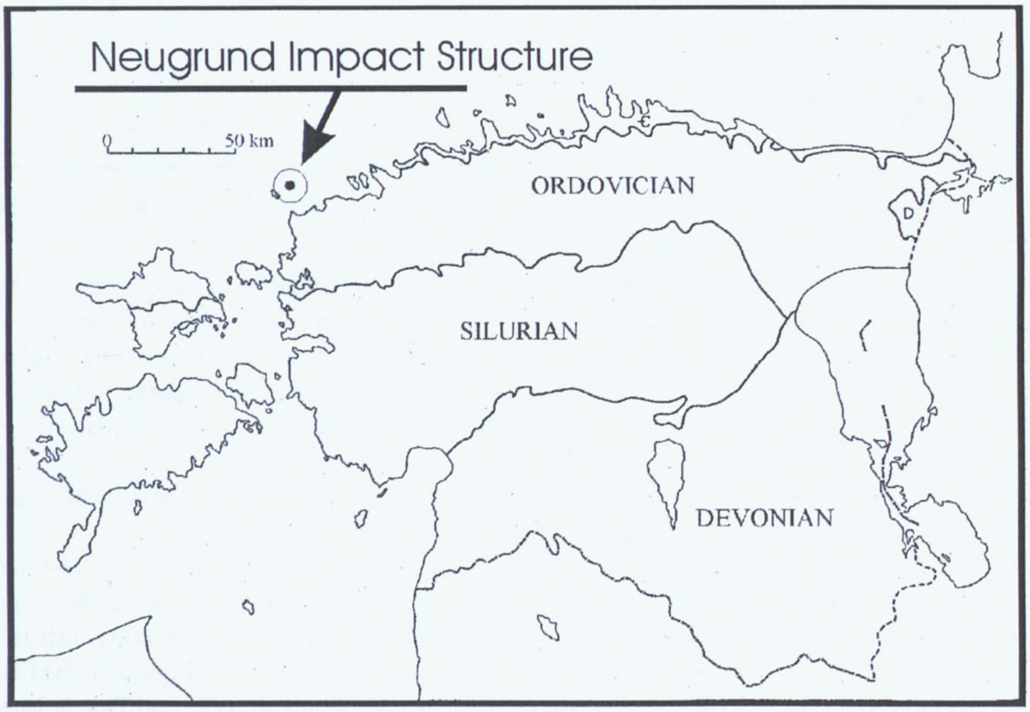

Fig. 1. Geographic position of the Neugrund Impact Structure.

The impact crater is well reflected in the seabottom topography. The crater with rim-to-rim diameter 7 km has a nearly circular limestone plateau, nearly 4 km in diameter, in the middle part. The plateau is bordered by a narrow circular canyon in its northern part. In southern part it is gradually merging with seabottom between the Osmussaar Island and the Ristinina Cape. The canyon is formed between the crater rim wall, consisting of brecciated metamorphic rocks of the crystalline basement, and the infilling with a continuous sedimentary succession in the internal slope.

The sampling of the sedimentary infilling was performed in course of diving (by S. Suuroja). Two more complete series of samples have been at our disposal. One of them was collected from the central-eastern part of the limestone plateau, the other series comes from the western margin of the infilling. These samples are well positioned, as the slope was very steep in this locality.

The set of samples from the western slope of the crater infilling was investigated for the rock composition and the grain size of the insoluble residue. The macrolithological log and the data on the content of insoluble residue are shown in Figure 2.

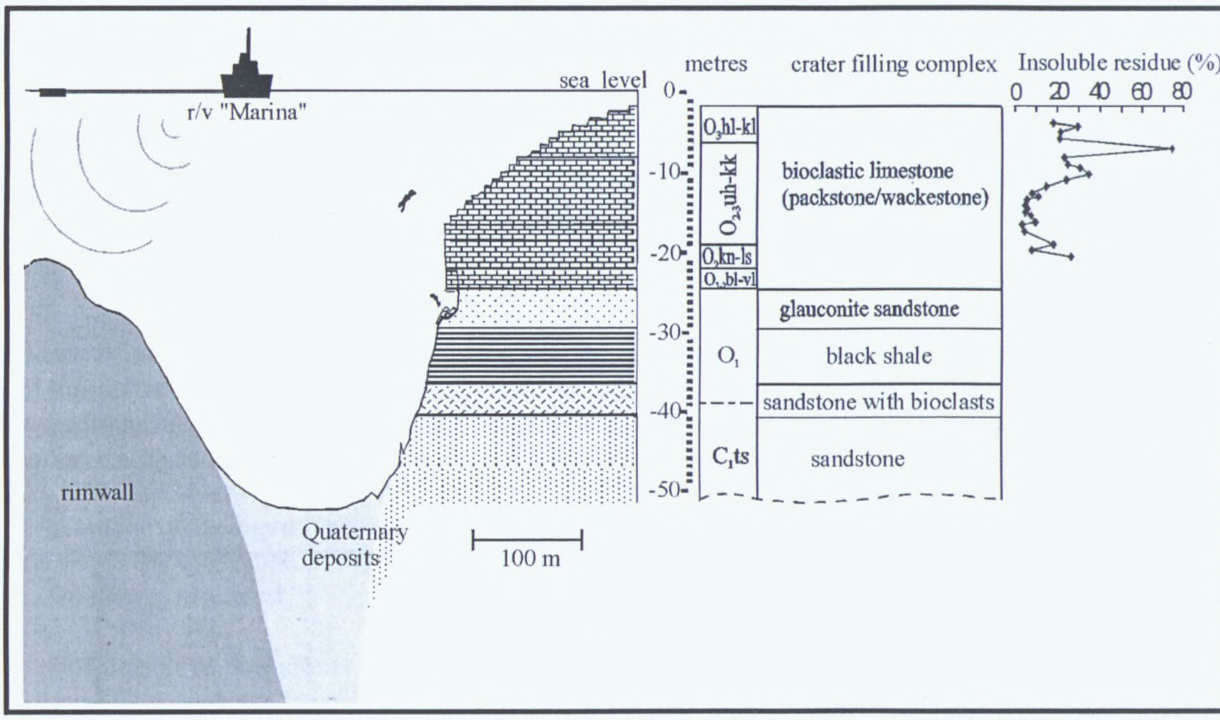

Fig. 2. Topography of the western slope of the crater infilling, the macrolithological log and insoluble residue content of the limestones.

The lower part of the underwater section is represented by the sandstones resembling those of the Lower Cambrian Tiskre Formation. This unit is overlain by the biodetrite-rich sandstones which most likely should be attributed to the Kallavere Formation of the Late Cambrian–Early Ordovician age. The thickness of this unit reaches 4 metres. Two marker beds are appearing higher up in the sequence. The dark brown bituminous argillites (*Dictyonema* Shale, thickness about 7 m) and glauconite sandstone (the Leetse Formation, nearly 5 m) are easily distinguishable.

The limestone succession begins 24 m below the sea level. Its top can occasionally be reached at the depth of 2 m; thus the thickness reaches nearly 22 m.

In both sampled sections, the micropalaeontological investigations were carried out. Ostracodes were identified altogether in twenty samples. In the sample set from the limestone plateau, the samples come from the depth interval of 2.2–21 m. The sample series from the western slope of the crater infilling consists of twelve samples from the depth interval of 14.9–24 m. The results are shown in the Figure 3. The faunal record of the studied sections was compared with the data by Sarv (1959), Meidla et al., 1998 and Tinn & Meidla, 2001.

The sampled sections cover the full extent of the limestone sequence, with some overlap. The overlapping part of the section is identified by the local range of abundant *"Cytherellina" magna"* which occurs in the interval of 16.5–21,0 m in the central-eastern section and between 14.9 and 18.0 m in the northwestern section (see figure 3). In both sections, its appearance level is marked by occasional *Tallinnella dimorpha* and the disappearance level with the first record of the genus *Sigmoopsis*.

The recovered material allows to distinguish the Volkhov Stage at the bottom of the limestone sequence. In the ostracode record, it is characterised by *Rigidella mitis* and *Ogmoopsis bocki*. The thickness of the Volkhov Stage reaches about 2 m like in many sections of northern Estonia. Higher up, the thickness of the Kunda, Aseri and Lasnamägi stages (together) reaches about 3 m. The lower boundary of this interval is marked by the appearance of *Glossomorphites* grandispinosus but more detailed subdivision of the unit is not possible, due to poor faunal record. The thickness of the Uhaku Stage seems to be rather limited in the area, however, the thickness of the Uhaku and Kukruse stages together reaches about 7 m.

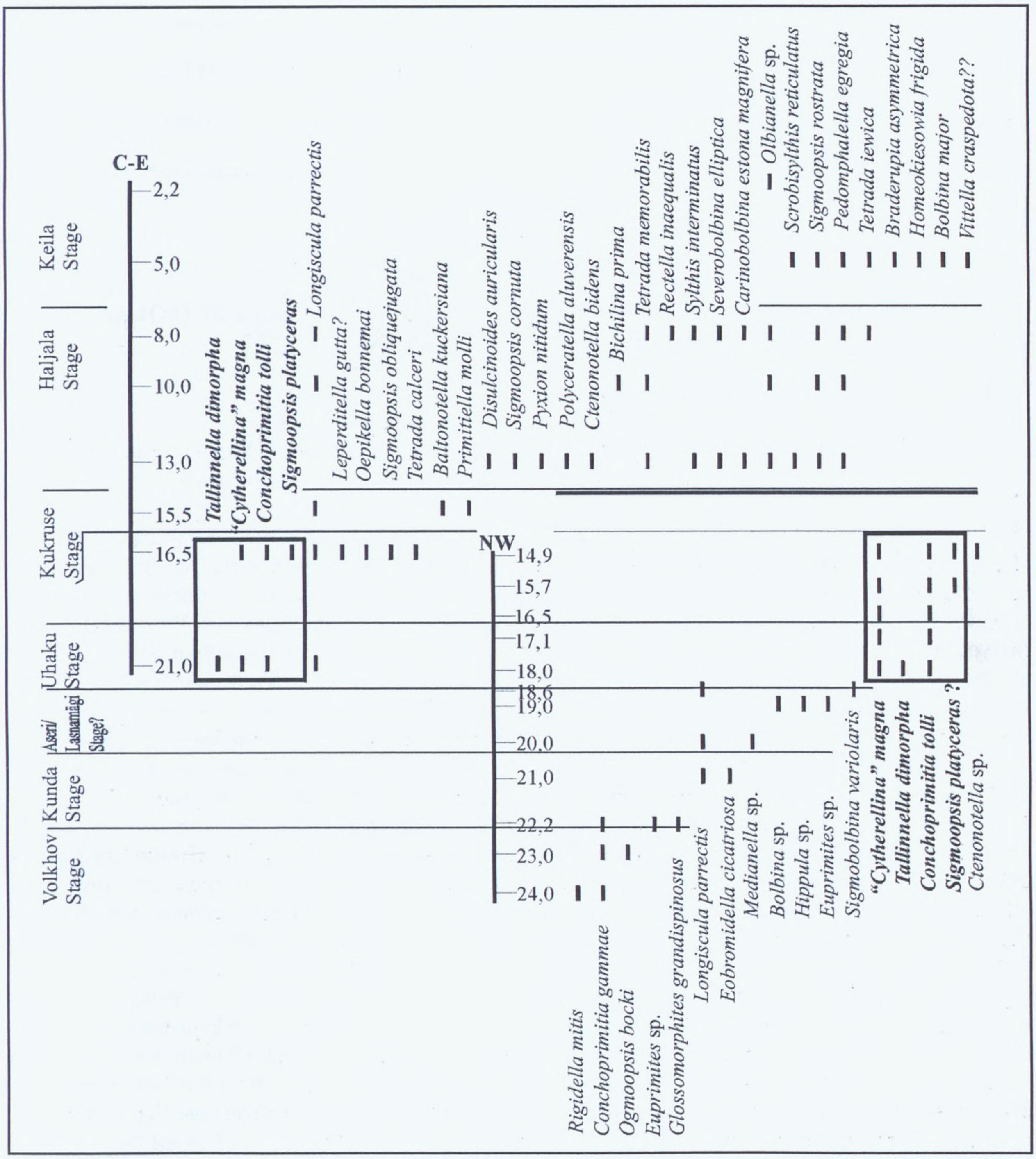

Fig. 3. Distribution of the selected ostracode species in the studied sections, with the stratigraphic interpretation of the sections. Species in bold are used for the correlation of the sections. Abbreviations: C–E – central-eastern section; NW – northwestern section.

The lower boundary of the Haljala Stage is very distinct, being marked by the appearance of a great number of new taxa (*Polyceratella aluverensis*, *Tetrada memorabilis*, etc.) in the central-eastern section. In the northwestern section the corresponding interval was not studied palaeontologically, due to the limited amount of the rock material available. Higher up in the sequence, the disappearance of *Tetrada memorabilis* in the interval of 5.0–8.0 m is most probably indicative of the lower boundary in this particular interval.

As a summary, the youngest sediments in the impact crater infilling are probably of early Keila age. The Neugrund structure must be considered the northernmost locality in Estonia where the Upper Ordovician strata are represented.

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SEDIMENTARY INFILL OF THE POMERANIAN, RADOM – LUBLIN (POLAND) AND PRIPYAT BASINS (BELARUS) DURING THE DEVONIAN

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The Devonian deposits in Poland are broadly extended, mainly under the thick cover of the younger sediments (Fig. 1). They are known from the boreholes on the territory of the Western Pomerania (see Fig. 1) and from some drillings offshore. In these sites, the Devonian beds rest unconcordantly on the Caradoc or Silurian and are covered by the Carboniferous or Permian. The Devonian sequence starts from the Upper Emsian (Turnau & Matyja, 2001). The occurrence of the older Devonian deposits is not palaeontologically proven.

The Devonian is known offshore from some boreholes drilled in the Polish Economical Zone of the Baltic located about 80 km North of the Hel Peninsula. In this case the Lochkovian (oldest Devonian) deposits rest on the youngest Silurian (Pridolian) ones. On the contrary, the Devonian deposits are not known southwards, on the territory of Eastern Pomerania (see Fig. 1). Most probably they were epigenetically eroded.

The boreholes due to technical reasons do not penetrate the Devonian in Central Poland.

The Devonian beds are relatively well investigated by the numerous boreholes and by a lot of seismic lines in the Radom–Lublin area, SE of Warsaw. In this area the Lochkovian strata rest conformably, without break, on the Pridolian ones. The Devonian sequence is stratigraphically full. Due to epigenetic erosion, the Visean or Permian or Triassic or Jurassic unconcordantly covers the different Devonian beds. The Devonian structures, beds and strata continue to SE on the territory of Western Ukraine, where occur in the Volhynia, Podolia, Lviv area and crop out on the Dniester River and its tributaries.

Southwards of the Radom–Lublin area, the Devonian crops out in the Holy Cross Mountains, near Kielce (**K** – see Fig. 1) and in the vicinity of Cracow (**Kr** – see Fig. 1). The Devonian is also known under different cover of younger deposits in other parts of SE Poland. It is known, even to S of the Carpathian Front (**CF** – see Fig. 1), under the Carpathian nappes, and as the exotic blocks and boulders within the Carpathian Flysh. The Devonian beds on the hypothetical territory of Medrzechow – Brzozow Land (see Fig. 1) probably never have been deposited.

In the SW Poland, the Devonian crops out in the some sites in the Eastern and Western Sudetes Mountains and is known from rare boreholes southwards of the Variscan Front (see Fig. 1).

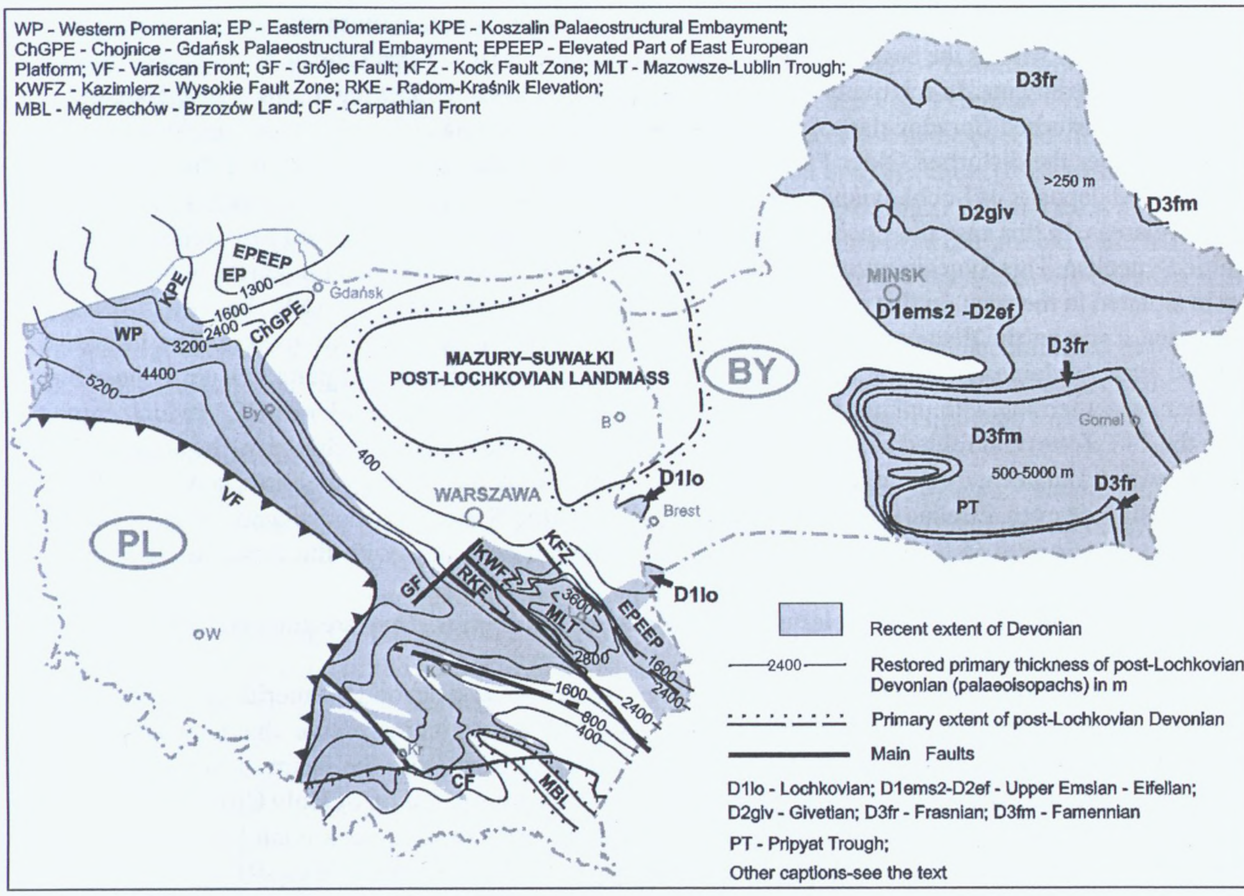

Fig. 1. Sketch map of Devonian in Poland and in Belarus.

The Devonian is also broadly extended on the territory of Republic of Belarus. Recently, the oldest Devonian deposits (Lochkovian – see Fig. 1) are known only from the very limited area nearby the Polish – Belarussian state border. The younger Devonian beds (of Late Emsian till Youngest Famennian age) occur in the eastern half of Republic. They infill the deep Pripyat Trough (see Fig. 1) and some shallow depressions northwards of this Trough. The Devonian deposits in Belarus have been penetrated by the hundreds of boreholes and thousands of kilometres of seismic measurements have been done.

General structural framework of the basins under study was determined by the structure of SW passive margin of the old East European Platform as well as by the relationships between this border and more or less suppositional terranes docked to this margin. The above mentioned relationships are still not fully understood. Even the strict delineation of the margin of EEP has been the matter of different meanings. There are opinions that the Teisseyre–Tornquist Zone is not the border of the EEP and this border is situated far south-westwards.

G. V. Zinovenko (1986) who compiled one of first broad interpretation of the geological structure and history of deposition along this border zone considered south-western border of EEP as the Baltic – Dniester zone of peri-cratonic depressions. Similar opinion have also the Polish geologists which published the recent summary of knowledge of the geology of Polish part of this zone in the collective work: "Sedimentary Basin Analysis of the Polish Lowlands" (1998). Similarly, the comprehensive characteristic of geology of Belarus is done in fundamental work "Geology of Belarus" (2001). The present writers do not pretend to resume these works. We want only to give some remarks on the common history of deposition in the Poland and Belarus during the Devonian and to point out the similarities and the differences.

The Radom – Lublin Devonian sedimentary basin and its prolongation Lviv – Dniester one was developed on the SW margin of EEP. The model of the flexural bending of the plate margin induced by the docking of some supposed terranes from SW is oversimplified and can be considered as the first approach. The models of rift basin or pull-apart basin are also possible.

The geotectonic setting of the Western Pomeranian Devonian basin is more mysterious. Firstly, we did not know exactly what is the basement of Devonian beds and what is the exact age of beginning of deposition of Devonian sediments. In relatively rare boreholes piercing down the Devonian, the non-metamorphosed but slightly disturbed Caradoc dark shales (in some cases the Silurian ones) lie under the Devonian beds. May be, under the disturbed Older Palaeozoic rocks which could be overthrusted, rest the epicontinental undisturbed deposits of Lochkovian and Older Palaeozoic similar to those from the Eastern Pomerania and offshore areas. In this case the Lochkovian basin on Western Pomerania area could be extended to NE into Baltic Syneclise. This Syneclise would exist in that time as the paleostructural embayment of the main broad basin situated in the west. In this case the crystalline basement of EEP extends to SW, beyond of Teisseyre–Tornquist Zone on the distance of several tens or even several hundreds of kilometres. The influence of docked East Avalonia terrane on the Baltica (as the part of Fenno-Sarmatia) plate margin is possible. In another case a terrane with unknown basement has been collided with the border of EEP (which coincides with the T–T Zone) and folded basinal and shelf deposits of Old Palaeozoic. In this case, the area of Western Pomerania and adjacent offshore, became the uplifted, eroded territory during Lochkovian Age and later, till Late Emsian or even Eifelian. Early Devonian basin of Baltic Syneclise was isolated. In both cases, the tectonic events occurred here between Caradoc and Eifelian coincides with the Acadian tectonic phase *sensu* W. S. McKerrow et al. (2000).

The geotectonic setting of the Belarussian Devonian basins is clear. They are situated on the EEP, fairly distant from its SW border.

The Lochkovian basin in Belarus had not very big extent to the east beyond the meridian of Brest (Fig. 1). Its extension was inherited from the Late Silurian basin and, due to its regressive character was narrower than Silurian one. The connection of Belorussian Lochkovian basin with the basin in Volhynia, Podolia, Lviv area (Ukraine), in EPEEP, in MLT, on the RKE and in Lysogory Region of Holy Cross Mountains near Kielce (**K** – see Fig. 1) is unquestionable. The sedimentary infill of the Lochkovian basin represents the regressive part of the Silurian – Early Devonian Transgressive-Regressive Cycle (T–RC) and is classically developed as mainly siliciclastic, coarsening upwards sequence. This sequence begins from marine claystones and is finished by the red fluviatile siliciclastics of ORS type.

During the Late Lochkovian, Pragian and Early Emsian, the Belarussian Massif and Ukrainian Massif (Ukrainian Shield) have been uplifted. This uplifting can be connected with eustatic lowering of sea level. The accumulation of sediments did not take place on the Belarus territory. Most elevated south-western extension of Belorussian Massif in Poland (which in Polish geological literature is known as Mazurian Massif or Mazurian–Belorussian Anteklise) has been uplifted too and along the line of deep-seated Grojec Fault (GF) closed the spreading of the sedimentary basin to NW. On our map (Fig. 1), it is named as Mazury – Suwalki post-Lochkovian Landmass (MSpLL). The sedimentary basin was developed in the Radom–Lublin area. The sedimentary infill of the basin in this time was also represented by the ORS type fluviatile deposits, which were accumulated under the condition of balanced rates of accumulation and subsidence. In the result up to 1500 m of uniform deposits have been accumulated.

The Late Emsian and Eifelian basins had transgressive character. The sedimentary infill has transgressive character too. The deposits are lithologically different. They are represented by the shallow marine carbonates and siliciclastics, by the lagoonal evaporates, even by the salt (in Belarus), by the deltaic siliciclastics and by the red fluviatile ones. The latter two depositional systems are especially well developed in Western Pomerania. The distribution of facies and thickness suggests that the transgression in Poland had the direction from SW to N and NE. In Belarus, the transgression had the direction from E and SE to N, W and S. During the Eifelian, the basins in Poland and Belarus have been connected each other. The primary thickness of deposits was not very big and rarely exceeded few hundred metres. The deposition of the sediments was topographically controlled. The average rate of subsidence was moderate but temporarily different. The minor T–R cyclicity is well visible in the Upper Emsian – Eifelian profiles.

The Givetian was the age of the new strong transgressive pulse in Poland (Milaczewski, 1981; Narkiewicz & Narkiewicz, 1998; Narkiewicz et al.,1998b, Matyja, 1998). This event took place in the *varcus* time (Pelcza Mb). The transgression had the general direction from west to east. The maximal transgressive pulse was recorded as a distinct carbonate bed with the marine different fauna in the SE part of Radom–Lublin area and even far eastward, in Volhynia (Ukraine). The pattern of basin-infill in the Radom–Lublin area was subordinated to structure of the SW border of EEP. The thickness of sediments increases from NE to SW. It is smallest in the elevated part of EEP (see Fig. 1). It is mean in the Mazowsze – Lublin Trough (see Fig. 1) and become bigger, up to 600 m, in the Radom – Krasnik Elevation (see Fig. 1) and in Lysogory Region, north of Kielce (**K** – see Fig. 1). Also the Sand/Shale and Clastic Ratios decrease in the direction from NE to SW.

In Belarus, the fairly uniform marine terrigenous deposits (up to 200 m of thickness) represent the Givetian (Polotsk Horizon) [Kruchek in Makhnach et al., 2000]. Pripyat Trough (see Fig. 1) did not exist in Givetian, so the thickness of deposits are also not very different. *Nota bene*: in this time did not exist the MLT and RKE as separate structural units in the Radom – Lublin area. Similarly to the latter area the Givetian sediments in Belarus were deposited in one T–R cycle.

In Western Pomerania and adjacent offshore the Givetian deposits were also accumulated in one T–R cycle, with abundance of carbonates in middle part of the cycle and prevailing of siliciclastics in the lower and upper ones (Matyja, 1998). The general direction of the Givetian transgression was from SSW to NNE. The thickness changes in this direction from about 600 m to 400 m.

In Frasnian, the geological revolution took place in Western Pomeranian, Radom – Lublin and Pripyat Trough areas. In the first area, the deep dysaerobic shelf basin was developed in the *punctata* time, probably due the increased subsidence and extension on the SW border of EEP. In the Radom – Lublin area, the increasing of subsidence took place a little later in Late Frasnian. The developing of MLT and separating of the EPEEP and RKE along the Kock Fault Zone (Fig. 1) and Kazimierz – Wysokie Fault Zone (Fig. 1) was induced by the transpression of the border of EEP (Narkiewicz et al., 1998a). This event was connected with the broad Frasnian transgression and with the start of "carbonate factory". The development of the deep shelf basin, known from Frasnian in Western Pomerania, started in MLT in the Early Famennian (Bychawa "Formation"). The beginning of rapid subsidence and development of Pripyat Trough as a rift between the Ukrainian Shield and Voronezh Massif took place in Late Frasnian. The extension of the basement was connected with the fast accumulation of salt complex and strong basic volcanism events.

Famennian basin-infill of the three basins described was different. In Western Pomeranian basin some influences of proximity of eroded Baltic Shield are visible in Upper but not Uppermost Famennian deposits. These deposits in northern part of Western Pomerania and offshore are represented by the mixed siliciclastic/ carbonate, in places, evaporitic peri-littoral sediments. Other Famennian deposits are the products of "carbonate factory". They are developed as different limestones and marls, also the characteristic nodular ones. Two minor T–R cycles were distinguished in the Famennian profile (Matyja, 1998). Upper cycle is extended to the Tournaisian Time. In the Radom – Lublin area, the deposition of sediments took place in one big T–R cycle, which begins from the shelf basin sediments (Bychawa "Fm."). The characteristic nodular limestones (Firlej Fm.) represent the middle part of the cycle as the carbonate – marly ramp deposits. The upper one is represented by the dolomite marls as the shelf basin deposits (Niedrzwica "Fm."). The vertical movements along the KFZ are manifested by the deposition of carbonate sandstones, marls and mudstones of Hulcze Fm. This cycle is extended to the Tournaisian too. The Famennian has a considerable thickness about 1500 m. The rate of subsidence was big and different.

In the Pripyat basin, the Famennian deposits are characteristic for next pulse of rifting of the area. The very thick profile can be subdivided into three parts: lower is the product of "carbonate factory", that is very different limestones and marls (Infra-Salt Beds). During the early Famennian the huge basic, very deep-seated volcanism events took place too. The Middle Famennian is developed mainly as the enormous complex of different salts (Upper Salt-bearing Beds). Mainly the shallow marine terrigenous deposits represent the Upper Famennian. The total thickness of the Devonian deposits in the Pripyat Trough achieves 5000 m, whereas in the other part of Belarus does not exceed 1000 m (Kruchek in Makhnach et al., 2001).

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POST-HOLSTEINIAN INTERGLACIALS – ONE OR MORE?

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The upper part of the stratigraphical schemes of the Quaternary deposits in all the three Baltic states compiled before the mid- 1990s are rather similar. Within the second half of the Middle Pleistocene, two glacial episodes – Sangaste/Letiza/Dainava/ and Ugandi/Kurzeme/Žeimena separated by Karuküla/Butėnai/ Pulvernieki interglacial deposits – are distinguished on the basis of two till beds traceable in the Baltic region. In the scheme of Lithuania one more stratigraphic unit tentatively assigned to the separate Late Middle Pleistocene (Snaigupėlė) interglacial is introduced by Kondratienė (1973) within the penultimate glacial period dividing it into two independent – Žemaitija and Medininkai – glaciations. However, our recent electron spin resonance (ESR) dating of freshwater mollusc shells from one of the parastratotypical sections of the Snaigupėlė Interglacial at the Valakampiai site (Molodkov, Bolikhovskaya, Gaigalas, 2002) revealed that these deposits were actually formed during the last (Merkinė/Eemian) interglacial stage. The attempts to correlate the Baltic stratigraphical schemes with the global oxygen isotope-based reference levels have also failed. The isotope record demonstrates a more complicated palaeoclimatic pattern than that presented in the Baltic schemes. It can be explained either by the absence of some palaeoenvironmental episodes in the Late Middle Pleistocene history of the Baltic countries or by the circumstance that these evidence-based deposits have not yet been recognised here. The latter explanation seems more reasonable because the glacial erosion could have been very intensive on the territories of the Baltic countries, especially in Estonia and Latvia, and, therefore, the time interval reflected in a particular stratigraphic section at a given locality may consist of a number of hiatuses. As a result, only a small portion of the geohistorical record is actually preserved here. Due to the absence of reliable criteria for timing and recognising coeval glacigenic deposits and a wide variability of their compositional and physical characteristics, the interregional correlation of the existent lithostratigraphic units is significantly hampered.

During the last years, a palyno-chronostratigraphic framework has been developed for Northern Eurasia on the basis of two independent sources of palaeoenvironmental information: electron spin resonance (ESR) chronology of warm-climate-related deposits and palynological record of vegetation response to climatic variability and palaeoenvironmental events (Molodkov, Bolikhovskaya, 2002). It allowed us to create a record of palaeoenvironmental changes and provide absolute chronology for continental-scale climatic events over the past 600,000 years. The correlation of palynostratigraphic sequence with the ESR-chronological record suggests that the distinguished climate episodes have a large regional, or even global significance. Therefore, such palyno-chronostratigraphic framework may offer an excellent guide for searching of missing stratigraphical units equivalent to palaeoenvironmental events revealed. Hence, from our point of view, some of problems related to the Middle Pleistocene stratigraphy in the Baltic states could be solved now by a twofold approach. First, by using this record as a palaeoenvironmental reference for the Central and Eastern Europe and the Baltic area. Second, by carrying out purposeful search for the deposits correlative with the main palaeoclimatic markers recognised by us in the neighbouring and more remote areas, and by dating them with the latest and most promising geological dating methods – ESR and OSL (optically stimulated luminescence).

In the paper we shall also discuss the palaeoenvironmental and temporal structure of the time interval under consideration and potential possibilities for providing geochronological constraints on stratigraphic units of interest.

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SUPRACRUSTAL ROCKS IN THE CRYSTALLINE BASEMENT OF LITHUANIA: IMPLICATION FOR PALAEOPROTEROZOIC STRATIGRAPHY

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The International stratigraphic chart, ratified by IUGS suggest subdivision of Proterozoic sequence in to three erathem and nine periods (Plumb, 1992); Explanatory note…, 2000). The basic principles to be used in subdividing Precambrian should be the same as for Phanerozoic rocks (Hedberg, 1994).

However, early Precambrian stratigraphy is complicated by obvious lack of fossils and deformation of stratified sequence. It is particularly complicated in terranes involved in orogeny and suffered high-grade metamorphism, and deformation of whole sequence by folding, faulting and block movements. Therefore reconstruction of Precambrian stratigraphy requires special approach and methodology. Particular tasks are identification of primary origin and type of supracrustal (sedimentary and volcanic) protolith of metamorphic rocks; estimation of the source rocks and provenance area of sediments; degree of alteration of primary material. Particularly important is identification of tectonic setting and tectonic facies of sedimentation and volcanism (Robertson, 1994). Based on regularity of plate tectonic processes there are possible to apply the methods of tectono-stratigraphy and to distinguish at least the largest stratigraphic unites (groups, series, formations) and to estimate their relative age. Radiological dating of particular lithologies of such unites or intrusive rocks cutting them can provide more certain limits of their age.

These tasks might be achieved by integrated interpretation of the results of various methods: petrological, geochemical, izotopical and other including dating, as well as regional correlations.

In Lithuania, situated on the Western margin of the East European Craton, the crystalline crust has been formed in course of Svecofennian orogeny (Mansfeld, 1996). The oldest supracrustal rocks suffered metamorphism of amphibolite to granulate facies, reaching level of ultrametamorphism with partial melting and migmatisation, in places also local metasomatism and later tectonisation. These rocks appear mainly in relic bodies among granitoid, charnockitoid and other intrusive missives, or in form of palaeosome in migmatites. Nevertheless they are only representatives of primary supracrustal sequence and provide possibility for the reconstruction of its lithological composition, formation history and stratigraphy.

The purpose of this work is to characterise primary sedimentary and volcanic rocks in crystalline basement of Lithuania and implications on environment and age of their formation and stratigraphy. New ICP analysis made in ACME Analytical Laboratories and summary of previous data have been used.

The crystalline crust of Lithuania belongs to two main domains – West Lithuanian Granulite Domain (WLD) and East Lithuanian Domain (ELD) (Skridlaitė, Motuza, 2001). These domains differ in upper lithosphere structure, formation environment and evolution history. The transitional zone between these domains is identified as Middle Lithuanian Suture Zone (MLSZ), having some specific features of lithological composition.

In ELD the primary supracrustal sequence might be reconstructed as formation of interbedding of felsic gneisses and amphibolite or mafic granulite, with rear beds of marble. The proportions of these lithologies varies in different places of the area, but in general mafic volcanics are predominant, while felsic gneisses are more abundant in limited areas.

In WLGD predominant supracrustal rocks are felsic to intermediate gneisses, while mafic supracrustals are almost lucking, presenting rear beds among the metasediments or xenolites in intrusive rocks.

In ELD main lithological varieties of felsic gneisses are: biotite-plagioclase-quartz gneisse rarely with garnet, sillimanite; muscovite-biotite-microcline-plagioclase-quartz gneisse with sillimanite; amphibole-biotite- plagioclase-quartz gneisse. Rocks are characterized by fine and even grained lepidogranoblastic texture, banded structure, caused mainly by variation of mineral composition. Such banding is regarded as primary bedding. In places there are present rounded or lens shaped fragments of quartz, quartz feldspar, in places with sillimanite composition, reminding deformed pebble or pyroclastic lapilli (wells 403, 716, 719). Zircon is typically well-rounded, lucking primary magmatic crystallographic features and might be detrital. It is confirmed by age of particular fractions of zircon (Pb-Pb and U-Pb method) varying from 2.16 till 2.71 Gy. (Marfin i dr., 1987; Mansfeld, 1996). The chemical composition implies psammitic (grywackes, arcoses) or semi-pelitic (shale) protolith possibly with amendment of felsic volcanic material in some beds. Some gneisses might be volcanogenic.

The REE abundance diagrams form rather uniform swarm with small negative Eu anomaly, characteristic for post-archaean metasediments. The trace element diagrams indicate better affinity to North American shale as to Postarchaean Australian shale.

The mineralogical and geochemical data indicate that the principal source of ELD metapsammites might be local mafic and felsic volcanic rocks, as well as granitoids from the distant Archaean continental crust. The last is suggested by detrital zircon ages presented above and eNd values (-2 in metapsammites in well 403) indicating blending of Archaean (-9) and Palaeorozoic (+4 in amphibolite, well 404) material (Mansfeld, 1995). The clastic material in metapsammites suffered week chemical weathering what also indicate prevailing its local source.

Amphibolites and mafic granulites are fine, even grained, nematogranoblastic with relics of primary volcanic textures (porphyry, ophitic). Petrochemical composition suggests basalt, rarely komatiitic basalt protholit, mainly of tholeiitic affinity. The REE diagrams and other geochemical data imply the tectonic setting of mafic volcanics in the active marginal part of the oceanic lithospheric plate in volcanic island arc tectonic facies, in the subduction related tectonic environment. The geochemical signatures of subduction zone increases in Western part of ELD, in the realms of MLSZ, forming border zone between ELD and WLGD.

The supracrustal sequence might be the part of the accretionary prism, where alternation of deformed and tectonically juxtaposed blocks (wedges) of metavolcanic and metasedimentary rocks takes place. Interbedding of mafic and felsic supracrustals indicates their formation in roughly same time span.

The WLGD felsic supracrustals are of two main types: Bt-CPx-Pl-Qtz gneises, presumably metapsammites and alumina rich Sil-Grt-Bt-KFs-Pl-Qtz +/-Hz, Mag, gneisses, presumably metapelites. The presence of felsic volcanics or volcanoclastic rocks is suspected, but not proven with certainty.

Reconstruction of protholith of these rocks suggests Fe-shales, shales, wacke. Geochemical features of metapelitic gneisses differ, by essentially higher content of Fe and Ti, negative correlation between Al_2O_3 and $CaO+Na_2O$ and positive between Al_2O_3 and FeO, what is characteristic for pelitic rocks. On the $CaO+Na_2O$ – Al_2O_3 – K_2O triangle diagram reflecting weathering of primary clastic material, WLGD metapelites exhibit much higher degree of exogenic alteration of protholith.

Many geochemical patterns of WLGD metapsammitic rocks are comparable with those of ELD. Nevertheless, in general, supracrustals of WLGD clearly differ from those in ELD by lower alkali and higher Fe content in respect to alumina (FeO versus Al_2O_3; Alkalies versus Al_2O_3 diagrames), higher content of Co, Ni, Ti and other elements characteristic for mafic magmatic rocks. Geochemical data suggests more uniform and predominantly mafic source of clastic material for the formation of protholith of WLGD metasediments.

Mafic metavolcanic rocks are subordinated in WLGD, forming small bodies between felsic supracrustals or xenolites in granitoids and charnockitoids. Their geochemical features are similar to those of most evolved metabasites in ELD. They are mainly tholeiitic or approaching to calc-alkaline.

The tectonic setting of WLGD supracrustals is less evident. Geochemical data suggest formation of mafic metavolcanics in subduction related volcanic island arc environment, identical as mafites of ELD. Nevertheless REE abundant diagrams normalized by Primitive mantle not always exhibit negative Nb anomaly, characteristic to subduction zone. So it might be presumed, that WLGD also form part of subduction related assemblage, but the relation to subduction zone is not so close. There is possible presumption on formation of WLGD supracrustals in forearc or back-arc basin.

The possible time span of supracrustals can be estimated based on following data and speculations. The Sm/Nd model ages estimated both in Lithuania and surrounding areas indicate Palaeoproterozoic age of the crust. So its formation might be with certainty related to Svecofennian orogeny (Mansfeld, 1995). The lover limit of formation of Svecofennian supracrustals in Finland and Central Sweden is 1.95–1.904 while upper 1.888–1.87 Gy (Nironen, 1996). In Belarus the estimated age of oldest Okolovo volcanics is 2.0 Gy (Taran, 2001). U-Pb age of felsic gneisses in Staicele, Latvia, interpreted as metarhyodacite is estimated as 1.87 Gy (Mansfeld, 1996). In Lithuania there is only one U-Pb in zircon data of metadiabazes in well Lazdijai-4 – 1.864 Gy (Rimsha et al., 2000). This is subvolcanic rock which might be related to volcanism, but can be also later in respect to supracrustal rocks. So its age cannot be safely attributed to supracrustals. The upper possible limit of the age of supracrustals indirectly is indicated by intrusive granitoids and charnockitoids dated in adjacent areas of Belarus as 1.9–1.85 Gy. The age of felsic intrusives in WLGD is 1.81–1.8 Gy (Mansfeld, 1995; E. Bibikova, pers. com.). Same age is of monazite in migmatites in supracrustal sequence indicating time of peak metamorphism. It means that the sedimentation took place essentially earlier, probably before 1.85 Gy (Skridlaitė et al., 2001). These dates indicate the possible interval of the formation of supracrustals in ELD and WLGD as 2.0–1.85 Gy. Thus, in spite of possible differences in age the formation of supracrustals in both domains of the crystalline basement of Lithuania on the recent level of knowledge fall in same time span, which corresponds to Orasirian period of Palaeoproterozoic – 2.05–1.80 Gy.

Summarising, supracrustals of Lithuanian crystalline basement can be subdivided in three petrotectonic units. All being parts of subduction related assemblage, formed in different tectonic facies.

ELD unite, composd by interbedding mafic volcanics and psammites, presumably have been formed in accretionary prism and/or forearc basin tectonic setting.

MLSZ unite, composed predominantly by mafic volcanic and subordinated intermediate volcanic or volcanoclastic rocks, probably represent eroded volcanic arc complex.

WLGD unite, composed predominantly by metasedimentary and, probably, felsic to intermediate metavolcanics might be attributed to forearc or back arc basin environment.

All these petrotectonic unites, being parts of same subduction-related assemblage fall in the same time span – 2.0–1.85 Gy, which corresponds to Palaeoproterozoic Orasirian period.

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INFLUENCE OF CALEDONIC VOLCANIC ACTIVITY FOR MARINE PALEOENVIROMENT: EVIDENCE FROM THE LLANDOVERIAN GRAPTOLITES OF BALTIC SEDIMENTARY BASIN

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During Cambrian–Silurian time the Baltic sedimentary basin was a foreland basin. All that time the width and the depth of Baltic sedimentary basin was under the influence of marginal tectonic processes of Baltica paleocontinent. The docking of Avalonia paleocontinent and the movement of Avalonia Caledonides on Baltica continent, most significantly increased the shape of Baltic sedimentary basin (Lazauskiene, 2000 after Sliaupa et al., 1997). Colision-subduction processes were accompanied by the volcanic activity of Reic Ocean island arcs and continent margin.

The numerous layers of ilitic clays (metabentonites) we could find in Baltic sedimentary sequence as the evidence of Ordovician–Silurian volcanic activity. Metabentonites represent the relicts of volcanic iron enriched dust ore ash, transported by atmosphere from the active continental margin volcanoes. Such fine volcanic material could form the layer only in the quiet and deep part of the basin. In Kurtuvėnai-161 drill core there are numerous metabentonites beds with thickness from 1 to 50 mm. Those layers are distributed in argillite matrix, which was deposited in the basin with depth up to 200 m. The mass of piroclastic material use to be really huge for to form the 5 cm thick layer in such depth keeping in mind the distance from source. This volcanic pollution has an influence on the marine environment and disappearance of some graptolite specimens and species.

The graptolite fauna was studied in Kurtuvėnai-161 borehole. This borehole is in the West of Lithuania, in the zone of clayey facies of the Baltic sineclyse. In the central part of the Baltic sineclyse Llandoverian graptolites are found from ***cyphus*** to ***lapworthi*** zones. Llandoverian succession in this borehole is the most complete. Llandoverian graptolites were studied in 75 samples in Kurtuvėnai-161 borehole. There were found about 60 species and subspecies of graptolites. 8 graptolite zones were separated from Llandoverian in Kurtuvėnai-161 borehole (Fig. 1). The overall thickness of Llandoverian rocks is 51.9 m.

There were found 4 metabentonitic layers in this section: in the depth of 1474.6 m, 1471.5 m, 1465 m and 1450 m. According to geochemical analysis of Silurian metabentonites, the piroclastic material was the product of mature shoshonitic magmatic series (Fig. 2), from active continental margin tectonic environment (Fig. 3) (Motuza et al., 2001). Most significant feature of this tectonic environment is the huge volcanoes with catastrophic eruptions and worldwide spread piroclatic material. According to classification of piroclastic rocks, the size of piroclastic dust grains is < 1/16 mm (Shelley, 1993). Then the huge mass of piroclastic material can be distributed 1000 km distance within 10 days. The grayish or blackish volcanic dust could pollute the upper part of seawater, which would cause the light permeability decrease and this pollution could stay for the years.

Some graptolites species cross through these layers. The zones with metabentonitic layers are crossed by graptolites that have narrow, weak, spiral, strongly curved colonies (*Campograptus communis*, *Demirastrites convolutes*, *Oktavites planus*, *Streptograptus nodifer*, *Globosograptus crispus*, *Monograptus veles,* etc.). These species of graptolites lived in the deeper layer of water (Bates, Kirk, 1989). Species such as *Monograptus halli*, *Metaclimacograptus hughesi*, *Pristiograptus regularis* that had massive rabdosomes and lived in the upper layer of water do not cross metabentonitic zone.

It was related to the volcanic pollution of the upper layer of the seawater. The decreased of light permeability in the upper layer of water caused decrease of phytoplankton, which was the main food for graptolites and their abundance decreased as well. Volcanic ashes subsided on the sea flora at different speed, they dispersed in the lower beds of the water and ashes influence on organisms was less. Graptolites that lived in the deeper beds of the water demanded less light and probably that predetermined graptolites salvation after sedimentation of the volcanic ashes. Extinction of graptolites related to volcanism had only local nature.

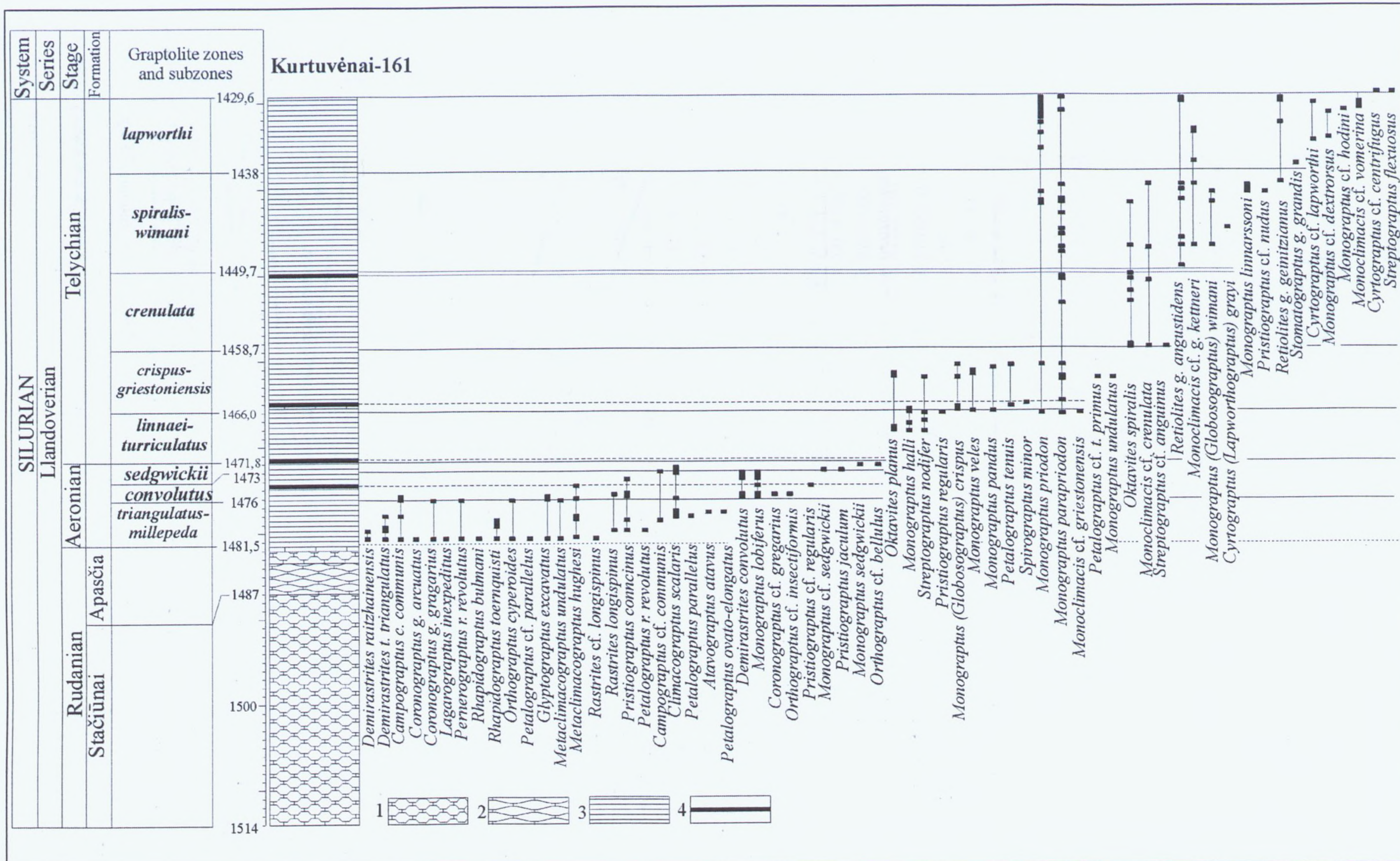

Fig. 1. Distribution of the graptolite fauna and layers of metabentonites in the Llandoverian geologic section (Kurtuvėnai-161 borehole). 1 – nodular limestone, 2 – aphanitic limestone, 3 – argillite, 4 – metabentonite layer.

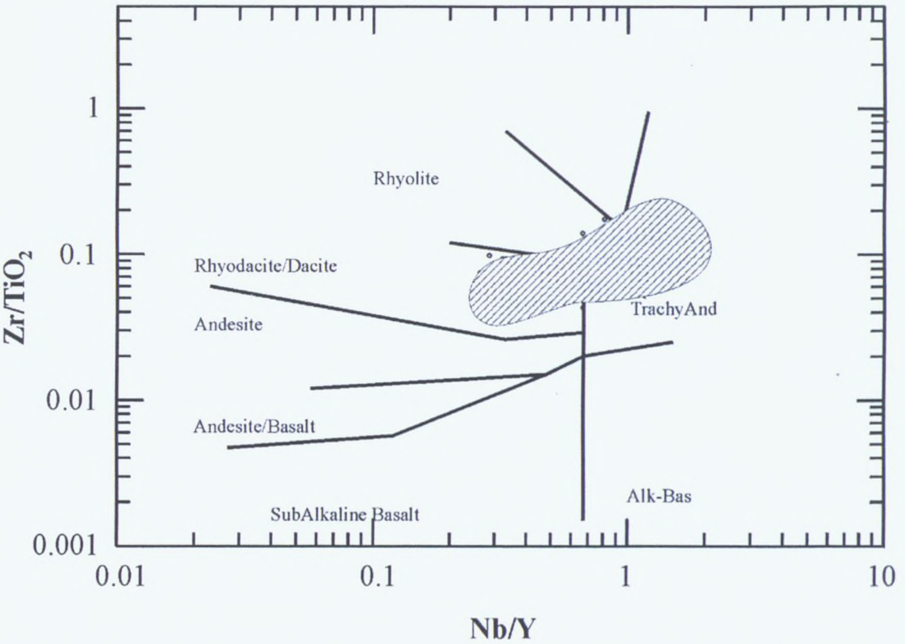

Fig. 2. Primary geochemical composition of volcanic ash (metabentonites).

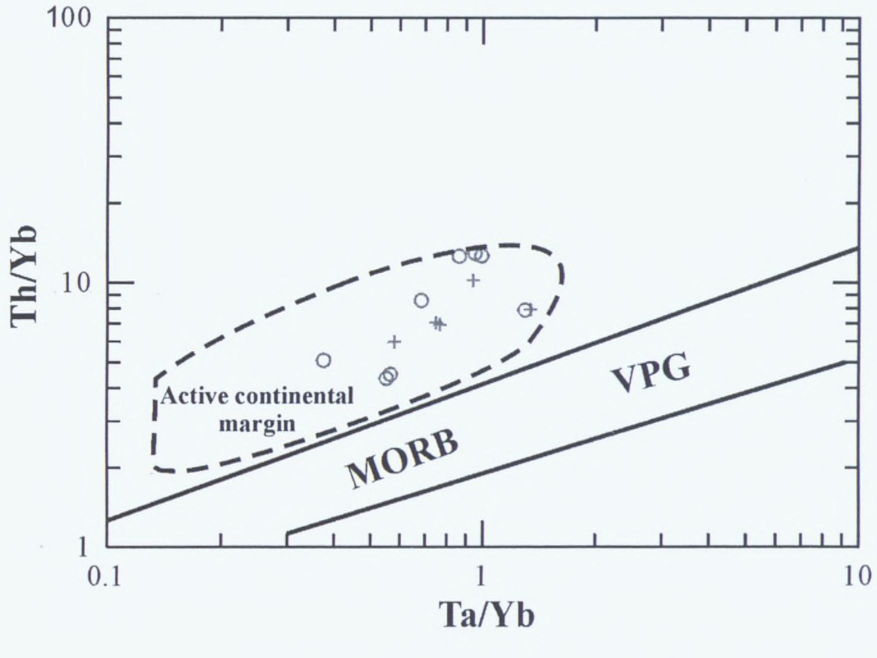

Fig. 3. Tectonic facies of volcanic ash source.

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TECTONIC SETTING OF THE ORDOVICIAN-SILURIAN K-BENTONITES OF THE BALTIC REGION

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The Baltic basin represented the passive continental margin basin during Cambrian–Middle Ordovician that evolved into the foreland basin of the North German–Polish Caledonides during the Late Ordovician–Silurian times. Numerous K-bentonites are documented in the Ordovician–Silurian carbonate-shaly succession of the basin, thus recording volcanic activity along the interacting plate margins. Good stratigraphic control provides possibility to trace major trends in evolution of the volcanism and governing tectonic processes.

Closest to Caledonides the K-bentonites are mapped throughout the whole Ordovician section, while they are confined to Arenig, Lower–Middle Caradoc (most voluminous) and Ashgill intervals only further east. Lateral distribution of K-bentonites indicates transportation of volcanic ash from the western and northwestern sources. Occurrence of K-bentonites sharply increases in the Lower Silurian and they are only scarce in the Upper Silurian. The latter can be related either to secession of volcanic activity or terrigenous dilution. The latter mechanism seems more reliable taking into consideration drastically increased sedimentation rates due to denudation of advancing Caledonian orogen in Late Silurian. Mineral composition of K-bentonites is dominated by mix-layered I–S, the kaolinite composes 0–60%.

REE and trace element composition of Ordovician and Silurian K-bentonites can be used to reconstruct parent lithologies and their tectonic setting. The samples were collected from Estonia, Latvia, Lithuania and Kaliningrad District (eastern half of the basin). They revealed dominant felsic composition of volcanic source rocks. Caradoc K-bentonites show affinity to dacites and rhyolites, while the Ashill bentonites resemble trachyandesites that might be interpreted as an indication of maturation of the subduction processes. Silurian K-bentonites were likely sourced from dacites and trachyandesites (Fig. 1). Trace and REE element geochemical cross-section though Ordovician and Silurian of the Baltic sedimentary basin indicates several distinct sources of the K-bentonites. The Middle Caradoc K-bentonites geochemistry is dominated by calc-alkaline trend, yet revealing the second less distinct tholeiitic source that gradually gave way to transitional one. The Caradocian K-bentonites bear significantly different geochemical features from

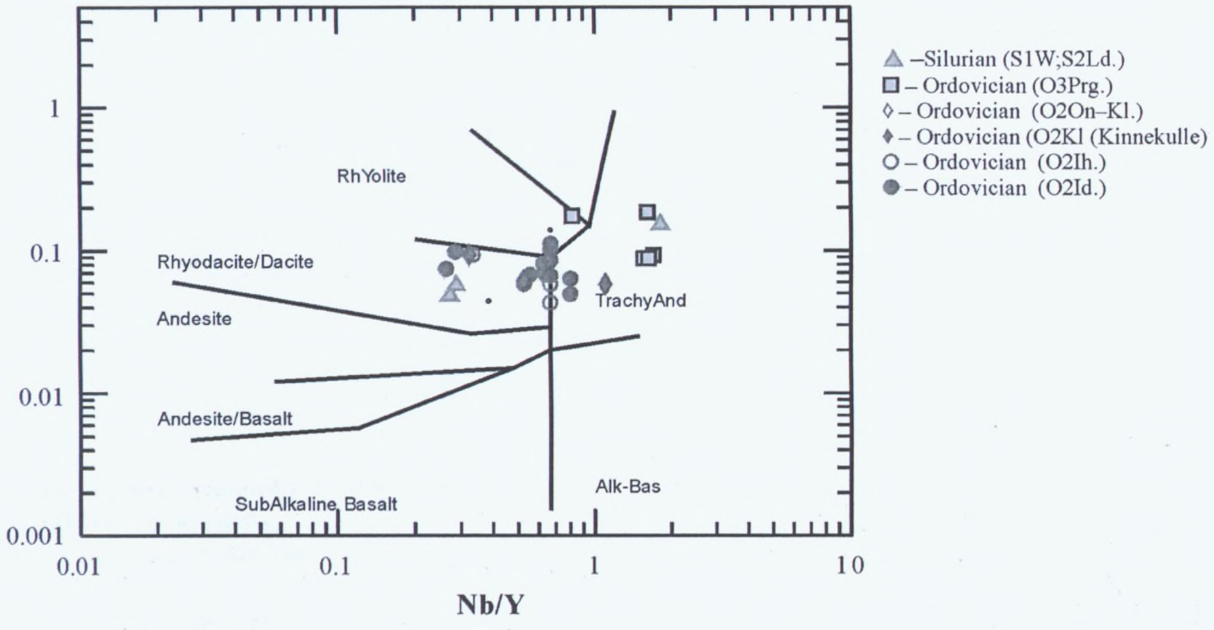

Fig. 1. The discrimimant plot of K bentonites primary combosition.

those of Ashgill and Silurian. The former show affinity to subduction related (continental) volcanic arc felsic rocks (Fig. 2), while geochemistry of the two latters points to collision-related tectonic setting and continental type of volcanic arcs, though some Lower Caradoc and Silurian samples show oceanic setting. Trace elements point to general trend towards more evolved parent lithologies from

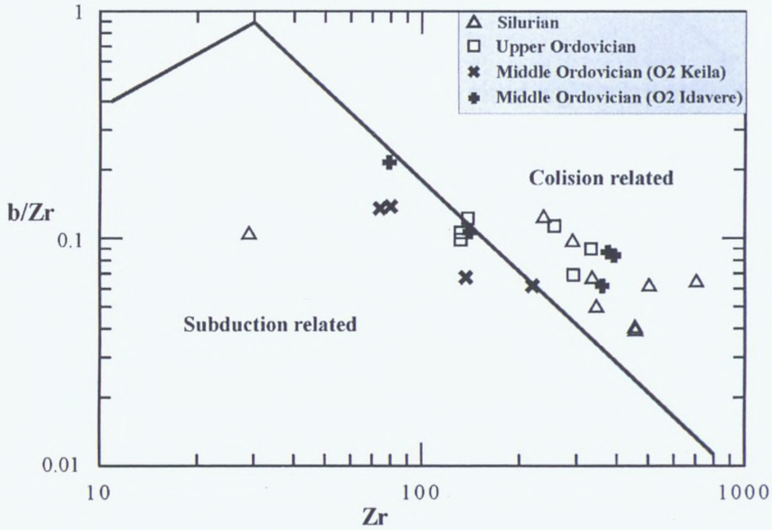

Fig. 2. K-bentonites, parental source relations with plate tectonic processes.

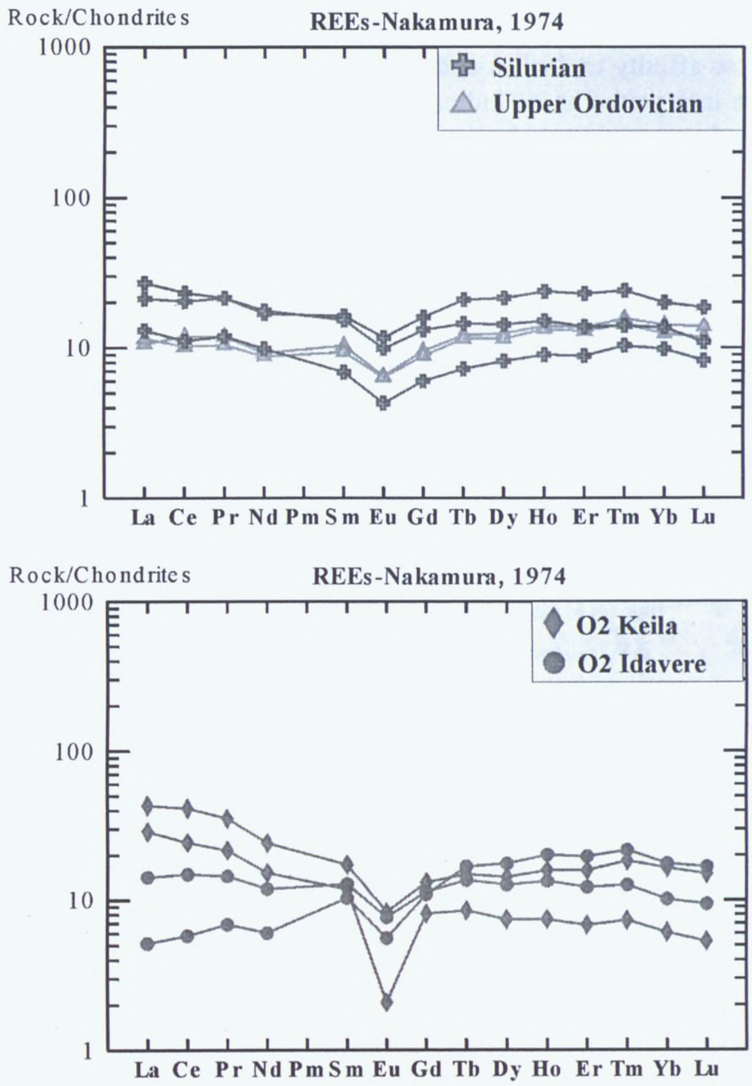

Fig. 3. Chondrite-normalized REE distribution of Ordovician and Silurian K-bentonites.

Lower Caradoc to Ashgil and Silurian. It is remarkable that Ashgil and Silurian K-bentonites show flat REE chondrite-normalized patterns (Fig. 3). One of possible explanations could be inheritance of the geochemical signatures of the subducted oceanic crust.

Geochemical features of Baltic K-bentonites indicate evolved character of volcanic arcs that developed along the converging margins of Eastern Avalonia, Baltica and Laurentia during Ordovician–Silurian, showing maturation trend of the converging system in time. Majority of volcanic ash was sourced from continental island arcs during Caradoc (Fig. 4). Yet, some Lower Caradoc K-bentonites might be products of oceanic arcs that matured in Middle Caradoc. During Ashill and Silurian volcanic activity was dominated by collisional tectonic setting pointing to closure of Tornquist and Japetus oceans during this time.

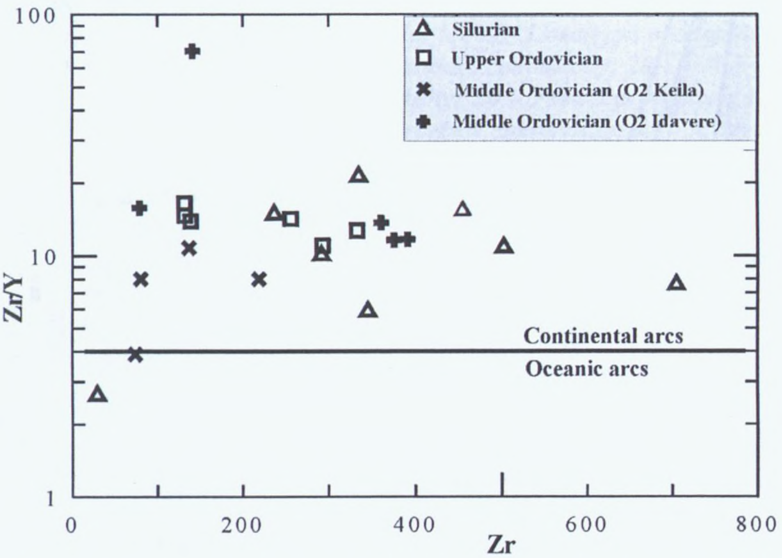

Fig. 4. Discrimant plot of tectonic setting of K- bentonites parent rocks.

THE DEVELOPMENT OF SILURIAN BRACHIOPOD COMMUNITIES IN LITHUANIA

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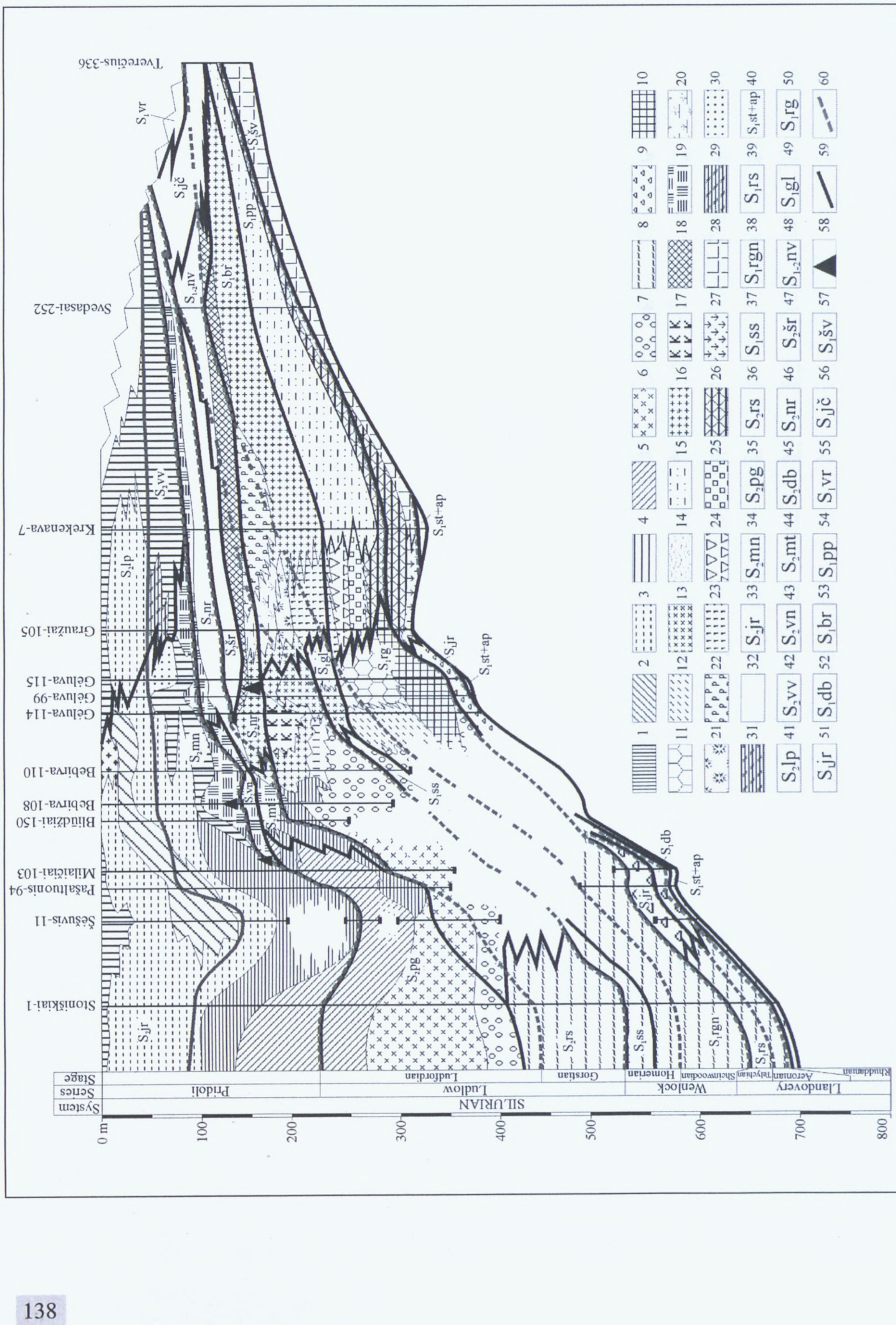
There were no brachiopods recorded from the **Rhudanian** of Lithuania. In **Aeronian** (***convolutus – sedgwickii*** time) only deep water communities which were living under quiet water dysaerobic environments below the wave base are represented. The shallower facies and communities didn't remain in the sections. These low and moderate diversity comminities from offshore to onshore made the folowing lateral sequece: ***pelagic community – "Clorinda" sp.*** (BA 5) – ***Jonsea grayi*** (BA 4–5) – ***Eoplectodonta duvalii-Lissatrypa obovata*** (BA 4–5) (Fig. 1).

In **Telychian** (*crenulata* time) more shallow high diversity communities appeared. The following lateral sequence was established from offshore: **pelagic community** – ***"Clorinda" sp.*** (BA 5) – ***Jonsea grayi*** (BA 4–5) – ***Eoplectodonta duvalii-Lissatrypa obovata*** (BA 4–5) – ***Skenidioides lewisii-Pentlandella tenuistriata-Dicoelosia paralata*** (BA 4) – ***Dicoelosia paralata-Skenidioides lewisii*** (BA 4) – ***Gotatrypa hedei*** (BA 3).

During **Sheinwoodian** in ***riccartonensis*** time few new dysaerobic communities appeared instead of ***Eoplectodonta duvalii-Lissatrypa obovata*** community making up lateral sequence **pelagic community** – ***"Clorinda" sp.*** (BA 5) – ***Jonsea grayi*** (BA 4-5) – ***Lissatrypa obovata-Plagiorhyncha depressa*** (BA 4) – ***Lissatrypa obovata-Skenidioides lewisii*** (BA 4) – ***Dicoelosia paralata-Skenidioides lewisii-Lissatrypa obovata*** (BA 4) – ***Dicoelosia paralata-Skenidioides lewisii*** (BA 4). At the end of **Sheinwoodian** together with the basin regression shallow water communities appear making up onshore lateral sequence **pelagic community** – ***Jonsea grayi*** (BA 4–5) – ***Lissatrypa obovata-Plagiorhyncha depressa*** (BA 4) – ***Lissatrypa***


System
Series
Stage
SILURIAN
Llandovery
Wenlock
Ludlow
Pridoli
Rhuddanian
Aeronian
Telychian
Sheinwoodian
Homerian
Gorstian
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0 m
100
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Stoniškiai-1
Šešuvis-11
Pašaltuonis-94
Milaičiai-103
Bliūdžiai-150
Bebirva-108
Bebirva-110
Gėluva-114
Gėluva-99
Gėluva-115
Graužai-105
Krekenava-7
Svedasai-252
Tverečius-336
S_2jr
S_2lp
S_2vv
S_2mn
S_2nr
$S_2šr$
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S_1br
S_1pp
$S_1šv$
S_1gl
S_1rg
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S_1ss
S_1jr
S_1st+ap
S_2rs
S_1db
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1 2 3 4 5 6 7 8 9 10
11 12 13 14 15 16 17 18 19 20
21 22 23 24 25 26 27 28 29 30
31 32 S_2jr 33 S_2mn 34 S_2pg 35 S_2rs 36 S_1ss 37 S_1rgn 38 S_1rs 39 S_1st+ap 40
S_2lp 41 S_2vv 42 S_2vn 43 S_2mt 44 S_2db 45 S_2nr 46 $S_2šr$ 47 $S_{1-2}nv$ 48 S_1gl 49 S_1rg 50
S_1jr 51 S_1db 52 S_1br 53 S_1pp 54 S_1vr 55 $S_1jč$ 56 $S_1šv$ 57 58 59 60

Fig. 1. Brachiopod communities distribution. 1 – Microsphaeridiorhynchus nucula community; 2 – Isorthis ovalis-Protochonetes piltenensis community; 3 – Isorthis ovalis community; 4 – Dayia-Isorthis community; 5 – Dayia community; 6 – Lissatrypa obovata community; 7 – Jonsea grayi community; 8 – pelagic community; 9 – "Clorinda" sp.; 10 – Lissatrypa obovata-Plagiorhyncha depressa comminity; 11 – Lissatrypa obovata-Eoplectodonta duvalii-Plagiorhyncha depressa comminity; 12 – Lissatrypa obovata-Isorthis clivosa comminity; 13 – Lissatrypa obovata-Protochonetes minimus comminity; 14 – Isorthis amplificata community; 15 – Dicoelosia-Skenidioides community; 16 – Atrypa reticularis community; 17 – Kirkidium knightii community; 18 – Sphaerirhynchia wilsoni community; 19 – Atrypoidea community; 20 – Atrypa reticularis-Leptaena sp. community; 21 – Rhynchotreta cuneata- Atrypa reticularis- Isorthis amplificata community; 22 – Pentamerus gothlandicus community; 23 – Skenidioides lewisii-Isorthis amplificata community; 24 – Dicoelosia biloba-Skenidioides lewisii community; 25 – Dicoelosia biloba-Ptychopleurella lamellosa comminity; 26 – Dicoelosia paralata-Skenidioides lewisii- Lissatrypa obovata community; 27 – Lissatrypa obovata-Pentlandella tenuistriata- Dicoelosia paralata community; 28 – Gotatrypa hedei community; 29 – Lissatrypa obovata-Isorthis amplificata comminity; 30 – Pseudoprotathyris infantilis community; 31 – Pseudoprotathyris infantilis-Morinorhynchus orbignyi community; 32 – facies without brachiopods; 33 – Jūra Formation; 34 – Minija Formation; 35 – Pagėgiai Formation; 36 – Rusnė Formation; 37 – Siesartis Formation; 38 – Ragainė Formation; 39 – Rasytė Formation; 40 – Stačiūnai and Apaščia Formations; 41 – Lapės Formation; 42 – Vievis Formation; 43 – Ventspils Formation; 44- Mituva Formation; 45 – Dubysa Formation; 46 – Neris Formation; 47 – Širvinta Formation; 48 – Nevėžis Formation; 49 – Gėluva Formation; 50 – Riga Formation; 51 – Jūrmala Formation; 52 – Dobele Formation; 53 – Birštonas Formation; 54 – Paprieniai Formation; 55 – Verknė Formation; 56 – Jačionys Formation; 57 – Švenčionys Formation; 58 – bioherms; 59 – formation boundaries; 60 – time boundaries.

lewisii-Isorthis amplificata (BA 4) – ***Dicoelosia biloba-Ptychopleurella lamellosa*** (BA 4) – ***Dicoelosia biloba—Skenidioides lewisii*** (BA 4) – ***Atrypa reticularis*** (BA 3) – ***Sphaerirhynchia wilsoni*** (BA 2–3) communities.

During **Homerian** in ***lundgreni*** time the following lateral sequence was established: **pelagic community** – ***Jonsea grayi*** (BA 4–5) – ***Lissatrypa obovata-Plagiorhyncha depressa*** (BA 4) – ***Lissatrypa obovata-Eoplectodonta duvalii-Plagiorhyncha depressa*** (BA 4) – ***Lissatrypa obovata-Skenidioides lewisii*** (BA 4) – ***Lissatrypa obovata-Protochonetes minimus*** (BA 3–4) – ***Lissatrypa obovata-Isorthis amplificata*** (BA 3–4) – ***Skenidioides lewisii-Isorthis amplificata*** (BA 3–4) – ***Isorthis amplificata*** (BA 3) – ***Rhynchotreta cuneata-Atrypa reticularis-Isorthis amplificata*** (BA 3) – ***Atrypa reticularis-Craniops implicatus*** (BA 3) – ***Atrypa reticularis-Leptaena sp.*** (BA 3) – ***Atrypa reticularis*** (BA 3) – ***Pentamerus gothlandicus.*** (BA 3) – ***Sphaerirhynchia wilsoni*** (BA 2–3) communities.

During **Gorstian** in ***nilssoni*** time **pelagic community** – ***Lissatrypa obovata*** (BA 4–5) – ***Jonsea grayi*** (BA 4–5) – ***Lissatrypa obovata-Isorthis amplificata*** (BA 3–4) – ***Skenidioides lewisii-Isorthis amplificata*** (BA 3–4) – ***Isorthis amplificata*** (BA 3) – ***Atrypa reticularis*** (BA 3) – ***Atrypa reticularis-Leptaena sp.*** (BA 3) – ***Atrypa reticularis-Gypidula galeata*** (BA 3) – ***Pentamerus gothlandicus.*** (BA 3) – ***Sphaerirhynchia wilsoni*** (BA 2-3) communities were recorded

During **Ludfordian** in ***leintwardinensis*** time the following seqence of communities was established: **pelagic community** – ***Lissatrypa obovata*** (BA 4–5) – ***Jonsea grayi*** (BA 4–5) – ***Dayia*** (BA 4) and ***Dayia-Isorthis*** (BA 3–4) – ***Atrypa reticularis*** (BA 3) – ***Kirkidium knightii*** (BA 3) – ***Spirigerina marginalis*** (BA 3) – ***Sphaerirhynchia wilsoni*** (BA 2–3) community.

In ***bohemicus-kozlowskii*** time **pelagic community** disappeared and communityies number significantly decreased with the basin regression. ***Lissatrypa*** dominated communities were totally replaced by ***Dayia*** dominated communities. So the following community sequence was recorded: ***Dayia*** (BA 4) and ***Dayia-Isorthis*** (BA 3–4) – ***Isorthis amplificata*** (BA 3) – ***Atrypa reticularis*** (BA 3).

In ***formosus*** time ***Microsphaeridiorhynchus nucula*** (BA 2) and ***Atrypoidea*** (BA 2) communities developed due to the intensive growth of reefs with a steep forereef slope. It was the reason why no more communities where recorded in between ***Dayia-Isorthis*** (BA 3–4) and ***Microsphaeridiorhynchus nucula*** (BA 2) communities, i.e. their belts of distribution was too narrow to recover them with the existing density of wells.

At the beginning of **Přidoli** in **Varniai** time the reefs gradually decreased in the basin and wide shallow slope developed where shallow water communities appeared. The following sequence was recorded: ***Dayia*** (BA 4) – ***Dayia-Isorthis*** (BA 3–4) – ***Isorthis ovalis*** (BA 3) – ***Isorthis ovalis-Microsphaeridiorhynchus***

nucula (BA 2–3) – ***Microsphaeridiorhynchus nucula*** (BA 2) – ***Pseudoprotathyris infantilis*** (BA 2) – ***Pseudoprotathyris infantilis-Morinorhynchus orbignyi*** (BA 2) – ***Atrypoidea*** (BA 2) communities.

In the mid **Přidolian Girdžiai** time reefs disappeared in the basin which became shallow. It led to the disappearance of ***Atrypoidea, Pseudoprotathyris infantilis, Pseudoprotathyris infantilis-Morinorhynchus orbignyi*** and ***Dayia*** communities. So, the ***Dayia-Isorthis*** (BA 3–4) moved basinwards, onshore – ***Isorthis ovalis*** (BA 3) – ***Isorthis ovalis-Protochonetes piltenensis*** (BA 2–3) – ***Isorthis ovalis-Microsphaeridiorhynchus nucula*** (BA 2–3) – ***Microsphaeridiorhynchus nucula*** (BA 2) communities developed.

At the end of **Přidoli** in **Kelmė–Rietavas** time the basin degradated, ***Dayia-Isorthis*** community disappeared, ***Isorthis ovalis*** (BA 3) community moved westwards and onshore ***Atrypa reticularis*** (BA 3) – ***Isorthis ovalis-Protochonetes piltenensis*** (BA 2–3) – ***Isorthis ovalis-Microsphaeridiorhynchus nucula*** (BA 2–3) – ***Microsphaeridiorhynchus nucula*** (BA 2) communities developed.

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PRINCIPLES OF THE STRATIGRAPHIC DIVISION OF THE BURIED CRYSTALLINE BASEMENT AND UNRESOLVED PROBLEMS OF THE EARLY PRECAMBRIAN STRATIGRAPHY OF BELARUS

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All the geological investigations involving geological mapping, mineral resources studies and forecast mineral assessment of territories are based on a reliable stratigraphic chart. The stratigraphic division of the Lower Precambrian is based on a chronostratigraphic approach [4].

The problem of the stratigraphic division of the Lover Precambrian may be reduced in general to a differentiation of large geological bodies widespread over vast enough territories and corresponding to structural-material complexes (SMC) [1]. Stratigraphic units involved in each specific complex are confined to definite basement structures, show rather similar material composition and thermodynamic parameters of metamorphic and other transformations. Metamorphic members of each SMC are associated as a rule with well-defined ultrametamorphic and magmatic formation complexes. All the above suggests that the formation of each SMC was confined to a large independent stage of the Earth's crust evolution in this region, and SMC themselves can be considered as the main subdivisions of the stratigraphic chart of the region.

The relative position of individual SMC in the regional stratigraphic chart is specified by geochronometric data, or when those latter are absent or few in number by a degree of regional metamorphism correlated with the other regions, where Early Precambrian rocks are widespread.

The extent to which the crystalline basement was studied by drilling in various regions of Belarus becomes an important factor of the basement stratigraphic subdivision, as the sedimentary cover of considerable thickness overlies the basement almost everywhere. In this connection the stratigraphic chart of the Early Precambrian of the studied region is mainly based on data obtained from the most elevated basement parts of Belarus – Belarussian massif and Mikashevichi–Zhitkovichi uplift. Besides, an extremely complicated structure of the basement formations strongly reworked by repeated metamorphic and ultrametamorphic processes should be considered too, as it caused a prevalence of datings that reflect the more recent stages of the basement transformation and an almost entire absence of relict age values [2]. At present about 200 radiological age datings were obtained for the crystalline basement of Belarus using K-Ar method, about 60 datings – using various modifications of U-Pb (the isochron one included) method, and 150 – using kinetic Pb-Pb method by single zircon grains [3]. The data available seem to be inadequate to establish reliable age properties of numerous stratified and intrusive complexes of the basement rocks and to elaborate a consistent chart of the Early Precambrian of the region. By comparison, the stratigraphic chart of the Ukrainian Shield of 1995 was based on 134 reference datings obtained with an isotopic method [5].

The division of the crystalline basement of Belarus was based on a differentiation of well-defined complexes of metamorphic rocks showing similar structural and textural features, approximate peculiarities of the material composition and almost equal degree of metamorphic transformation. A type of association of metamorphic and ultrametamorphic rocks formed under similar physical and chemical and geodynamic conditions and being therefore of the similar age was taken as an additional criterion, when metamorphic subdivisions of the stratigraphic chart were differentiated. So, rocks of the enderbite-charnockite ultrametamorphic complex and the Golenovo complex of anatectic granitoids overlie the Schuchin series granulites and migmatites, granite-gneisses and anatectic granitoids of the Polonka ultrametamorphic complex occur on amphibolite-gneissic strata and blastomilonites of the amphibolite facies. A relationship between metamorphic rock complexes and some magmatic complexes was also taken into consideration though this relation is not so well defined as that between metamorphic and ultrametamorphic complexes.

The relative position of the distinguished subdivisions in the general sequence of Early Precambrian formations was determined from the interpretation of the rock ratios observed in sections, the succession of metamorphic transformations shown in changes of mineral parageneses, not numerous isotopic datings and the correlation with the more extensively investigated regions (stratotype ones included) of the Early Precambrian rock occurrence. The areal distribution of the stratigraphic chart subdivisions was determined from drilling data correlated with characteristics of anomalous magnetic and gravitational fields. When the stratigraphic chart was compiled, a requirement was fulfilled that subdivisions differentiated in the stratigraphic chart of the buried crystalline basement should be easily identified from their visual appearance or using a minimum of additional investigation. This eliminates difficulties that may emerge sometimes when the chart is used for geological explorations.

Magmatic complexes that break metamorphic and ultrametamorphic strata were distinguished by their material composition, formational belonging, association with one or other stratified subdivision and their mutual relations observed in the core of boreholes. A degree of metamorphic transformations of magmatic rocks, as well as data of geochronometric studies were considered too.

Among the major unresolved problems of the crystalline basement stratigraphy in Belarus are as follows: still uncertain age of the Schuchin series and Rudma strata (Early Archean or Early Proterozoic?) and, respectively, the general Early Precambrian evolution of the crystalline basement in the region; similar age of various strata (Ozery, Peretoka and Yurovichi) of the amphibolite-gneissic complex and Okolovo series, which became evident from the kinetic method data, but has not yet been certainly proved. Additional investigations are needed to divide the Okolovo series.

There is a number of unresolved problems with magmatic complexes too. So, the independence of some granitoid complexes (Vygonovo, Bobovnia ones and others) still remains to be proved; the situation with the Berezovka complex differentiation and with the general subdivision of the basic rock complexes seems to be complicated enough. The belonging of such a magmatic almost not metamorphosed complex as the Kamenets gabbro-dolerite complex to the Late Proterozoic or even to the Phanerozoic must not be ruled out.

In conclusion it should be noted that in course of further investigations special attention should be given to searching for relict datings and to obtaining reference ages with the most foolproof methods. Further research needs to be performed with the kinetic Pb-Pb method which is less expensive and, respectively, more extensively used. Standard lithologic and stratigraphic methods used for subdivision and correlation of Early Precambrian formations should be of great importance as before. In this respect geochemical studies seem to be very promising as help in reconstructing the primary nature of metamorphic formations and, therefore, in differentiating rocks that compose outwardly resembling metamorphic strata, but formed under different deposition conditions.

Neverthless, the suggested stratigraphic chart of the Lower Precambrian of Belarus may be used today in studies of geology and ore content of the crystalline basement, in compiling legends of geological maps of the new generation. The chart application in industry will increase the efficiency of geological surveying and exploration works.

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STRATIGRAPHY AND MAGMATISM OF THE CRYSTALLINE BASEMENT OF BELARUS

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The crystalline basement of Belarus is one of the most extensively studied areas of the Russian Plate basement even though it is overlain by a rather thick sedimentary cover within the most part of the territory. It is stripped by more than 4,500 boreholes very unevenly distributed over the area. About a half of them are located within a small (about 240 sq. km) Mikashevichi–Zhitkovichi Horst, more than 1,500 boreholes were drilled in the Belarussian Massif, almost 400 – in the territory of the Pripyat Trough.

The Stratigraphic Chart of the Early Precambrian of Belarus now in force was compiled in the early eighties with a participation of all the geological organizations of Belarus, was adopted at the stratigraphic meeting in Gomel (Belarus) in 1981 and approved by the USSR Interdepartmental Stratigraphic Committee in 1982 [6]. The Chart involves the ideas about the basement stratigraphic division mostly used and accepted by the majority of geologists of Belarus. The use of the Chart permitted a unification of the basement mapping methods employed by various organizations and promoted a correlation of Precambrian subdivisions of Belarus and the other regions. At the same time, the Chart has several disadvantages. The most essential of them is the absence of a stratigraphic subdivision corresponding to amphibolite-gneissic strata metamorphozed under the amphibolite facies conditions (metamorphism of the amphibolite facies), that are widespread in the central and western areas of Belarus. In addition, the independency of some intrusive complexes was poorly substantiated.

A considerable body of new data on the stratigraphy of the crystalline basement of Belarus obtained within last twenty years at the Laboratory of the Early Precambrian of the Institute of Geological Sciences of the National Academy of Sciences of Belarus made for a differentiation of several new stratigraphic divisions and magmatic complexes absent in the previous Chart and provided an additional substantiation of the earlier known divisions not included in the Chart of 1982 for one or another reason.

The working stratigraphic chart of the Archean and Early Proterozoic of Belarus approved by the USSR Stratigraphic Committee [6] with some modifications and additions was used as a basis of the suggested version of the stratigraphic chart of the Early Precambrian of Belarus. The age of subdivisions differentiated was indicated to erathem. The following age limits were accepted: 3,200±50 mln. yrs. between the Lower and Upper Proterozoic.

The final stratigraphic chart assumed that the crystalline basement of Belarus is subdivided into two regions: western and central parts of Belarus and eastern part of Belarus.

The suggested version of the stratigraphic chart of the Archean and Early Proterozoic of Belarus (see Table) includes the following stratified subdivisions: the granulite Schuchin and Kulazhin series and Rudma strata [3] that are supposed to be of Archean (early or late) age, three amphibolite-gneissic Ozery [1] and recently propose Peretoka and Yurovichi [2] strata, that obviously are of the similar Late Archean or Early Proterozoic age; Early Proterozoic schistose- amphibolite-gneissic Okolovo and essentially schistose Zhitkovichi series [6]. The Spushanka suite which formations were proved to be of metasomatic nature and the Yachenka strata which was placed earlier in the Okolovo series top and which rocks are also of metasomatic nature were excluded from the chart. The Dumichi suite was placed in the Okolovo series bottom [7]. A possibility that the Okolovo series formations (or a part of them) may appear of the same age as the amphibolite-gneissic strata must not be ruled out.

Various age subdivisions of the stratigraphic chart differ from each other firstly in a degree of metamorphism and usually in the petrographic and chemical composition. When two or more divisions are of approximately the same age (Schuchin and Kulazhin series, Ozery, Peretoka and Yurovichi strata, Okolovo and Zhitkovichi series), they differ in both the material composition, and confinement to various structural zones, that were clearly recognized in the crystalline basement of Belarus. So, the Schuchin series is widespread within the Belarussian–Baltic granulite belt, the Kulazhin series is confined to the Bragin granulite massif, the Ozery, Peretoka and Yurovichi strata occur in the Inchukaln folded zone, Smolevichi–Drogichin suture zone and Osnitsk–Mikashevichi volcanic-plutonic belt, respectively.

Table. Regional Stratigraphic Chart of the Archean and Early Proterozoic of Belarus (2000).

Subdivisions of the General Stratigraphic Chart		Western and Central regions		Eastern region	
		Metamorphic formations	**Magmatic and ultrametamorphic formations**	**Metamorphic formations**	**Magmatic and ultrametamorphic formations**
PR	**$PR_{1\text{-}2}$**		Kamenets and Mosty magmatic complexes	**Bobruisk series** Myshkovichi suite Luchki suite	Nichiporovka and Zagorbashie magmatic complexes
	PR_1	**Okolovo series** Shashki suite Gumenovschina suite	Peski, Kossovo, Zhukhovichi, Vygonovo, Bobovnia and Rusinovka magmatic complexes	Belevo suite **Zhitkovichi series** Kozhanovichi suite Ludenevichi suite	Berezina, Nagornaia, Zhitkovichi, Mikashevichi and Volkhva magmatic complexes
AR_2-PR_1		**Ozery and Peretoki strata**	Migmatite-granite-gneissic and blastomilonite ultrametamorphic complexes and Korelichi magmatic complex	**Yurovichi strata**	Migmatite-granite-gneissic ultrametamorphic and Anisimovka magmatic complexes
AR_1(?)		**Rudma strata** **Schuchin series** Ditvin strata Zaborie strata	Enderbite-charnokite ultrametamorphic complex, Osmolovo, Golenovo and Berezovka magmatic complexes	**Kulazhin series**	

A list of magmatic rock complexes that were absent in the chart of 1982 includes the Osmolovo and Kossovo complexes of hypersthene granitoids revealed within the Belarussian–Baltic granulite belt, the granitoid Bobovnia and Vygonovo complexes determined earlier (Smolevichi–Drogichin suture zone), the gabbroid Anisimovka complex (Osnitsk–Mikashevichi volcanic-plutonic belt) and the Berezina and Zagorbashie complexes recently distinguished also within the volcanic-plutonic belt by N. V. Aksamentova [3].

The suggested version of the stratigraphic chart of the Early Precambrian of Belarus was considered and discussed at the III All-Russian conference "General issues of Precambrian subdivisions" in Apatity (June 13–17, 2000) [5].

A great disadvantage of the new chart is its poor substantiation by geochronometric evidences. It should be also considered that the majority of the radiometric datings now available show most likely the latest transformations of metamorphic rock complexes rather than the time of their formation [4].

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USE OF K-BENTONITE BEDS AS TIME-PLANES IN SEQUENCE STRATIGRAPHIC ANALYSIS OF CARADOC (UPPER ORDOVICIAN) CARBONATE SEDIMENTS IN ESTONIA

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The Baltoscandian Palaeozoic epicontinental palaeobasin developed as carbonate platform on the Baltica continent from Middle Ordovician until the end of Silurian. During this time drastic climatic changes took place when the Baltica Craton drifted from the southern high latitudes to the tropical realm (Torsvik et al. 1992). In the development of the Baltoscandian Middle Ordovician–Silurian basin, five stages were differentiated (Nestor 1990; Nestor and Einasto 1997).

In the Middle and early Late Ordovician (Arenig–Caradoc, transgression and unification stages by Nestor 1990) basin developed in temperate climate conditions. Sedimentation rates were slow and constant in this sedimentary starving basin due to the proximity stable Baltic Shield. The Caradoc age was period of relative tectonic and eustatic stillstand on the East-European Platform, which terminated as the short transgressive phase at the beginning of the Haljala stage, longer stillstand in the middle Caradoc and possible shallowing at the end of the Keila age (Nestor and Einasto 1997). The regression and new transgression at the end of the Keila stage is described as type 1 sequence boundary (Ainsaar and Meidla 2001).

There are few attempts made to describe development of Baltoscandian basin in sequence stratigraphic categories (Dronov and Holmer 1999; Ainsaar and Meidla 2001). In present study the early Caradoc basin infilling history in North Estonia is analysed in detail. It became obvious that regional stages in Ordovician are too rough units to describe sequence- and system tract-scale trends in sedimentation. We need more detail time-planes to divide the sedimentary sequences into time units and see the dynamics of sedimentation. In Caradoc good opportunity for detail stratigraphy gives the presence of numerous K-bentonite beds in the sections.

During the Haljala and Keila stages numerous volcanic ash intercalations formed giving evidence of growing volcanic activity in the adjacent Iapetus Ocean, perhaps connected with its transition from the opening phase to the closing state (Nestor and Einasto 1997). Numerous studies and geological mappings have been shown the correlation of distinctive K-bentonite beds. Erika Jürgenson (1958) separated five more important K-bentonite beds in the Caradoc (Haljala and Keila stages) sections in Estonia and denominates them as "a", "b", "c", "d" and "e" beds. Bergström and others (1995) applied geographical names to prominent K-bentonite beds as follows: "a" and "b"-bed of Jürgenson (1958) – Grefsen, "c"-bed – Sinsen, "d"-bed ("main bentonite bed") – Kinnekulle, and "e"-bed – Grimstorp. Thickness of K-bentonite beds various from 2 to 30 cm in Estonia. All of the named K-bentonite beds or bed complexes can be recognized throughout Baltoscandia, except the eastern part of it (NW Russia), and the correlation of some of them with North American beds is proposed (Huff et al. 1992). The Grefsen and Kinnekulle beds have used in formal stratigraphical schemes as boundaries of regional stages or substages.

Leslie and Bergström (1997) have been used Caradoc K-bentonite beds (Deicke and Millbrig beds) in eastern North America as a stratigraphic framework, in which litofacies distributions and changes through time, and net rock accumulation rates can be examined. Their study showed that K-bentonite beds are useful as time-planes for such studies because of their large-scale spreading throughout different facies zones.

In present study we show the sedimentation dynamics in upper-middle ramp situation in Estonia during Caradoc epoch (Haljala and Keila ages) using K-bentonites as time-planes. The Haljala and Keila Stages comprises argillaceous bedded to nodular limestone (wackestone and packstone) with marl intercalations. (Põlma et al. 1988; Ainsaar and Meidla 2001). This sedimentary sequence of the upper Haljala and lower Keila stages has been divided into four units using bentonite layers as boundaries for these beds, thickness maps of which are compared. Data of more than one hundred drillcores has been used for the analyses. The thickness of separated units varies between 0.4 and 10 m.

Generally, in all of these beds the thickness is bigger in northern part of the study area (landward) and thinner in south (basinward). Analysis of the maps allows following the maximum thickness area indicating the deposition axis on the maps. The maximum thickness belt has a shape of arc, spreading in west-east direction. Each bed thins 2–3 times southward from the deposition axis. In some maps slight thinning in northern (landward) direction can be observed.

In Bed I (between Grefsen and Sinsen K-bentonites; Fig. 1–I) the maximum thickness reach to 7.4 m in NE Estonia. The northward thinning cannot be observed, it probably existed outside nowadays erosional boundary. Thus, the deposition axis was probably north from the outcropping area.

In Bed II (between Sinsen and Kinnekulle K-bentonites; Fig. 1–II) the maximum thickness area is in western part of Estonia (max. 7.6 m). Comparing to the previous unit, the maximum thickness belt and deposition axis has been shifted southward.

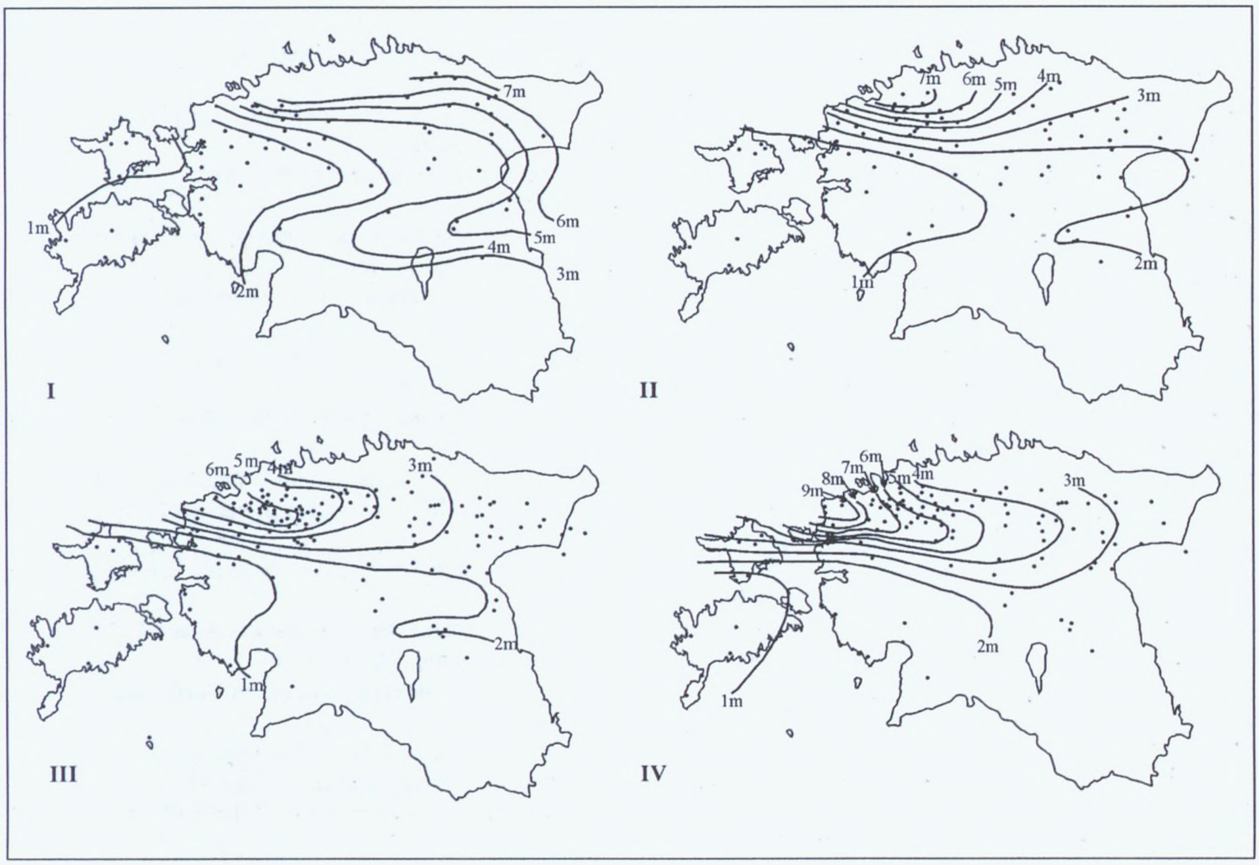

Fig. 1. Thickness maps of the K-bentonite-separated beds of the Haljala and Keila stages. Bed I is between Grefsen and Sinsen K-bentonite layer; bed II – between Sinsen and Kinnekulle layer; bed III – between Kinnekulle and Grimstorp layer; bed IV – between Grimstorp K-bentonite layer and massive limestone bed of the Pääsküla Mb.

In Bed III (between Kinnekulle and Grimstorp K-bentonites; Fig. 1–III) the deposition axis is shifted more south (ca 10–15 km) comparing to the previous bed. Thinning of bed in southern and northern directions can be clearly observed.

In Bed IV (between Grimstorp K-bentonite and well-traced massive limestone bed on the base of the Pääsküla Member; Ainsaar 1993; Fig. 1–IV) continuous southward shift of deposition axis can be seen (approximately 20 km comparing to previous unit). Southward thinning is considerable, reaching up to 3 times comparing with maximum area, northward thinning does not exceed 1.5 times in the area.

This analysis shows constant drift of maximum deposition area (deposition axis) in northern Estonia in southward direction and this is evidence for clear progradational character of sedimentation in this time. According to the sequence stratigraphic interpretation, these units represent the prograding cycle in high stand system tract of the sequence. This sequence, called as Haljala-Keila sequence, starts with hiatal Kukruse/Haljala stage boundary, interpreted as type 1 sequence boundary. The Haljala-Keila sequence is terminated by another type 1 boundary in upper part of the Keila Stage (Ainsaar and Meidla 2001).

Very similar progradation pattern in the Uhaku and Kukruse stages has been demonstrated by Männil, et al. (1986), Bauert (1989) and Saadre and Suuroja (1993a, b) by detail analysis of oil-shale bearing beds. The maximum thickness belts of the beds together with kukersite layers shifted from northeast Estonia towards to central Estonia during the Kukruse time (Männil, et al. 1986; Saadre and Suuroja 1993a, b). This pattern can also be interpreted as progradational high stand system tract of the previous, Uhaku–Kukruse sequence. The progradational pattern can be explained as evidence of gradual fulfilment of the sedimentary space with sediments. It shows, that bioclastic-micritic sedimentation was driven by relative sea level changes creating necessary accommodation space in upper-middle ramp. The low-rate deposited temperate carbonates were able to infill the accommodation space during long-lasted high-stand system tracks, which might happen in situation of lack or minor tectonic movements near the Baltic Shield.

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STRATIGRAPHIC GAPS, AND PROBLEMS OF THE TELYCHIAN (UPPER LLANDOVERY, SILURIAN) BOUNDARIES IN THE NORTHERN EAST BALTIC AREA

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In western Estonia, a subregional unconformity was established between the Raikküla and Adavere regional stages long ago (V. Nestor 1976). By now its lateral extent and maximum stratigraphic range are rather well documented (e.g., H. Nestor 1997). Recently a remarkable gap was recognized also at the contact of the Adavere and Jaani stages in southwestern mainland Estonia (H. Nestor, V. Nestor 2002) but its lateral extent and stratigraphic span are not exactly studied yet. The unconformities approximately correspond to the boundaries of the Telychian and were recognized due to biostratigraphic correlation of drill sections by chitinozoans. The figure shows the position and range of the gaps in the Ikla, Ohesaare and Aizpute-41 drill cores, where it has been possible to integrate zonal successions of chitinozoans and graptolites (Loydell et al. 1998, and in press), and to determine the stratigraphic range of the gaps.

The gap between the Raikküla and Adavere stages is most precisely dated in the Ohesaare core, where it is overlain by the upper part of the *Spirograptus turriculatus* graptolite Zone (*Torquigraptus proteus* Subzone, Loydell et al. 1998) and underlain by the *Ancyrochitina convexa* chitinozoan Zone, which in the Aizpute-41 core coincides with the lower part of the *Demirastrites triangulatus* graptolite Zone (Loydell et al. in press). This shows that the lowermost Telychian and most of the Aeronian, corresponding to the Rumba Formation and to the upper part of the Saarde Formation, are lacking in the Ohesaare section. Southeastwards the gap between the Raikküla and Adavere stages decreases gradually (H. Nestor 1997). However, no criteria for recognition of the temporal analogues of the *Stimulograptus sedgwickii* and *S. halli* graptolite zones have been discovered in the Ikla core in southwestern mainland Estonia. Thus it is possible that unconformity between the Raikküla and Adavere stages extends up to there. The Rumba Formation, which was earlier indirectly correlated with the Aeronian *S. sedgwickii* graptolite Zone (H. Nestor 1972), is now included into the Telychian as in the Ikla section it contains *Stricklandia laevis*, a guide fossil of the Telychian in the stratotype area, and chitinozoans of the *Eisenackitina dolioliformis* Biozone, also characteristic of the Telychian (Verniers et al. 1995; H. Nestor, V. Nestor 2002).

At the southeastern margin of the Baltic basin in eastern Lithuania, the Švenčionys Formation, containing Telychian graptolites at the base, unconformably overlies different uppermost Ordovician (Ashgillian) and lowermost Silurian (Rhuddanian) strata, so that the whole Aeronian is lacking (see Paškevičius 1997). In the Aizpute-41 core in western Latvia there is no biostratigraphic evidence of the topmost Aeronian *S. halli* Zone (Loydell et al. in press), as nowhere in the graptolitic sections of the Baltic Syneclise (see Paškevičius 1997). This allows us to suppose that the unconformity at the base of the Adavere Stage and Telychian could embrace a large territory, if not the whole East Baltic area.

The unconformity between the Adavere and Jaani regional stages, approximately correlatable with the Llandovery–Wenlock boundary (or the upper boundary of the Telychian), was recently established in south-western mainland Estonia and adjacent northern Latvia (H. Nestor, V. Nestor 2002). In the Ikla drill core, red marlstones barren of chitinozoans (Interzone III in the chitinozoan succession), characteristic of the lower part of the Velise Formation, are directly overlain by the Riga Formation of the Jaani Stage (by Interzone IV), showing that the *Angochitina longicollis*, *Conochitina proboscifera*, and *Margachitina margaritana* chitinozoan zones are lacking in this section. The stratigraphic gap is increasing northeastwards. In the Ristiküla and Ipiki drill cores, the entire Velise Formation and the lower part of the Jaani Stage are missing, because the *Pentamerus oblongus* beds of the Rumba Formation are unconformably overlain by calcareous mudstones of the Tõlla Member of the Riga Formation (Jaani Stage) with zonal graptolite *Monograptus riccartonensis*. No exact biostratigraphic data are available yet allowing us to trace the south- and westward lateral extent of the unconformity. However, in the graptolitic sequence of the Aizpute-41 core (western Latvia), Loydell et al. (in press) could not definitely establish upper *Cyrtograptus lapworthi*, C. *insectus* and C. *centrifugus* graptolite zones as well as *Conochitina acuminata* and *Margachitina banwyensis* chitinozoan zones, known from the uppermost Telychian in central Wales (Mullins, Loydell 2001). There is no

SERIES	STAGE	GRAPTOLITE BIOZONES (Loydell 1998)	CHITINOZOAN BIOZONES (V. Nestor 1994)	IKLA	OHESAARE	AIZPUTE	REGIONAL STAGE
WENLOCK	SHEINWOODIAN	*M flexilis*	*C tuba*	Paramaja Mb.	Paramaja Mb.	Riga Fm.	JAANI
		S rigidus (S antennularius)	*C mamilla*	Tõlla Mb.	Tõlla Mb.		
		M riccartonensis	Interzone IV				
		M firmus					
		C murchisoni	*M margaritana*				
?	?	*C centrifugus*	?		?		ADAVERE
LLANDOVERY	TELYCHIAN	*C insectus*					
		C lapworthi	*C proboscifera*		Velise Fm.	Jurmala Fm.	
		O spiralis	*A longicollis*				
		M crenulata	Interzone III	Velise Fm.			
		M griestoniensis					
		S sartorius	?				
		M crispus					
		S turriculatus	*E dolioliformis*				
		S guerichi (R linnaei)		Rumba Fm.			
	AERONIAN	*S halli*	Interzone II	?		?	RAIKKÜLA
		S sedgwickii					
		D convolutus	*C malleus*	Saarde Fm.			
		P leptotheca	*C alargada*			Dobele Fm.	
		N magnus					
		D triangulatus	*A convexa*				
	RH	*C cyphus*	*E electa*		Saarde Fm.	Rm	

biostratigraphic evidence of the presence of these graptolite zones and of the *M. banwyensis* chitinozoan Zone in the Ohesaare core either. However, in the Ohesaare core an about 9 m interval, barren of graptolites occurs between the lower *Cyrtograptus lapworthi* and *C. murchisoni* graptolite zones (Loydell et al. 1998). The presence of the chitinozoan species *Conochitina acuminata* through the whole barren interval (see V. Nestor 1994) shows that a gap, corresponding to the *Cyrtograptus insectus* and, perhaps, to *C. centrifugus* graptolite zones and to the *Margachitina banwyensis* chitinozoan Zone, probably occurs in the Ohesaare section. The former graptolite zone has not been established anywhere in the East Baltic graptolitic sections, therefore a regional unconformity is supposed to exist at the top of the Telychian in this region (Loydell et al. in press).

D. Loydell (1998) established major sea-level lowstands just prior to the end of the Aeronian (*Stimulograptus sedgwickii* to *S. halli* chrons) and Telychian (late *Cyrtograptus lapworthi* to *C. insectus* chrons), which were probably related to the glaciations in South America (Caputo 1998). The subregional to regional unconformities at both boundaries of the Adavere Stage, roughly corresponding to the Telychian, were evidently caused by the above-mentioned glacio-eustatic lowerings of sea-level. The analogous unconformities are established in shallower-water sequences of different continents (e.g. Panuara unconformity at the end of the Aeronian in central New South Wales). It is quite possible that similar gaps occur also in the type sections of the Telychian and Wenlock in the marginal part of the Welsh Basin, where the graptolite succession is incomplete. This complicates determination of the boundaries of these units in other areas.

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BENTONITE IN THE UPPERMOST KUKRUSE STAGE IN NORTH ESTONIA

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In the North Estonian sections located in the stratotype or near the outcroping area of the Kukruse and Haljala stages there is a wellknown and remarkable gap in sedimentation between these stratons (Nõlvak 1972, fig. 4; Männil 1984, fig. 2; Männil, Bauert 1984; Hints *et al.* 1995). In comparison with the stratotype of the Kukruse Stage (Savala core) the southward sections in central Estonia are more complete on account of the beds of the uppermost part of the Kukruse Stage. Among these younger layers there occurs a specific and easily recognizable "blue layer", which is conventionally termed here as a bentonite. However, it has a practical value in local stratigraphy. According to the data available this layer seems to be also the oldest one connected with volcanic ash-falls documented from the East Baltic Ordovician sections.

In 1967 during the detailed bed-by-bed investigations of the acid-resistant microfossils there was found a layer under description from the Kamariku core section at a depth 153.21–153.25 m laying 1.4 m below the palaeontologically proved lower boundary of the Idavere Substage of the Haljala Stage. This rock is insoluble using weak (15%) hydrochloric acid, except the filling limestone in burrows.

DESCRIPTION

By lithological classification all under- (up to 0.5 m) and overlying (up to 2.2 m) beds with the layer of bentonite under discussion belong to the Upper Peetri Member of the Viivikonna Formation (see Männil 1984, fig. 2). In almost all investigated sections the litology of these wackestones occurring just below and above the bentonite is primarily light grey medium-bedded (5–10 cm) argillaceous limestone. Some layers have higher kukersite content that gives them more brownish colour, and a small (up to 3 cm) wavy kukersite seam can be followed between the nodules of the kukersite limestone layer. However, the latter feature is less clearly indicated in western Estonia. Broken trilobite, echinoderm, brachiopod, bryozoa, ostracode debris (skeletal sand) may be locally abundant and is often oriented according to the direction of burrows.

Specific features of the bentonite bed are as follows:

(1) The layer is easily recognizable having a clear blue tint among generally greyish or brownish bedded or partly nodular skeletal calcarenites (containing more than 10% of skeletal particles).

(2) Almost all beds in this interval are thoroughly bioturbated.

(3) In Estonian sections a clear order of individual beds can be followed, except the core 1927 near Slantsy (St. Petersburg Region). Layers are different in their colour and content. Below the bentonite there occurs greenish-grey limestone (up to 10 cm), often with more or less distinct pyritic discontinuity surface on its upper boundary. Above that bentonite layer occurs light bluish-grey limestone (up to 10 cm), which is covered with light-brownish nodular kerogenous limestone (3–8 cm), and higher by seam of argillaceous kukersite (up to 5 cm).

(4) The lower boundary of the bentonite bed ("metabentonite", or "altered volcanic ash" in early literature, see Hints *et al.* 1997, or "K–bentonite", see Bergström *et al.* 1995) is always distinct; the upper, on the contrary, is very transitional according to the concentration of some dark blue silicified material in it. The layer has thickness from 1 cm up to 5 cm, and is also highly bioturbated. Burrows (diameter 1–4 mm) are fulfilled with more argillaceous limestone. In this bed the content of the insoluble residue was evaluated as being 50–80%, containing equally sandy-aleurolitic and pelitic fractions. The aleurolitic fraction (0.01–0.1 mm) consists of amorphous silica (chalcedone and opal) 64–70%, feldspar 1–3%, muscovite 1–10%, pyrite 3–15% and organic matter 3–10%, also minerals apart from pyrite: zircon, garnet, corundum, biotite (data by A. Rahu, 1971). Pelitic fraction (0.001 mm) consists of hydromica 90% and chlorite 10% with admixture of feldspar and quartz (data by laboratory of Tartu University, 1972: from Kamariku, Tudu and Palvere cores). There are absent montmorillonite and mixed-layer illite widely distributed in Estonian /meta/bentonites (see Utsal, Jürgenson 1971).

Nevertheless, its structural and textural peculiarities, and the mineralogical composition (supplements of secondary silica and brownish biotite), also higher content of feldspar in the smaller fractions, give the evidence that this bed can be interpreted having primarily been a volcanic ash, or, at least this ash-fall influenced changes in the sediments on the sea-bottom. In all likelihood much harder framework of this rock is caused by higher content of silica.

TYPE SECTION for this bentonite bed: Kamariku core at a depth 153.21– 153.25 m.

Other investigated core sections (with depths): core 1927 (near Slantsõ) – 114.40–42 m (pers. comm. L. Põlma, 1969); Taga-Roostoja (25A) – 67.35 m; core 6305 (near Tudulinna) – 70.10–12 m; Virunurme (6309) – ? m; Tudu – 79.03–07 m; Rannapungerja – 88.69–72 m; core 502 (near Kamariku) – 105.3 m; Kiltsi – 154.4 m (pers. comm. P. Vingisaar, 1971); Ellavere – 168.4 m; Albu (7049) – 123.90–95 m; Kautla (7034) – 122.76–81 m; Vardja (7020) – 97.30–34 m; Palvere (7019) – 92.41–43 m; Rapla – 157.3 m; core 1149 (near Saue) – 15.26–30 m; core 1098 (near Vasalemma) – 42.15–20 m; Rabivere (307) – 116.0 m (pers. comm. T. Saadre, 1982); Valgu (V-97) – 206.4 m (pers. comm. A. Põldvere, 1997); Rumba – 223.3 m (pers. comm. S. Mägi, 1971); Vihterpalu (357) – 33.6 m; Haapsalu (203) – 140.57–60 m. This described bed is absent in the sections southwards from the line: Are – Äiamaa – Jõgeva – Mustvee, where the beds of the uppermost portion of the Kukruse are again more restricted, or volcanic material did not form any clear individual bed. From the line Keila – Alavere – Tamsalu – Roela – Lipu (see Nõlvak 1972) – Mäetaguse northwards there is a stratigraphical gap in the uppermost Kukruse succession: the thickness decreases rapidly. So, this bentonite can be recognized in a relatively narrow area (tongue), which in west Estonia is about 60 km and in eastern part about 30 km wide, where this bed is distributed and preserved.

MICROFOSSILS AND CORRELATION

In the core samples from the bentonite bed all organic-walled microfossils are very rare compared with limestones below and above. Relatively very rapid sedimentation rate of that mixed sediment had to be taken into account when calculating the abundance of microfossils. Also in the bentonite layer most of the chitinozoan vesicles have been broken into small unidentified pieces by crystallization (probably feldspar), except some small forms from the fillings of burrows, where these specimens were buried later.

Altogether, in addition to lithology very similar chitinozoan associations prove the correlation of the layers under discussion in Estonian sections. The faunal log of chitinozoans from the Fjäcka section in Dalarna (Sweden) gives a hint on a possible correlation of that bentonite bed (Nõlvak *et al.* 1999, fig. 2). In Fjäcka *Conochitina tigrina* Laufeld (1967) is known only from the middle part of Dalby Topoformation, where Jaanusson (1963; 1976, text-figs. 9, 12) described a bentonitic clay layer. This short-ranging chitinozoan species occurs only below and above this layer in Sweden, as in many Estonian sections (Kamariku, Rapla, Kautla, Palvere, Virunurme and Rannapungerja) at a level now regarded by the taxonomic composition of chitinozoans and other acid-resistant microfossils as the uppermost part of the Kukruse Stage.

In Västergötland contemporaneous bentonite is suggested from two other cores: Kullatorp, at a depth 87.35 m and Norra Skagen at a depth 11.54 m based on data in Jaanusson (1964), however, chitinozoans from these sections are not described.

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THE LOWER ORDOVICIAN PILEKIIDAE AND PLIOMERIDAE (TRILOBITA) OF BALTOSCANDIA

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Trilobite zonation of the Lower Ordovician of Baltoscandia elaborated by Tjernvik (1956) was historically based on asaphids. Tjernvik divided the Planilimbata Limestone into two groups: the Hunneberg Group (*Megistaspis armata* and *Megistaspis planilimbata* Zones) and the Billingen Group (*Megalaspides dalecarlicus* and *Megistaspis estonica* Zones). Later he defined (Tjernvik and Johansson 1980) the Transition bed – the zone of *Megistaspis aff. estonica* in the uppermost part of his previous *M. planilimbata* Zone. The new zone turned to be the base of the Billingen Stage (Table), and was characterized by the appearance of the nominate species co-occurring *Pricyclopyge gallica, Raymondaspis brevicauda* and some undescribed new species. The lack of more precise trilobite data the ecophenotypic variation of taxa complicate the correlation of the sections of different parts of the Baltoscandian Palaeobasin.

Representatives of the suborder Cheirurina (Pliomeridae, Pilekiidae) appeared nearly simultaneously with the beginning of carbonate sedimentation in the Baltoscandian Palaeobasin in the Late Tremadoc. First they appeared in the Oslo Region (Ebbestad 1999, Hoel 1999), Västergötland, South Bothnian District and Öland (Tjernvik 1956, Tjernvik and Johansson 1980), later on extended into northern Estonia and Ingria (Balashova 1966, Aru 1980).

The earliest pliomerid *Pliomeroides primigenus* (Angelin, 1854) is quite common in the basal limestone beds of the Bjųrkåsholmen Formation, Tremadoc, and is found in several localities of the Oslo Region together with non-cheirurine trilobites *Apatokephalus serratus, Bienvillia angelini, Ceratopyge acicularis, Shumardia pusilla, Euloma ornatum and Symphysurus angelini* (Ebbestad 1999). It occurs also in Västergötland, Sweden (Tjernvik 1956, p. 115, 177). A very rare pilekiid *Parapilekia speciosa* (Dalman, 1827) is known from a slightly younger bed of the same formation in the Oslo Region (Ebbestad 1999, p. 104–105), and in Öland and South Bothnian District in Sweden (Tjernvik 1956, p. 163).

Hintzeia actinura (Dalman, 1824) is a relatively common pliomerid known from the *M. planilimbata* Zone at Skultorp, Västergötland (Lundqvist *et al.* 1931, p. 54), at Lanna, Närke (Tjernvik 1956, p. 135, Hoel 1999, p. 277) and in South Bothnian District (Tjernvik 1956, p. 163). The species is also known from the hard, grey limestone embedded in dark-grey shale on Andersön Island in Lake Storsjön, Jämtland (Tjernvik 1956, p. 173) and in Dalarna (*ibid.* p. 166) belonging to the *Megistaspis aff. estonica* Zone. Its close relatives are known from the older beds of the *M. armata* Zone in South Bothnian District (*ibid.* p. 163) and from a younger bioherm of the *M. estonica* Zone at Hällabrottet in Närke (Tjernvik and Johansson 1979, p. 188).

Very rare *Pliomerops*? *linnarssoni* (Wiman, 1906) has been described from greenish grey limestone at Skultorp, Västergötland, where it occurs together with *Pliomera mathesii* Angelin, 1954 (Lundqvist *et al.* 1931, p. 54) [= *Hintzeia actinura* (Hoel 1999)], as well as with *M. planilimbata* and *Varvia breviceps* (Wiman 1906, p. 292–293), which mark the topmost part of the *M. planilimbata* Zone. Material closely related to *Pliomerops*? *linnarssoni* occurs in the glauconitic sandstone of the Mäeküla Member in the Lamoshka

Table. Faunal zonation and distribution of the Lower Ordovician trilobites of the families Pilekiidae and Pliomerida in Baltoscandia.

Global Series	British Series	Baltoscandian biozones: Graptolite zones	Baltoscandian biozones: Conodont zones	Baltoscandian biozones: Trilobite zones	Stage	Pilekiidae and Pliomeridae: Western Baltoscandia	Pilekiidae and Pliomeridae: Eastern Baltoscandia
Middle	Arenig	*Didymograptus hirundo*	*Baltoniodus navis* *Baltoniodus triangularis*	*Megistaspis (M.) polyphemus*	Volkhov		
Lower Ordovician	Arenig	*Phyllograptus ang. elongatus*	*Baltoniodus triangularis* / *Oepikodus evae*	*Megistaspis (P.) estonica*	Billingen	*Pliomerops oelandicus* *Hintzeia aff. actinura*	*Evropeites lamanskii* *Pliomerops? cf. linnarssoni*
Lower Ordovician	Arenig	*Phyllograptus densus*	*Oepikodus evae*	*Megalaspides (M.) dalecarlicus*	Billingen	*Evropeites toernquisti*	
Lower Ordovician	Arenig	*Didymograptus balticus*	*Prioniodus elegans*	*Megistaspis (P.) aff. estonica*	Billingen	*Hintzeia cf. actinura*	
Lower Ordovician	Arenig	*Tetragraptus phyllograptoides* (upper, lower)	*Paroistodus proteus*	*Megistaspis (P.) planilimbata*	Hunneberg	*Hintzeia actinura* *Pliomerops? linnarssoni*	
Lower Ordovician	Tremadoc	*Hunnegraptus copiosus* *Araneograptus murrayi*	*Paroistodus proteus*	*Megistaspis (E.) armata*	Hunneberg	*Hintzeia aff. actinura*	
Lower Ordovician	Tremadoc	*Kiaerograptus supernus*	*Paltodus deltifer*	*Apatokephalus serratus*	Varangu	*Parapilekia speciosa* *Pliomeroides primigenus*	

locality, St. Petersburg Region. The main difference between those is in the number of thoracic pleurae. The specimens from sandstone have at least 14 segments while that from limestone has of 13 segments. However, this variation can be explained by ecophenotypic control described by Sheldon (1987) but might also be indicative of speciation. Generally, pliomerids have a quite variable number of thoracic segments contrary to cheirurids. The shallow-water conditions were also favorable for the appearance of different kinds of heterochronic patterns. Thus, it is difficult to decide whether these two specimens are conspecific or not. Also, this genus is spectacular by the decreased number of pygidial spines. It has only four pairs of spines.

Another example of two nearly contemporaneous species inhabiting rather different palaeoenvironments concerns the representatives of the genus *Evropeites* Balashova, 1966. *E. toernquisti* (Holm, 1882) is known from the Billingen greenish gray limestone intercalated with layers of greenish clayey shale in Dalarna, Sweden, and *E. lamanskii* (Schmidt, 1907) from the lowest layers of glauconitic sandstone of the Mäeküla Member in Estonia and St Petersburg Region.

Remarkably, this genus has also decreased number of pygidial spines. It has only two pairs of spines similar to acanthoparyphinines, but otherwise it resembles the pilekiids usually carrying three or four pairs. In conclusion, it appears to be specific for the pliomerids and pilekiids of Baltoscandia to have a decreased number of pygidial segments compared with those from other palaeogeographical regions.

The number of pygidial pleurae has had great weight in the classification of the superfamily Cheirurina at the family and subfamily levels. However, the Baltoscandian pliomerids and pilekiids well show that the number of pleural spines cannot have such weight in familial derivation, as it have had.

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ACRITARCHS FROM THE STRATOTYPES OF THE LITHOSTRATIGRAPHIC UNITS OF THE LASNAMÄGI AND UHAKU REGIONAL STAGES, MIDDLE LITHUANIAN DEPRESSION

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INTRODUCTION

The modern Ordovician stratigraphic scheme of Lithuania is a result of detailed and comprehensive both lithologic and biostratigraphic studies. It demonstrates the unified regional units of the entire Baltic and also the local lithostratigraphic subdivisions recording the specific peculiarities of different facial zones (Laškovas et al., 1993). It is necessary to gain as complete as possible palaeontologic characteristics of the stratotypical sections for a successful recognition and correlation of the local (formations) and regional (regional stages) stratigraphic units on the territory of Lithuania. The acritarch research of the stratotypes has been initiated to clear the taxonomic composition, to ground the regional and local stratigraphic units with a further palaeontologic characteristic, to obtain the characters of the vertical and lateral distribution and to assess the correlational potential of the microphytoplankton. This paper presents the results of the acritarch study of the Lasnamägi and Uhaku regional stages (Middle Ordovician) from the Middle Lithuanian depression. They have been investigated in 21 samples taken from two borehole cores (Figure).

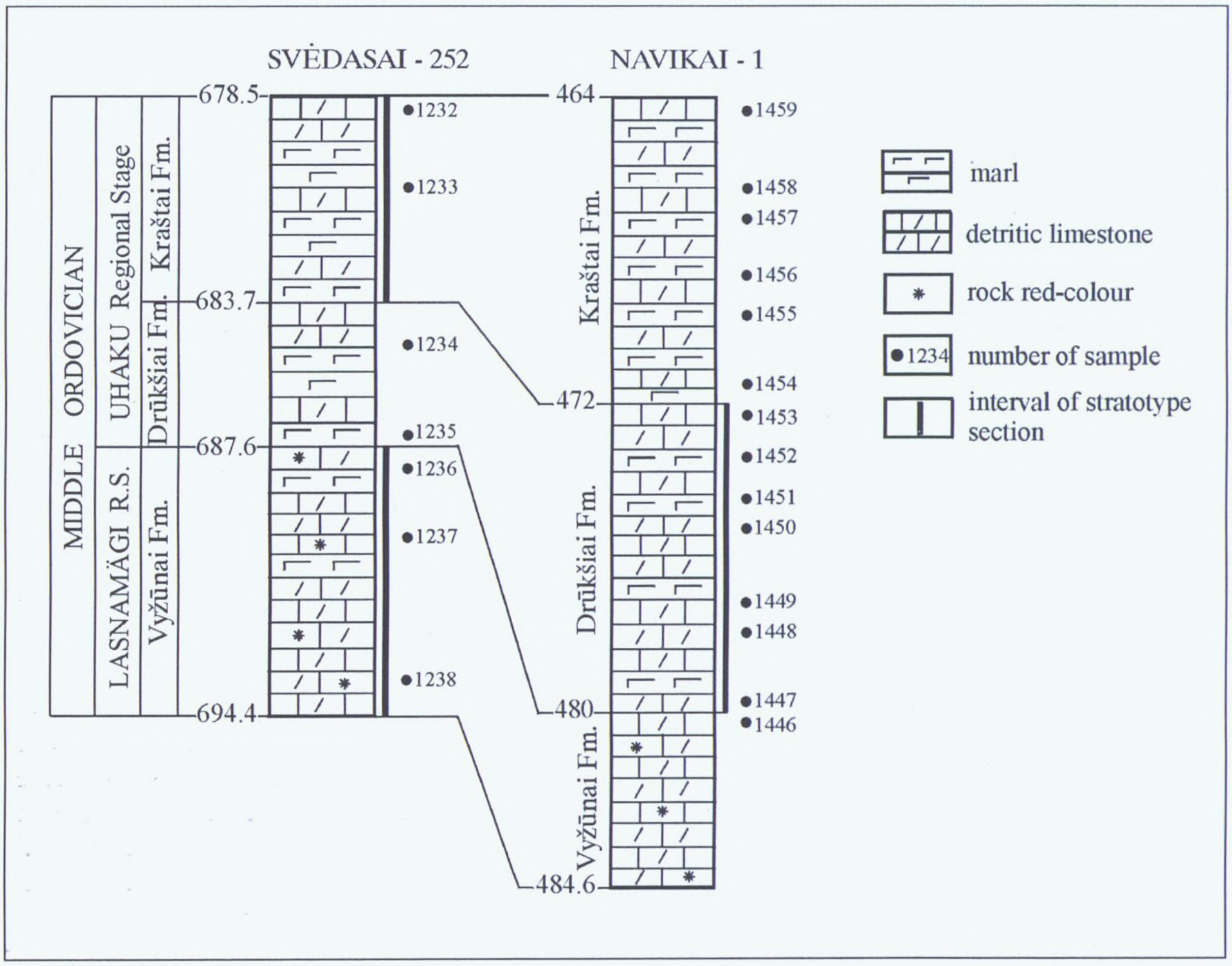

Figure. Lithology, acritarch samples and correlation of the sections studied.

STRATIGRAPHY AND LITHOLOGY

The Middle Lithuanian depression involves central and eastern Lithuania (Lashkov, Paškevičius, 1991). The stratotypes of the most of Middle Ordovician lithostratigraphic units have been arranged in the Svėdasai-252 and Butkūnai-241 boreholes placed in the axial part of the depression (Lashkov et al., 1984). The Vyžūnai Formation was attributed to the Lasnamägi and the Kraštai Fm. – to the Uhaku Regional Stage. The detailed core study of the boreholes from the Drūkšiai area, situated on the eastern slope of the depression, has cleared several lithologic specific characters of the Ordovician profile caused to distinguish a new, Drūkšiai Fm. (Lashkov et al., 1993). The age of the Vyžūnai, Drūkšiai and Kraštai formations and the correlation of sections are based on the palaeontologic (brachiopods, ostracodes, etc. fauna), lithologic and geophysic data (Lashkov et al., 1984; Sidaravičienė, 1996).

The stratotype of the Vyžūnai Fm. is arranged in the Svėdasai-252 borehole at a depth 687.6–694.4 m. The formation is composed of variegated and grey organogenous, detritic limestones with lenses and interbeds of marls. Frequent surfaces of sedimentary gaps are covered by ferrous oxides (Figure).

The Drūkšiai and Kraštai formations are ascribed to the Uhaku Regional Stage. The stratotype of the Drūkšiai Fm. is chosen in the Navikai-1 borehole at a depth 472–480 m. It is composed of grey organogenous, detritic limestones and intercalated grey and greenish grey clayey or dolomitic marls with the wavy surfaces of sedimentary gaps covered by pyrite and phosphate. The stratotype of the Kraštai Fm. is arranged in the Svėdasai-252 borehole (678.5–683.7 m). The volume, if compared with that primary defined (Lashkov et al., 1984), is reduced, leaving the name Kraštai for the upper section part only, containing more marls. The formation is composed of intercalations of grey and green marls and grey organogenous, detritic, unevenly clayey limestones.

RESULTS AND DISCUSSION

The stratotype of the Vyžūnai Fm. has ielded abundant acritarchs in respect with the genera, species and number of specimens. Sixty-five species belonging to 22 genera are identified. The representatives of *Baltisphaeridium* genus are evidently predominant (24 taxa). Another the most common fossil group forms *Pachysphaeridium, Veryhachium* and *Polygonium* genera, are represented by 4–7 species (Table). The remaining genera have 1–3 species. The most of acritarchs of the Vyžūnai Fm. have wide vertical ranges occupying some stages or even series of the Ordovician. Among them, except for the baltisphaerids, *Multiplicisphaeridium alloiteaui, M. irregulare, Micrhystridium nannacanthum, M. stellatum, Veryhachium trispinosum, V. reductum, Pachysphaeridium robustum, Rophaliophora palmata, Leiovalia similis*, etc. are worth mention. The phytoplankton of the Vyžūnai Fm. in the Navikai-1 core is studied in a single sample (Figure). It is taken 0.2 m below the formation top and according to the taxonomic composition of acritarchs is similar to the sample 1236 of the Svėdasai-252 core. There are 31 species belonging to 16 genera identified here. The most stratigraphically valuable acritarchs for the definition of the Lasnamägi Regional Stage and correlation of the rocks are *Pterospermopsis tranvikensis, Micrhystridium nanodigitatum, Excultibrachium concinnum, Arkonia* sp. cf. *A. semigranulata, Polygonium tenuispinosum* and *Orthosphaeridium densiverrucosum.*

In the stratotype of the Drūkšiai Fm., 109 species and 27 genera of acritarchs are identified. The transitional species of *Baltisphaeridium* genus predominate (27 taxa), as in the Vyžūnai Fm. Among the occurring species, *B. verrucatum, B. riegnellii, B. folkeslundianum, B. arboreum* and *B.* sp. C are mentioned. The genus *Veryhachium* is represented by 10 species. Some species such as: *V. trisulcum, V. domasioides, V. reductum* and *V. rhomboidium* are very abundant in specimens among them. *Pachysphaeridium* genus demonstrates high diversity too (9 species). Another the most common group of the genera, including each 4–6 species, forms *Peteinosphaeridium, Polygonium, Multiplicisphaeridium* and *Micrhystridium.* As an exclusive character of the phytoplankton association of the Drūkšiai Fm., is the abundance of *Lophosphaeridium* and *Aremoricanium* genera (5 and 3 taxa respectively) and the first occurrences of *Stelliferidium, Stellechinatum, Leiofusa* and *Strablosphaeridium.* The ranges of two stratigraphic valuable species are worth attention. The first of them, *Pterospermopsis tranvikensis,* is attached to the lower member of the formation, and the second, *Aremoricanium simplex,* is for the first time identified in the rocks of the upper member. Otherwise, *A. simplex* in the Svėdasai-252 is not found, whereas the similar *Aremoricanium* sp. A Playford & Martin, 1984 together with *P. tranvikensis* is defined 0.2 m higher the lower boundary of the Drūkšiai Fm. (sample 1235). The acritarch association of the Drūkšiai Fm. from the Svėdasai-252 borehole is analogous to that of the stratotype. The differences obtained are not essential. In the Svėdasai-252, *Dictyotidium* genus has yielded more numerous specimens, but the lower diversity is characteristic to *Pachysphaeridium, Multiplicisphaeridium* and *Baltisphaeridium*; *Lophosphaeridium* and some species of *Veryhachium* (*V. lairdi, V. cymosum*) are not found. Moreover, some taxa lacking in the stratotype such as: *Stellechinatum celestum, Pirea* sp. A, ?*Tranvikium polygonale* and *Visbysphaera* sp. n. are identified here.

The species of the transitional *Baltisphaeridium* genus (31 taxa) are also predominant in the Kraštai Fm. Among the other genera with the great diversity of species and quantity of specimens, *Veryhachium* (7 species), *Pachysphaeridium* (6 species), *Polygonium, Micrhystridium* (4 species each), *Aremoricanium* and *Ordovicidium* (3 species each) can be mentioned. The remaining genera are represented by 1–2 taxa. *Aremoricanium simplex, Baltisphaeridium? bulbosum* and *Strablosphaeridium biparietalis* make the first occurrences in the Svėdasai-252 core. *Navifusa ancepsipuncta, Pterospermopsis tranvikensis* and the *Stellechinatum* taxa are absent in the Kraštai Fm.; acritarchs of the *Stelliferidium* genus are found with the essentially smaller quantity. The Kraštai Fm. of the eastern slope of the depression (Navikai-1 borehole) contains a more diversified acritarch association represented by the increased quantity of specimens per sample. Together with the specimens of *Baltisphaeridium* genus, the increased quantity belongs to *Ordovicidium, Veryhachium* and *Micrhystridium* genera, whereas in some samples (e.g., 1454) also to

Table. Distribution of selected acritarch taxa in studied sections, formations and regional stages.

Taxa — Regional Stage	Lasnamagi		Uhaku			
Formation	Vyžūnai		Drūkšiai		Kraštai	
Section	Svėdasai	Navikai	Svėdasai	Navikai	Svėdasai	Navikai
Aremoricanium deflandrei	cf		x	x	x	x
Veryhachium trispinosum	x		x	x		x
Veryhachium cymosum	x			x		x
Veryhachium domasioides	x		x	x	x	x
Veryhachium brevitrispinum	x		x	x	x	x
Veryhahium reductum	x	x	x	x	x	x
Pachysphaeridium sp. n. 2	x		?	x	x	x
Pachysphaeridium robustum	x		x	x	x	x
Pachysphaeridium mochtiense	x	x	x	x		x
Pachysphaeridium kjellstroemii	x		x	x	x	x
Pachysphaeridium pachyconcha	x		x	x	x	cf
Ordovicidium nudum	x		x	x	x	
Ordovicidium elegantulum	x		x	x	x	x
Ordovicidium heteromorphicum	x	x	x	x	x	x
Peteinosphaeridium micranthum	x	x		x		x
Peteinosphaeridium eximium ?	x		x	x	x	x
Visbysphaera gotlandica	x			x	x	x
Baltisphaeridium llanvirnianum	x		x	x	x	x
Orthosphaeridium densiverrucosum	x	?	x	x	x	x
Gyalorhethium chondrodes	x		x	x	x	
Polygonium pellicidium	x		x	x	x	x
Polygonium tenuispinosum	x		x	x	x	x
P. polygonale polyacanthum	x		x	x	x	x
Pterospermopsis tranvikensis	x		x	x		
? Tranvikium polygonale	x		x		x	
Baltisphaeridum dasos	x		x	x		x
Dictyotidium sp. cf. *D. oculatum*	x		x	x	x	x
Arkonia sp. cf. *A. semigranulata*	x					x
Excultibrachium concinnum	x			x	x	
Gyalorhethium angustispinosum	x		x	x	x	x
Micrhystridium nanodigitatum	x	x				
Peteinosphaeridium trifurcatum		x	x	x		
Dasydorus cirritus		x	x	x	x	x
Cymatiosphaera spp.		x	x	x	x	
Stelliferidium striatulum			x	x		x
Stelliferidium stelligerum			cf.	x	cf.	x
Aremoricanium rigaudae			x	x	x	x
Veryhachium rhomboidium			x	x		?
Pachysphaeridium sp. n. 1			x	x	x	x
Stellechinatum uncinatum			x	x	?	
Stellechinatum celestum			x			
Aremoricanium sp. A			x			
Aremoricanium simplex				x	x	x
Veryhachium lairdi				x	x	x
Strablosphaeridium biparietalis				x	x	x
Veryhachium trisulcum				x		x
Pachyspharidium bulbosus				x		
Baltisphaeridum folkeslundianum				x		
Leiofusa granulacutis				x		
Veryhachium oklahomense					x	x
Veryhachium minutum					x	
Baltisphaeridium ? bulbosum					x	
Leiofusa fusiformis						x

Peteinosphaeridium. Leiofusa fusiformis, Arkonia sp. cf. *A. semigranulata* and representatives of *Stelliferidium*, which make rare occurrences in other localities, are found in the Navikai-1 core.

Stratigraphically most informative acritarchs of the Uhaku Regional Stage originated in the Middle Lithuanian depression are the species of *Stelliferidium* and *Stellechinatum,* also *Aremoricanium simplex, Baltisphaeridium? bulbosum, Strablosphaeridium biparietalis, Leiofusa fusiformis, L. granulacutis* and *Arkonia* sp. cf. *A. semigranulata.* As one of the important peculiarities of the Uhaku acritarch association is a great diversity and quantity of *Veryhachium* taxa.

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MICROBIAL PALEONTOLOGY AND CATHODOLUMINESCENCE – A TOOL FOR THE INVESTIGATIONS OF THE ORDOVICIAN PHOSPHATE – BEARING SEQUENCE OF THE BALTIC BASIN

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The Lower/Middle Ordovician sedimentary sequence in the extensive area of the East-European Craton reveal significant concentration of authigenic iron and phosphorus minerals. In the Llanvirnian wackestones and packestones enriched with the phosphatic horizons, the phosphorus can be concentrated as a 1. granular phosphorite which contains coated grains including ooids, peloids and phosphatized bioclasts 2. biolaminated phosphatic structures (laminae, hardgrounds, phosphatic matrix). The latter was a result of phosphate precipitation at the bottom stabilized by microbial benthic communities

This stratigraphic interval reveals the features of lithological, taphonomical and sometimes stratigraphical condensation (Podhalańska, 1992, 1995). The latter are of interest because they form distinctive stratigraphic levels and are used as indices of the transgressive – regressive cycles (Krajewski et al., 2000, Marcinowski et al., 1996).

Occurrences of phosphate in the Llanvirnian deposits have been investigated by the author in the Podlasie Depression, Baltic Syneclize, Łeba Elevation (NE Poland) and northern Estonia regions (Podhalańska, 1992, 1995; Podhalańska, Nõlvak, 1995). The lithologic and facial development of the Lower/Middle Ordovician deposits in NE Poland is most closely comparable to that in North Estonia and Lithuania.

The stratigraphy of these beds has been based on conodonts, whose distribution proves taphonomical condensation. The following conodont zones have been distinquished: *Amorphognathus variabilis, Eoplacognathus suecicus* and *Eoplacognathus reclinatus.* Index taxa from all zones and subzones are represented either in correct microstratigraphical order or as mixed fauna, composed of various elements in one stratigraphical level (Podhalańska, 1992, 1995).

The use of new methods of investigations – cathodoluminescence and microbial paleontology, together with electron microscopy and X-ray analysis of element distribution to investigate phosphatic structures, allows to determine their microstructure and to reconstruct depositional and early diagenetic processes which were responsible for the origin of the Ordovician phosphate-bearing sequence of the Baltic Basin.

SEM analyses show a great abundance and diversity of microbial filaments of different shape, mode of branching as well as spheroidal bodies interpreted as endo- and epibenthic fossil community in the layers enriched in phosphatic peloids, ooids and bioclasts and within the laminae.

Fossil microbial communities occur mainly in skeletal grains, in the rims of the coated grains and near the hardgrounds. A close association between microbial activity and the formation of apatite and the presence of bacterial communities in the phosphatic rims and laminae suggests microbial participation in the build up of these structures. Due to methods of bacterial paleontology, on the base of shape, dimension and mode of branching – fungal, cyanobacterial and green algae sheats have been distinquished (Podhalańska, 1998, 2001).

Biochemically precipitated phosphate around microbial bodies, during their life or in a consequence of the decay of the organic matter, has improved the preservation of the microbial structures as the fossil records.

Microbial communities, playing an important role in the Ordovician of the Baltic Basin have a high palaeoecological potential and can be used to identify and to interpret palaeoenvironmental conditions especially at the sediment – water interface.

The abundance of microbial structures indicates nondeposition, and the presence of autotrophic forms can suggest photic conditions near the bottom. Their activity during the life and bacterial decay of organic matter influenced:

- changes of chemical and biochemical factors in microenvironment (eutrophization, concentration and precipitation of apatite, oxygen deficiency),
- stabilization of bottom sediment and forming of the phosphatic hardgrounds
- occurrence of the phosphatic grains and phosphatic laminae.

Long before the numerous sedimentologic, petrographic and microstructural investigation of the phosphogenic system have revealed the role of microbial communities in phosphorite formation. As an example (Lucas, Prévōt, 1985; Krajewski et al., 1994, 2000; Southgate, 1986).

The first identification of globular, bacteria-like phosphatic microstructures in the analysed samples may confirm the role of microbial processes in phosphogenesis.

The evidenced process of the crystallisation of hexagonal apatite crystals on primary amorphic globular forms suggests not only that phosphate gel was precipitated prior to apatite crystallisation but also suggests the primary nature of phospatization.

Cathodoluminescence (CL) analyses were performed for the Llanvirnian phosphate – rich samples. This method has proved to be a very useful tool in the study of the phospate – bearing sequence but its application is much less known for the investigators than the examination under the scanning electrone microscope (SEM) and microprobe chemical analyses of phosphates (EDS). This method enables identification of different kinds and generation of minerals. Cathodoluminescence of apatite depends on the contents of the rare earth elements (REE) as well as Sr and Mn which substitute for Ca, P and F and effectively induce luminescence (Marshall, 1988).

The study was conducted to present a distribution and different forms of the phosphatic occurrences in the Ordovician of the East European Craton. This method enables identification of different kinds and generation of minerals. Previously CL was used to Cambrian phosphate investigation in NE Poland (Sikorska, 1998).

Cathodoluminescence study proved the variable intensity of the phosphatisation process and multiple of phosphatic primary coatings. CL spectra for apatite observed in the investigated samples show strong to medium blue luminescence. Blue CL is most frequently encountered in primary apatite from carbonatites. The color of CL images may vary from blue to violet, depending on the distribution of the REE activators, sensitive to geochemical changes in the crystallizing solutions. Also the phosphate distribution in the mineralized microbial laminae on the sea floor as well as zones of the primary phosphatic matrix formation and its distribution were obtained due to CL images.

The results obtained proved the usefulness of the applied methods to the investigation of the distribution, variability and genesis of the Ordovician phosphogenic sequence in the Baltic basin.

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UPPER CRETACEOUS RHYTHMOSTRATONS AND FORAMINIFERAL ZONES OF WESTERN SIBERIA

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As it has been previously stated, West-Siberian Cenomanian/Santonian assemblages, composed mostly by quartz-siliceous agglutinated foraminifera, are comparable with coeval assemblages (of about the same taxonomic composition) from Canada and Northern Alaska assigned to the Arctic paleobiogeographical realm. These foraminiferal assemblages are not correlatable with those from the adjacent regions assigned to the Boreal-Atlantic realm in terms of these organisms. The Late Campanian/Maastrichtian West-Siberian foraminiferal assemblages can be assigned to the Boreal-Atlantic realm by the similarity of their mostly calcareous secreted forms to coeval ones from the north of Middle Asia and Northern Europe.

The broad correlation of both Cenomanian/Santonian and Campanian/Maastrichtian foraminiferal assemblages to coeval taxa from within different realms (Arctic and Boreal-Atlantic) has been performed previously (Podobina, 1978, 1984). However, the application of only a paleobiogeographical method has proved to be inadequate. Within the littoral and shallow facies of West-Siberian province assigned to the Arctic realm for the Cenomanian/Santonian, there are assemblages of calcareous secreted foraminifera besides widespread quartz-siliceous agglutinated forms. That is why along with the paleobiogeographical zonation method, for the purpose of stratigraphy, the paleogeographical method has been applied with which zoning of this province has been performed with regard to the distribution of deep-dwelling and coastal foraminiferal assemblages. They diverge by the test wall chemical composition and, hence, by the taxonomy. For example, the Slavgorodskian horizon enclosing these assemblages has been dated to Santonian on the grounds of combined finds of quartz-siliceous agglutinated forms together with calcareous secreted foraminifera, characteristic for this age, (*Cibicidoides eriksdalensis* Brotzen, *Eponides concinuus* Brotzen *planus* Vassilenko) in littoral marine facies.

Within the most area of West-Siberian basin (a relatively deep-water central part), in the Slavgorodskian horizon, the assemblages of quartz-siliceous agglutinated foraminifera are distributed: those with *Ammobaculites dignus*, *Pseudoclavulina hastata admota* (Early Santonian) and *Cribrostomoides cretaceus exploratus*, *Ammomarginulina crispa* (Late Santonian). They correspond to the assemblages with *Cibicidoides eriksdalensis* in the east (Podobina, 1975, 1989) and to the *Discorbis* assemblage (the upper one with minute anomalinids) in the west (along the east slope of the Urals; Kipriyanova, 1977) of marginal littoral districts of the basin; these assemblages including calcareous secreted forms, but there are also combined assemblages.

As a whole, the Santonian assemblages from littoral facies of the Slavgorodskian and other horizons have developed in the more shallow warmer part of the basin; so they possess calcareous secreted tests and are similar in their taxonomy to coeval forms from Europe and other provinces (Sokolov, 1991).

Besides the conventional biostratigraphical, paleogeographical and paleobiogeographical methods for determining the stratigraphical position of West-Siberian Late Cretaceous foraminiferal assemblages, of great importance are approaches considering the transgressive-regressive cycles interrelated with the tectonic movement. Alterations observed in West-Siberian foraminiferal assemblages throughout the Upper Cretaceous section are connected with lithofacies. These latter are considered by Podobina V. M. as separate rythmostratons. A rhythmical character in the occurrence of foraminiferal assemblages throughout the Upper Cretaceous of the central part of Western Siberia has previously been investigated (Podobina, 1989). On the base of peculiarities in the averaged quantitative distribution of foraminifera, the generalized faunal curve (GFC) has been constructed which corresponds, as mentioned above, to the transgressive-regressive cycles of the development of the basin. In this paper, the additional interpretation of GFC is presented: there are three distinct rhythms standing out on GFC and corresponding to such great rhythmostratons as the Kuznetsovskian–Sedelnikovskian, Slavgorodskian and Gankinskian horizons. Every rhythm associated with certain rhythmostratons is separated in GFC by boundaries between two greatest curve knees corresponding to maximal transgressions (Table).

The GFC subdivisions corresponding to one or two horizons are designated by the authors as rhythmothems. American investigators gave the other name to analogous subdivisions: the marine cyclothems (Caldwell et al., 1993).

The foraminiferal characterization throughout the Upper Cretaceous section reveals changes in their taxa on the level of orders and families which are, as a rule, similar in their taxonomy within appropriate rhythmothems. Horizons or their portions subordinate to the rhythmothems are somewhat diverse in lithology and characterized by a definite generic composition. In the hierarchy of rhythmostratigraphical subdivisions they may be equated to such rhythmostratons as rhythmotherms (the designation has been proposed by the authors and is derived from the Latin *termus* – a cut-off portion of a branch). The local biostratigraphical (foraminiferal) zones distinguished by their species assemblages have been named rhythmolithes which are subordinate to rhythmotherms and these latter to rhythmothems. The accompanying table combines the scheme of the Upper Cretaceous foraminiferal zonal stratigraphy and the GFC diagram based on the relative quantitative distribution of foraminifera. The authors have established three transgressive (T_1–T_3) and three regressive (R_1–R_3) cycles of the basin.

Hence, the approach of the rhythmicity in the distribution of foraminifera involves the isolation of successive rhythmostratons throughout the Upper Cretaceous section and is based on peculiarities of transgressive-regressive cycles of the basin, which are interrelated, as it has been mentioned, with the tectonic regime of the territory.

Recently much attention has been given to the establishment of marine cyclothems (in our interpretation "rhythmothems") on the base of foraminiferal biostratigraphy (Caldwell et al., 1993). American investigators have come to the conclusion that foraminiferal assemblages directly prove eustatic variations in the World ocean's level also associated with the tectonic movement. These variations imprint themselves in rock series corresponding to those cycles. Foraminiferal assemblages also reflect variations in climatic and oceanographic conditions within a basin, including relative displacements of Boreal and Tethyc water mass as well as anoxic events of the basin. The foraminiferal taxonomy reflecting their habitat enables not only zonation sections, but also allows to apply these forms for the detailed biostratigraphical correlation throughout great intervals.

Two major stages of development have been defined by American investigators in the development of the Western Interior Late Cretaceous Basin (Sea way): Cenomanian/Santonian and Campanian/Maastrichtian (Kauffman, 1984); the same being also true for Western Siberia (Podobina, 1989, 2000). The earlier stage is coincident with two transgressive-regressive marine cycles reaching maximal transgressions in the beginning of Turonian and in the end of Santonian. Correlatable with the cycles are some portions of the Greenhorn and Niobrara formations of the Western Interior basin as well as the bottom layers of the Kuznetsovskian and Slavgorodskian horizons of West-Siberian basin.

Three transgressive-regressive cycles have been reported by American investigators for the Campanian/Maastrichtian of the Western Interior Late Cretaceous Basin. The considerable regression being revealed in the Early Campanian and the maximal transgression in the Late Campanian to Early Maastrichtian. The Campanian/Maastrichtian Bearpaw formation as well as West-Siberian Gankinskian horizon correspond to those three cycles. It should be noted that the Western Interior basin, unlike West-Siberian one, had an unobstructed connection with the Tethys and was directly influenced by it. That is why, even in the northern part of Canada (the province of Alberta), foraminiferal assemblages comprise a considerable amount of calcareous secreted forms in addition to quartz-siliceous agglutinated ones.

Table. Rhythmostratigraphical scheme of the Upper Cretaceous in Western Siberia combined with rhythms of quantitative distribution of foraminifera.

System	Series	Stage	Substage	Horizon	Rhythmothem	Rhythmoterm	GFC, %	Rhythmolith	Zone
CRETACEOUS	UPPER	Maastrichtian	upper	Gankinskian	3	2	40 R$_3$	2	*Spiroplectammina kasanzevi, Bulimina rosenkratzi*
			lower			1		1	*Spiroplectammina variabilis, Gaudryina rugosa spinulosa*
		Campanian	upper				T$_3$	2	*Cibicidoides primus*
			lower	Slavgorodskian	2	2	R$_2$	1	*Bathysiphon vitta, Recurvoides magnificus*
		Santonian	upper			1		2	*Cribrostomoides cretaceus exploratus, Ammomarginulina crispa*
			lower				T$_2$	1	*Ammobaculites dignus, Pseudoclavulina hastata admota*
		Coniacian	upper	Sedelnikovskian	1	2	R$_1$	2	*Dentalina tineaformis, Cibicides sandidgei*
			lower					1	*Haplophragmoides chapmani, Ammoscalaria antis*
		Turonian	upper	Kuznetsovskian		1		2	*Pseudoclavulina hastata*
			lower				T$_1$	1	*Gaudryinopsis angustus*
		Cenomanian	upper	Uvatskian				2	*Trochammina wetteri, Trochammina subbotinae*
								1	*Verneuilinoides kansasensis*

Legend

GFC – Generalized Faunal Curve derived from the averaged quantitative distribution of foraminifera; T$_1$–T$_3$ – transgressive cycles; R$_1$–R$_3$ – regressive cycles; - - - - - – borderline between rhythmothems; ⋀⋀⋀⋀⋀ – borderline between rhythmoterms; – ·· – ·· – ·· – – borderline between rhythmoliths

The table illustrates the rhythmostratigraphical scheme combined with the occurrence of foraminiferal assemblages within the Upper Cretaceous of West-Siberian basin. As it is demonstrated in the scheme, West-Siberian marine rhythmothems corresponding to American cyclothems are as follows: the Kuznetsovskian and Sedelnikovskian horizons (the 1st lower rhythmothem), the Slavgorodskian one (the 2nd middle rhythmothem) and the Gankinskian one (the 3rd top rhythmothem). Two rhythmotherms are subordinate to each rhythmothem: the bottom one corresponding to the transgressive cycle of the basin and the top one conforming to the regressive cycle. A rhythmotherm is subdivided into two rhythmoliths coinciding with the layers characterized by adequate foraminiferal assemblages and distinguished as biostratigraphical zones.

The Kuznetsovskian bottom layers (the Early Turonian zone *Gaudryinopsis angustus,* i.e. the 1st rhythmolith) correspond to the maximal transgression; its top layers (the Late Turonian zone *Pseudoclavulina hastata,* i. e. the 2nd rhythmolith) are appropriate to declining transgression. The Sedelnikovskian horizon is dated to the Coniacian and corresponds to the decline of transgression. Its bottom portion is the Early Coniacian zone *Ammoscalaria antis*, *Haplophragmium chapmani* (the 1st rhythmolith); its upper part is the Late Coniacian *Dentalina tineaformis, Cibicides sandidgei* zone (the 2nd rhythmolith). For the Santonian of Western Siberia, the maximal transgression is represented by the Slavgorodskian bottom (the Early Santonian zone *Ammobaculites dignus, Pseudoclavulina hastata admota*: the 1st rhythmolith), and the decline in transgression is reflected by the middle part of the Slavgorodskian horizon (the Late Santonian zone *Cribrostomoides cretaceus exploratus, Ammomarginulina crispa:* the 2nd rhythmolith). In the Early Campanian, the transgression reduced abruptly (the zone *Bathisiphon vitta, Recurvoides magnificus*: the 1st rhythmolith). In North-American Western Interior basin, the abrupt reduction of transgression occurred in this period (T_8–R_8; Caldwell et al., 1993); this fact being in complete agreement with our data on the Early Campanian of Western Siberia. These are tops of the Slavgorodskian horizon and the superjacent transitional rock member designated as Kargasokskian. The latter is distinctly traced in the east of West-Siberian plain and encloses the impoverished assemblage of the Early Campanian foraminifera with primitive forms dominating.

In the Late Campanian to Early Maastrichtian (the 2nd stage, the 1st rhythmotherm of the 3rd rhythmothem), the transgression was strengthening both in the Western Interior basin of North America and in Western Siberia. The expanded Turgay strait still connected West-Siberian basin with European and southern seas. Their effect resulted in abrupt changes of West-Siberian foraminiferal assemblages taxonomy. As stated above, that was the second stage in these organisms' development which differed from the first one in the taxonomy of orders (the first one had been from Turonian to Santonian). The lithological composition of host rocks also changed: the Gankinskian horizon was composed of grey clays admixed with calcareous material. Calcareous secreted and secreted-agglutinated foraminifera were characteristic for this horizon. The Late Campanian zone *Cibicidoides primus* (the 2nd rhythmolith of the 1st rhythmotherm) was established within the bottom layers of this horizon (the 3rd rhythmothem). Superjacent are the Early Maastrichtian zone *Spiroplectammina variabilis, Gaudryina rugosa spinulosa* (the 1st rhythmolith) and the zone *Spiroplectammina kasanzevi, Bulimina rozenkranzi* (the 2nd rhythmolith). This corresponds to three transgressive-regressive cycles of the Late Campanian/Maastrichtian of North-American Western Interior basin.

To summarize, the great lithostratigraphical units designated as "cyclothems" by the American investigators and as "rhythmothems" by us can be subdivided into subordinate rhythmostratons: rhythmotherms, nearly corresponding to horizons or regional stages, and rhythmolithes conforming to biostratigraphical foraminiferal zones.

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SHALLOW WATER CARBONATE FACIES OF THE ZECHSTEIN LIMESTONE (CA1) FROM THE LOWER SILESIA, POLAND, AND THE LITHUANIAN/LATVIAN BORDER: A COMPARISON

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The Zechstein rocks from the eastern margin of the Late Permian basin outcrop in three places: the Lower Silesia, the Holy Cross Mts. cover and the Lithuanian/Latvian border (Fig. 1). The distance between the furthest points is ca. 800 km. In between, the Zechstein rocks are buried at the depths of several hundreds to a few thousands meters and are available only from boreholes. The Holy Cross Mts. area is excluded from the present study because of a small number of outcrops and a monotonous lithology of the Zechstein.

Both areas of interest have a long research history, reaching XIX century. The Permian limestones from Lithuania and Latvia were detected by Grewingk (1857), and from the Lower Silesia – by Gruenewaldt (1851) and Geinitz (1861). The newer data are included, among others, in the studies by Kadūnas (1967), Suveizdis (1975) and Kuršs and Savvaitova (1986) from the Lithuanian–Latvian area, and by Gunia (1962), Krason (1967), Peryt (1976, 1978, 1992), Sliwinski (1988) and Raczynski (1996) from the Lower Silesia. However, there is still lack of a detailed comparison of both marginal parts of the Zechstein basin. Thus, the aim of the present project is to find similarities and differences in sedimentological and diagenetic development as well as in palaeontological, taphonomic and geochemical features. Conclusions will be based on the earlier results obtained by Lithuanian, Latvian and Polish geologists, and on own field and laboratory studies, including microfacies, organic matter, chemical and isotopic analyses.

Below, basic information on the Zechstein Limestone from both areas is presented.

LOCALITY

The Zechstein outcrops in the Lower Silesia area extend 70 km in the North-Sudetic basin. Old limestone quarries and anhydrite and copper ore mines, as well as several dozen of natural outcrops were investigated. Nowadays, only the anhydrite mine in the vicinity of Lwowek Slaski is open. In the majority of the outcrops limestones and marls, 2–25 m thick, occur. They are correlated with the Zechstein Limestone (Ca1) of the central basin. Evaporites, siliciclastic rocks and dolomites which in places lie above might be equivalents of the upper part of PZ1, PZ2 and PZ3 cyclothems. The Zechstein is covered with the Triassic clastic rocks.

The carbonate Naujoji Akmenė Formation, which is the equivalent of the Zechstein Limestone (Ca1) in the easternmost part of the basin, outcrops along the Lithuanian/Latvian border and reaches in places 35 m of thickness. Carbonate rocks are overlain by the Quaternary glacial tills and, further to the south, Triassic clays. The upper Zechstein cyclothems are known out of the area of outcrops, only from boreholes from middle and south Lithuania. Disused quarries at Nigrande and Satini, operating ones at Kumas, Karpėnai and Menčiai, as well as natural outcrops in the Venta River valley and in Paplaka were examined and sampled.

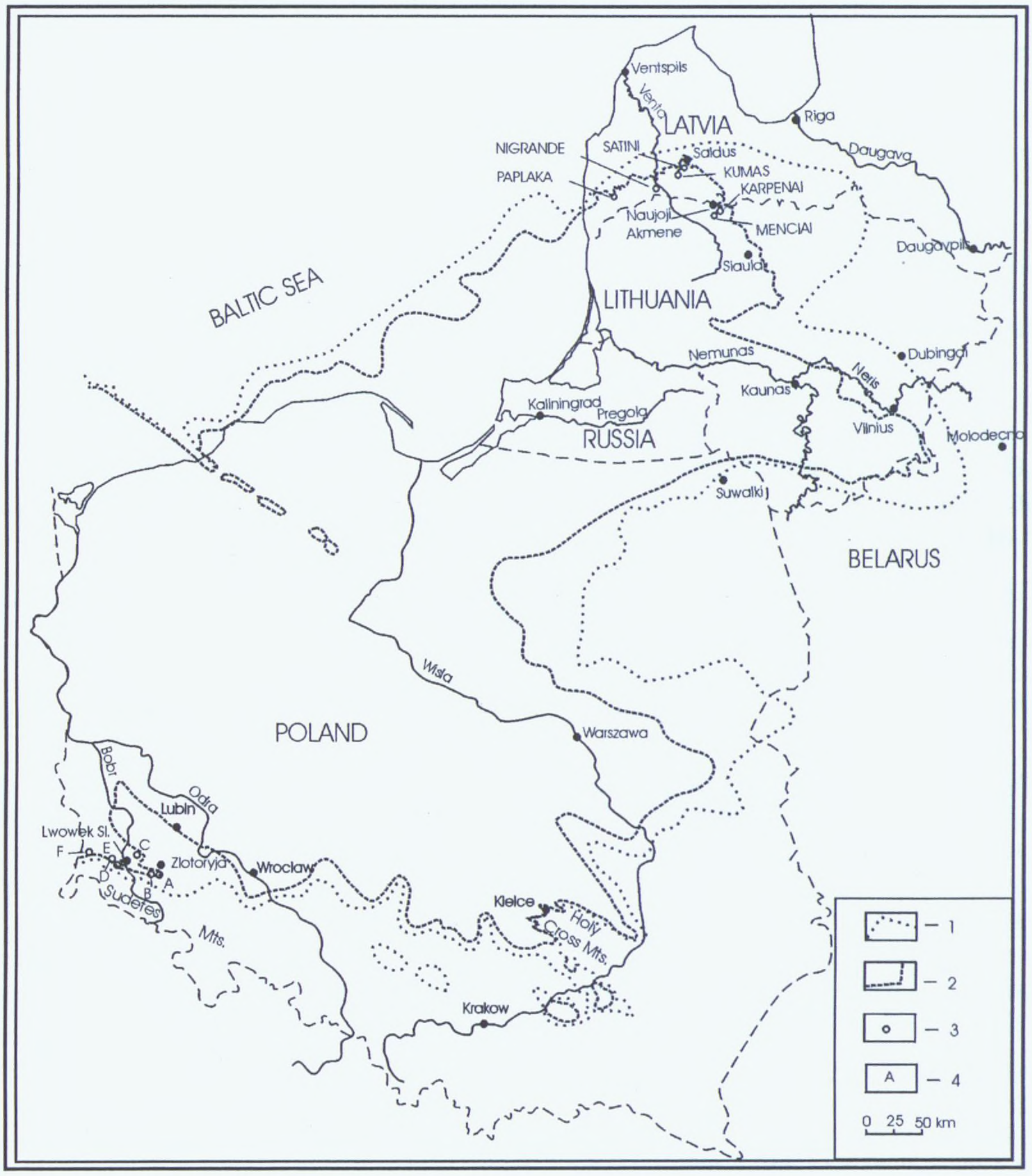

Fig. 1. Extent of the Zechstein Limestone in the eastern part of the European Basin (modified from Wagner, 1994). Explanations: 1 – original extent; 2 – current extent; 3 – studied outcrops; 4 – outcrops in the Lower Silesia: A – Leszczyna, B – Nowy Kosciol, C – Grodziec, D – Ploczki, E – Niwnice, F – Nawojow Slaski.

LITHOLOGY

The Lower Silesian Zechstein Limestone consists of three parts: thick layered limestones in the lower part, alternating, thin marls/limestones above, and limestones with a distinct siliciclastc admixture in the upper part. In places, dolomitic limestones were noticed in the upper part of the section.

In comparison to the Lower Silesia, the Lithuanian/Latvian Zechstein carbonate rocks contain far less terrigenic material. They are also divided into three parts: conglomerates/sands and marls in the lower one, limestones in the middle one and dolomitic limestones in the upper part (Kadūnas, 1967).

DEPOSITIONAL ENVIRONMENT

The Zechstein Limestone of the Lower Silesia was deposited in vast lagoons situated in a shallow reef-rimmed shelf. The lagoonal zone was ca. 50 km wide. Sedimentation took place both below and above storm wave base. Tempestites and shoal sediments form the majority of the carbonate sequence. There is a distinct record of relative sea-level changes with dominating shallowing-upwards trend in the Zechstein Limestone section. Deposits of small lagoons and shoals located very close to the land, originated in conditions of ecological stress, were preserved in the south-eastern part of the area. On the other part of this territory conditions were more normal marine, close to an open shallow shelf. The sandy-gravel coast, of relatively narrow width, was located close to the mountains built of metamorphic and volcanic rocks.

Similarly, the limestones of the Naujoji Akmenė Formation from Lithuania/Latvia formed in a coastal shallow shelf zone, between land and reefs. The remnants of the latter are preserved ca. 70–100 km south-west, in central Lithuania. The record of relative sea-level changes with a shallowing-upwards trend is also visible in the section. However, the coastal zone was much wider in comparison to the Lower Silesian, which was caused by an earlier long-term erosion of the platform and flattening of the bottom on a large territory.

FOSSILS

The most abundant and taxonomically diverse fossils were described from the lower part of the Zechstein Limestone. 150 genera of all groups of fossils characteristic for the Zechstein basin were identified in the Lower Silesian part (Riedel, 1917; Raczynski, 1996): bivalves, gastropods, brachiopods (articulate and inarticulate), bryozoans, echinoderms (echinoids, crinoids, ophiuroids), nautiloids, fishes, ostracods and foraminifers.

Similarly, 160 taxa of fossils were described from the Naujoji Akmenė Formation of the Lithuanian area (Suveizdis, 1975). The majority of them, however, were recognized from the boreholes. The limestones from the Lithuanian/Latvian border are rich in fossils although taxonomically less differentiated; bivalves, gastropods, foraminifers, ostracods and echinoderms dominate. Lack of bryozoans, articulate brachiopods and corals is striking. Moreover, many genera known from the area located further from the Zechstein Sea margin do not occur here.

Echinoderms, present in both studied areas, indicate normal water salinity. Epifaunal sessile organisms are more abundant in the Lower Silesia area, which suggests higher bottom stability. On the other hand, more bivalves in the fossilized living positions were noticed from Lithuania/Latvia, which indicates rather rapid burial of a soft sea bottom and more numerous mobile infauna.

DIAGENESIS

Widespread neomorphism, incipient dolomitisation and dissolution were dominating processes in the limestones of both compared territories. Significant karst phenomena occur in limestones of the Naujoji Akmenė Formation, especially in the vicinity of Naujoji Akmenė. The age of a sediment fill is considered as Jurassic, but there are still not sufficient evidences. Its statement is one of the project goals.

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ABNORMALITIES OF THE DEVELOPMENT IN SOME LUDLOWIAN *PRISTIOGRAPTUS* COLONIES

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Abnormalities of the development in the colonies of Wenlockian–Ludlovian monograptids were studied by Urbanek (1958, 1997) and Teler (1998). Similar abnormalities of Llandoverian diplograptids were studied by Paškevičius (1972), Ulst (1975) in the Baltic States.

The similar abnormal rhabdosomes of pristiograptids where obtained by chemical preparation from Milaičiai-103 borehole at the depth 1256.3 m (***Pseudomonoclimacis tauragensis*** biozone).

The collection contains colonies of pristiograptides, which have two virgulae (Fig. 1), colonies with additional thecaes in the proximal part of rhabdosome and the colonies which first thecae shape is not typical. The divergence of virgula on studied specimen begins at the edge of aperture of the second thecae, above the apex of prosicula. The second virgula is inclined to the dorsal part of rhabdosome at the angle of 20 ° from the main virgula. The second virgula is coverd by pseudochitinical membrane of the rhabdosom up to the divergence of virgula at the forth thecae.

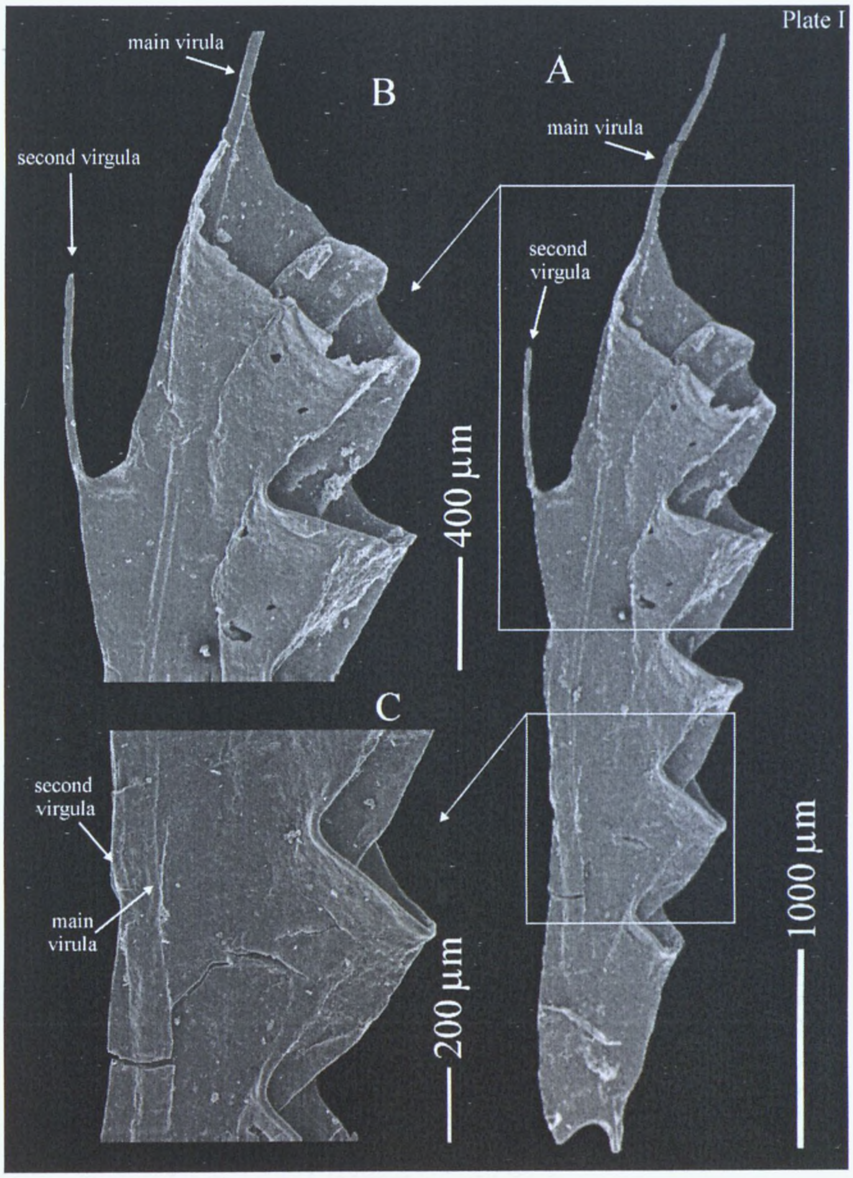

Fig. 1. **Abnormalities in the development of *Pristiograptus* sp. from Milaičiai-103 borehole, depth 1256.3 m, *Pseudomonoclimacis tauragenis* biozone.** ***A – general view; B – the distal part of rhabdosom with two virgulae; C – divergence of virgula.***

Possibly the virgula began diverge at primary stage of the colony development. Virgula started develop at the same time as the first thecae. The secondary virgula has diverged later from main virgula and it was growing bended independently. The secondary virgula probably finished its development due to the appearance of the fourth thecae, whilw the main virgula kept developing and caring out its function.

The divergence of virgula could be related to mechanical damage. There are two scenarios of abnormal virgula development which can be suggested: 1) provirgula was cleft and its separate parts developed individually at the sicula stage; 2) provirgula was broken off and two individual virgulae were developed from it. But this mechanical damage couldn't happen during rhabdosome developed, because the secondary virgula is covered by pseudochitinical membrane together with the main virgula. Although, these two scenarious can't explain other abnormalities (additional thecaes in proximal part of rhabdosom and not typical shape of the first thecaes) observed in other colonies in the same sample. May be the latter abnormalities are related to the biological degeneration of individuals or to their reaction to the environmental changes. More over, there is a large decreasing in graptolite fauna diversity during the ***tauragensis*** biozone and significant the graptolite fauna renewing appears above this zone.

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THE STRUCTURING OF TEXTUAL DATA FOR DATA MINING IN GEOLOGY

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ABSTRACT

Geological data are mainly descriptive ones, especially those related to formation of geological bodies in regions (like description of layers in Lithuania). Such presentation of data prevents to use powerful and popular methods of data mining in geology. The article presents a model for structuring textual data, enabling the application of data mining techniques in geology.

INTRODUCTION

Classical geology, being a completely descriptive science for a long time, still is on a wayside of data mining (DM) methods. Data mining is valuable for its ability to extract knowledge from statistical properties of data, and has an explosive growth in many applications but geology [7]. The reasons of such backwardness are: complexity of the domain itself, lack of impartial research methods, fuzzy description of objects, etc. Most of geological data are presented as a large scale descriptive text databases, and they do not fit to statistical analysis forthright [1, 2]. Even more, data cannot be converted to numeric or other precise form easily, and sophisticated structuring methods are needed. The article has a goal to present a model and corresponding software design developed for structuring textual data in geology. The resulting outcome of modeling procedures enables the techniques of data mining to be applied to geological data.

INITIAL DATASET OF GEOLOGICAL DATA

Geological data used to develop the model are related to Lithuania, and particularly they are presented in datasets resulted from broad drillings performed in a search of new oil fields. Primary dataset consists of

90 boreholes, 4000 layers overall [2]. Each layer is described with 35 lithological and palaenthological parameters for primary and secondary rocks both, plus 9 general parameters for the entire layer (e.g. collectors, other rocks, rock contacts, upper contact, and etc.). The data described refers to the Silurian system. This system in Lithuanian cross-section is one of the most complicated [5].

These data are presented in the form of visual description of characteristics of a geological layer by historically standardized lithological, palaenthological, and other definitions. Semantically distinct data are separated into different dimensions, e.g. name of the primary rock, name of the secondary rock, macro fauna of the primary rock, and etc. These datasets are not suitable to DM and cannot be understood in-depth by traditional DM techniques because of its features (descriptive text format, complex internal text structure and primary subjective estimation, etc.). Nevertheless, small subsets of descriptive text overlap with geochemical and geophysical datasets, which are numeric, structured and much more precise [2]. This feature can serve for additional control of data transformation.

STRUCTURE OF THE MODEL

The main idea behind a model suggested is to spread complicated, inner-structured, descriptive-textual formatted dimensions of the initial dataset into a number of scalar dimensions, according to suitable classification scheme. The resulting dataset should retain the maximum possible amount of information, and it should be possible to reverse the computer ready data into a human readable format (Fig. 1).

The model and computational system include the following building blocks:

1. functions (further procedures) – mainly for initial dataset pre-compilation (preparation for main algorithm);
2. data model – a number of entities (tables) for the functions and algorithms to work with;
3. forward-run algorithm for human-to-computer dataset transformation;
4. reverse algorithm for inverse process – computer-to-human dataset transformation.

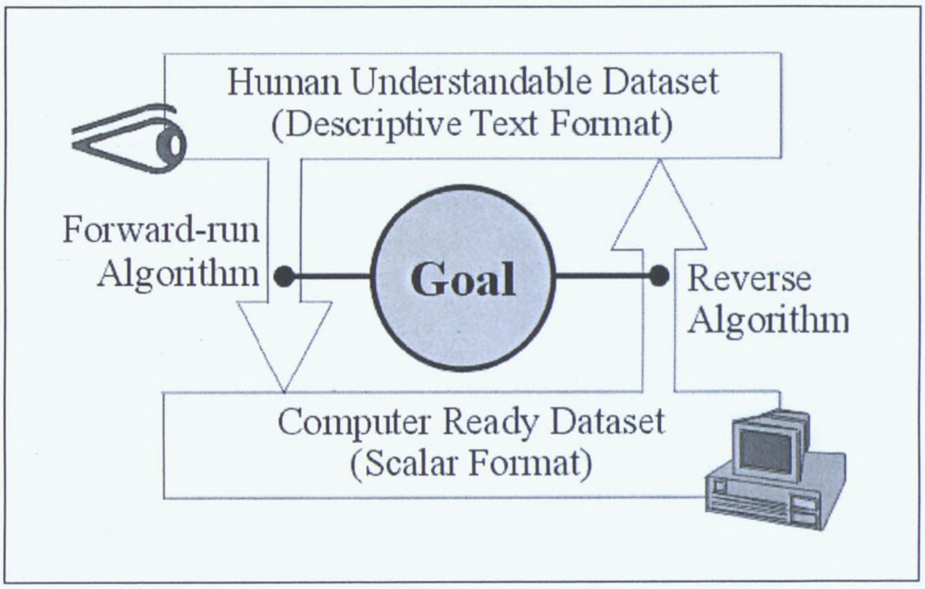

Fig. 1. The smooth, information preserving transformation between scalar and descriptive datasets.

FUNCTIONS OF THE MODEL

To begin a detailed description of computation algorithm, a number of functions must be introduced. These functions apply and work together with data model described in the "Data Model" section. The table 1 below provides an introduction to the sequence of functions that are hierarchically well defined and labeled. The root function is labeled by D, and is named 'Prepare textual descriptive geological data for analysis'. Further procedures evolve out of the atomic functions; or will remain for a grouping purpose. During the computation functions must be followed in the order given. To run these functions the data model must be constructed; and the main forward-run algorithm is implemented.

Table 1. Function hierarchy of the system.

Level	Label	Definition	Master Function	Atomic
1	D	Prepare textual descriptive geological data for analysis		
2	D0	Audit primary textual descriptive geological database	D	
3	D01	Audit borehole ID and coordinates	D0	A
3	D02	Audit borehole layers' consistency	D0	A
3	D03	Audit and correct parameter values	D0	
4	D031	Correct word conjunctions	D03	A
4	D032	Exclude typing errors	D03	A
4	D033	Exclude systematic errors	D03	A
2	D1	Group parameters	D	
3	D11	Group parameters according to descriptive domain (format Groups)	D1	A
3	D12	Group parameters according to subjective meaning (format Attributes)	D1	A
2	D2	Mapping dimensions to attributes	D	
3	D21	Dimension extraction	D2	A
3	D22	Unique word/expression extraction out of attribute	D2	A
3	D23	Expression ranking	D2	A
3	D24	Expression to dimension mapping (determination of expression to dimension weight)	D2	A

DATA MODEL

Functions given above work with particular data model. This model is presented as entity-relationship diagram (Fig. 2). For a relational database entities should correspond to tables, attributes – columns, relationships – primary/foreign key attribute pair. Note: attributes reflect a 'must demand', so there may and probably should be more attributes.

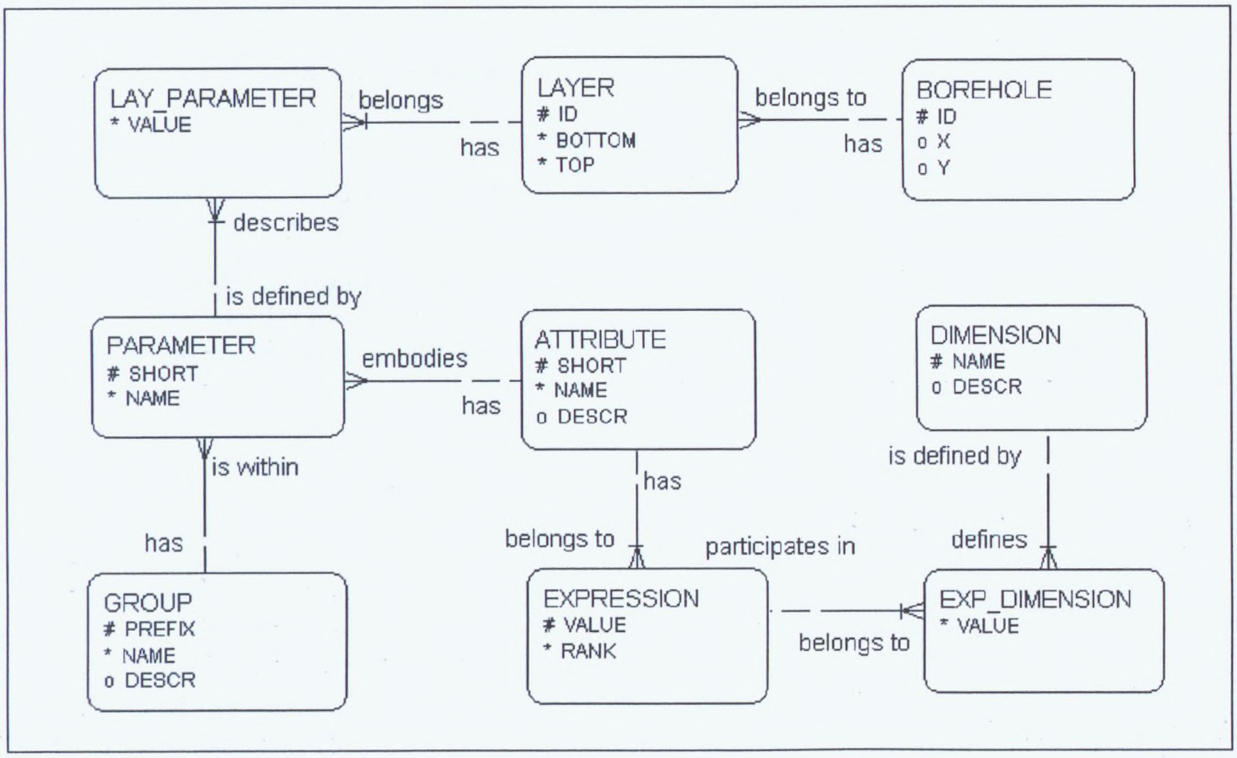

Fig. 2. ER diagram of Data Model.

FORWARD-RUN ALGORITHM

This algorithm works with data from the all entities. Below are given a possible algorithm in meta-language with comments.

```
//===== Create the resulting dataset
R¹.columns.new('borehole_id')
R.columns.new('layer_id')
//===== Main loop
for each b∈B².rows do
          for each l∈b.L³.rows do
//=== Insertion of new row into R
                    r←R.rows.new()
                    r.columns['borehole_id'].value←b.id
                    r.columns['layer_id'].value←l.id
                    for each p∈l.rows do
// === Looping expressions of the certain parameter
                              for each e∈p.attribute.E⁴.rows, in asc order by _
                                        e.columns['order'].value, do
// === Comparing a given expression with the real value and if it maches –
// === loop each dimenssion of the given expression
                                        if p.columns['value'].value matches _
                                        e.columns['expression'].value then
                                        for each d∈e.D.rows do
// === Get a column name for a resulting table: [group prefix]_[dimension name]
                                                  col_name←d.columns['name'].value
                                                  if p.group<>nil then _
                                                            col_name←p.group.columns['prefix'].value || _
                                                            "_" || col_name
                                                  end
// === If the specified column does not exists in the recordset – _
// === add it and set default value to nil
                                                  if not exists(R.columns[col_name]) then
                                                            R.columns.new(col_name)
                                                            R.columns[col_name].default←nil
                                                  end
// === If a given dimenssion is still not set by the more valuable expressions – _
// === lets do this now!
                                                            if r.columns[col_name].value==nil then
                                                                      r.columns[col_name].value←_
                                                                      d.columns['value'].value
                                                            end
                                                  end
                                        end
                              end
                    end
          end
end
```

R^1 – resulting dataset
B^2 – boreholes dataset (ref. to ER diagram entity BOREHOLE)
L^3 – layers dataset (ref. to ER diagram entity LAYER)
E^4 – expressions dataset (ref. to ER diagram entity EXPRESSION)

CRITICAL POINTS OF THE SEMANTICS OF DATA

The results of modeling depend on each function involved but the most sensitive is "Mapping Dimensions to Attributes" (D2) function group; especially its' atoms – "Dimension extraction" (D21), "Unique word/ expression extraction out of attribute" (D22) and "Determination of expression to dimension weight" (D24). Selected dimensions reflect all features of the attribute. The examples below provide a solution possible for domain – Silurian system in Lithuania, predominantly carbonaceous-clayey rocks.

ATTRIBUTE: Name of the Rock; DIMENSIONs: quantity carbonaceous material, quantity of terrigeneous material, roughness of terrigeneous material (particles), the level of dolomitization.

ATTRIBUTE: Texture of the Rock; DIMENSIONs: level of crystallization, particle roughness, quantity of organic material, level of debris ness.

ATTRIBUTE: Color of the Rock; DIMMENSIONs: quantity of red color (R), quantity of green color (G), quantity of blue color (B).

Unique word/expression extraction should be hardly evaluated. Some of terms can be omitted if they are not important and do not fit to dimensions selected; inclusion of some terms may cause addition of new dimensions. Weights of terms should also be hardly evaluated to fit the requirements. Weight margins must be standardized, e.g. consistent from 0 to 1 or Boolean: true – 1, false – 0.

FUTURE DM COMPUTATIONS

After the resulting scalar dataset is composed DM techniques may be applied. Classification methods for example [6, 8], produce rock classes – a standard attribute set that can be reversed into a human readable format for revision and further data input control.

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PROGRESS IN ESTONIAN QUATERNARY STRATIGRAPHY DURING THE LAST DECADE

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Stratigraphic studies aimed at compiling regional legends for geological mapping and different applied investigations have always been of high priority in Estonia. During Soviet time, based on several tens of thousands drillcores and medium- and large-scale mapping, rather well-grounded stratigraphical schemes of the Quaternary (Raukas & Kajak, 1995), Late-glacial (Pirrus & Raukas, 1996) and Holocene (Raukas et al., 1995) were compiled and published. These schemes were accepted on May 6, 1993 by the Estonian Stratigraphic Committee. But, in fact, there were not very many new scientific ideas in the field of the Quaternary stratigraphy after the First Baltic Stratigraphic Conference in Vilnius in 1976, when the main outlines of the current stratigraphic schemes were accepted. Not a single new interglacial or interstadial has been recognised in Estonia during the last decade. Also the number of new drillings is scanty and, therefore, there is little hope to find intermorainic sections, which could alter the existing stratigraphic charts.

In the official stratigraphical scheme of the Quaternary deposits in Estonia (Raukas & Kajak, 1995), lithostratigraphic terms are used as basic units. Formation, as a fundamental unit, is used in a meaning of glacial and interglacial episodes in the event stratigraphy. Big stadial episodes in a meaning of event stratigraphy are comparable with subformations. Although interglacial sediments are differentiated on the basis of spore and pollen and other fossil evidence, and pollen assemblage zones underlie their description, for the unity of the scheme, even here lithostratigraphical terms Prangli (= Eemian Interglacial) and Karuküla (= Holsteinian Interglacial) formations were preferred. All stratigraphical units (formations and subformations) are based on well-investigated stratotypes. For the interglacials, there are composite stratotypes, *i. e.* the localities where both boundary and unit stratotypes are recognised. For the tills, areal stratotypes are used. All proposed units are well mappable.

The Upper Järva (Upper Weichselian) late-glacial deposits are divided into Arctic (Bølling, Older Dryas) and Subarctic (Allerød, Younger Dryas) chronozones and based on regional type sections. According to the decision of the INQUA Congress in Paris in 1969, the Holocene/Pleistocene boundary is accepted as 10 000 ^{14}C years. As to the chronology, we have to point out in every single case which is the time scale we are speaking about: conventional radiocarbon years, dendrochronologically revised radiocarbon years, dendroyears, TL-years, varve years or the so-called "calendar" or "absolute"years. All these scales are based on various phenomena, and transitions from one scale to another are not always correct. Mechanical use of different time scales has already resulted in a lot of misunderstandings and the situation will worsen in coming years.

As there is little hope to get new sections due to limited drilling possibilities, in the last decade we have paid much attention to methodical problems and improvement of dating methods. In the Quaternary stratigraphy the age of tills is of special interest because it enables to correlate lithologically similar formations over a vast area. The age of tills is generally determined by bedding conditions, by their position with respect to interglacial or interstadial deposits. Unfortunately, the latter are rather uncommon. Besides, most of unconsolidated intermorainic organic deposits were strongly crushed by the advancing glacier during the succeeding glaciations, and they are often embedded as erratics in younger sediments. Some researchers (Liivrand, 1991) believe that practically all continental interglacial deposits have been displaced not only horizontally, but occasionally also a considerable up-thrusting and/or folding has occurred. Therefore, older blocks are found standing in a position of tens or even more than a hundred metres (as in the Karuküla section) above their normal stratigraphical position. As principally different opinions have been expressed about the bedding conditions of interglacial deposits in Estonia (Kajak, 1995), new dating methods are needed for solving the problem.

Unfortunately, there are practically no methods for direct dating of tills. Out of all types of Quaternary deposits, tills have proved the most complicated objects for the luminescence dating. The obtained data, as a whole, are in bad correlation with supposable geological ages of tills, because the basic prerequisite of the method – sufficient daylight exposure – is not normally fulfilled for such kind of deposits.

The occurrence of *Portlandia arctica* shells in glacigenic sediments of Central Latvia allowed to use effectively the electron-spin-resonance (ESR) method for dating of glacially transported raft (Molodkov et al., 1998) in a disputable section. Unfortunately, in Estonia we have not found Pleistocene sections with datable shells. The radiocarbon method is limited with the last 40 000 years and only a few of the radiocarbon dates for the Late Weichselian chronology are reliable.

A good potential to distinguish the main Middle–Late Pleistocene warm-climate/high sea-level events and correlate marine and terrestrial deposits over wide areas has an integrated approach based on two independent methods and sources of climato-stratigraphical information – the electron-spin-resonance (ESR) chronology of warm-climate related deposits and the palynological record of vegetation response to climatic variability and palaeoenvironmental events (Molodkov & Bolikhovskaya, 2002). This integrated approach allows establishing the number and stratigraphic position of the palaeoenvironmental events (stratigraphic units) missed in the stratigraphic schemes of the Baltic countries.

In the Late Pleistocene stratigraphy of the Baltic States, the following two problems are most topical: (1) the number of Weichselian (Järva) subformations (glaciations, big stadials) and (2) the beginning of the Late-glacial in Estonia.

In the Weichselian section, there are two clear till beds (Valgjärv and Võrtsjärv subformations). However, it is not yet clear whether the Valgjärv till belongs to the Lower or Middle Weichselian. We hope to solve this problem by OSL dating of all possible outcrops of intermorainic glacioaquatic deposits.

Traditionally, the beginning of the Late-glacial interval in Estonia is placed at the time the deposits of the Raunis Interstadial started to accumulate in Central Latvia (dated by different laboratories as 13390±500 (Mo–196), 13250±160 (TA–177), 13320±250 (Ri–39) conventional ^{14}C ages. Based on the results obtained through new excavations, drillings, and complex palynological and malacological investigations

(V Zelčs a.o.), the organic sediments in the Raunis section seem to be of Early Holocene age and are covered with pseudotill, probably with colluvial sediments. As the majority of dates obtained from submorainic and intermorainic sequences with organic remains in Estonia (Petruse, Viitka, etc.) are younger than one would expect on the basis of the conventional radiocarbon method and, at the same time, due to the hard-water effect some organic layers above the uppermost till are dated at 13 000–14 000 yr BP, new dating possibilities should be found.

Some promising results have been obtained through varve countings and palaeomagnetic studies of glaciolacustrine varved clays. The varve chronology produced for the Peipsi Basin (Hang, 2001) was extended to north-western Russia into the Luga and Neva basins (Hang et al., 2000) and via floating Finnish chronology to the Swedish time scale (Cato, 1987). However, there are too many gaps between Lake Peipsi and Sweden, and the accuracy of the estimated rate for the ice recession during those gaps is extremely disputable. Tentative age assignment for Lake Peipsi suggests that accumulation of varved clays in Peipsi Ice Lake took place between 13,500 and 13,100 varve years BP (Hang, 2001). This is comparable with the age of the Pandivere (Neva) ice marginal formations (13 300 "calendar" years BP) from the Lake Onega area by Saarnisto and Saarinen (2001) and agrees with the traditional opinion (Raukas et al., 1969) that the southeastern part of Estonia and the central part of the Haanja Heights emerged from under the ice sheet some 13 000 conventional radiocarbon years ago.

In the Holocene stratigraphy we have adopted the chronozone boundaries of the Nordic countries (Mangerud et al., 1974) and differentiated eleven pollen assamblage zones which are time-transgressive (Raukas et al., 1995). Stratotype sections for the Holocene deposits have not been officially accepted yet. This work is still in progress. Investigation of laminated lake sediments, which was started several years ago, inserts some hope for the improvement of local chronology.

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SUCCESSIONS OF CHITINOZOANS AND OSTRACODES AS TOOLS FOR TIME-ROCK CORRELATIONS IN THE ESTONIAN SILURIAN

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The best-known way for the use of fossils in dating sedimentary rocks is establishing various biozones by the occurrence of particular species or events in sections. Several zonal schemes are available for the chitinozoans (Nestor 1976; 1994) and ostracodes (Meidla, Sarv 1990; Meidla & Sarv 1997), all successfully used in chronostratigraphy of the region (Kaljo 1990; Nestor 1997). Problems arise when one wants to integrate biozones based on different groups (for instance, Loydell et al. 1998) or areas (Verniers et al. 1995).

The successions of fossils (their taxa) can be composed using various numerical methods, first of all to construct a scale for dating or time-rock correlation of sections characterised by them. Usually it means that total (Shaw 1964; Rubel & Pak 1984; Guex 1989) or optimal ranges of taxa (Agterberg 1990) are involved. The both ranges are not immediately observable in any particular section and the use of the timescales based on them for time-rock correlations is someway different from the use of zonal stratigraphic units based as a rule on local ranges of taxa. But proper timescales found by numerical methods have an obvious advantage, offering point-correlation needed in recent chronostratigraphy (identification of boundary stratotypes) as well as in sedimentolgy (estimation of sedimentation rates although in relative mode). Earlier at least two such attempts have been made concerning partly or completely the ostracodes and chitinozoans from the Baltic Silurian and touching also the integration of distributional data of different fossils into a common timescale (Rubel & Pak 1986; Rubel & Sarv 1996).

In the present case the algorithm DISTR (Pak1984; 1986) has been used to construct such successions on the Llandovery and Wenlock chitinozoans and Silurian ostracodes from the Baltic. The successions must reflect the origins and extinctions and not local appearance-disappearence of taxa. In this way environmental or biogeographical impacts on the successions are eliminated and they are an ordinal timescales. This will be achieved by selection of different initial sets of sections, intervals and taxa according to needed assumptions on the distribution of taxa in the studied sections. Then many possible scales with different resolution power and temporal-areal validity will be constructed for the selected stratigraphical interval.

A total of 151 chitinozoan species from the Llandovery to Wenlock part of 35 Estonian (most of them are drill cores) and 9 Latvian core sections were used to construct the corresponding timescales. One of them, based on all 151 taxa and 44 sections under study and processed according to the strategy 1 1 1 1 (see below) to maximize the number of taxa in the scale, includes 86 taxa arranged into the succession with 43 datum planes. In spite of the low percentage of the included taxa (57%), this scale has too many taxa with little or no temporal (zonal) value, for instance, those occurring only in one section or even in one sample which were excluded from the processing. The remaining 122 taxa from 44 sections gave the scale of 76 taxa forming 36 datum planes there. The use of this scale (briefly denoted as 122/76/36) in the time-rock correlations (dating) is yet unreliable as it is difficult to estimate the correlative value of some taxa included into the scale according to the strategy 1 1 1 1. The latters mean that equal weights are given to all factors (strong and weak contradictions, missing ties, frequency) for excluding taxa from the scale. Preferring taxa with wide occurrence in the possible scale by increasing the frequency coefficient (the last number in the strategy denotation) leads usually to decreasing a number of datum planes (resolution power of the scale) on account of removing taxa with small correlative (zonal) value from the scale. Thus the scale, got by the strategy 1 1 1 10, contained 73 taxa of chitinozoans (60% of all), including also 16 zonal species (Nestor 1994) and forming 33 datum planes there (briefly 122/73/33).

It was decided to use the scale 122/73/33 for checking contemporaneity of the existing positions of boundaries of the regional stages, as well as for the estimation of relative rates of sedimentation, especially gaps, in the studied sections and distribution of origins and extinctions of the studied species in time. Such datings indicated that the boundaries of stages have been mostly drawn into the same time interval. However, deviations of these estimations are observed in some sections. It is interesting to note that the boundary stratotype of the Jaani Stage itself is deviating if compared with the most of the boundaries of the same age in other Estonian and Latvian sections.

Sedimentation rates in the sections can be well compared between contemporaneous intervals of two sections by means of common to them palaeontological or other events. But comparison of sedimentation rates along the same section (in time) is possible assuming more or less regular nature of the scale used for that or, in our case, assuming the even rate in origins and extinctions of taxa in the used scales. Such a regularity is rare in practice if it exists at all, and there is no obvious reason to suppose it in the case of the chitinozoan scales here. Nevertheless, proper ordinal timescales must always indicate sedimentary gaps by means of the missing corresponding taxa or parts of their ranges as elements of the timescale used in a particular section. Such gaps or very low sedimentation rates (according to remarkably short section intervals for many timescale units) can be also discovered by the chitinozoan scales in the studied sections. It must be only noted that the discovered gaps have not so continuous areal or temporal distribution as supposed previously by means of the zonal schemes.

The use of the chitinozoan scale 122/73/33 to estimate their evolutionary rate according to their origins and extinctions requires the involvment of the taxa excluded from the scale by processing. Nevertheless, one can conclude by the scale taxa alone that in the studied interval and area, no remarkable evolutionary crises have occurred in the chitinozoan evolution.

The 351 species of ostracodes from nearly 200 sections with different age and extent (most of them outcrops) from the Silurian of Estonia, Latvia, Lithuania, Poland and Gotland Island represent a too motley database to be processed by the algorithm DISTR. In the given case the timescales desired are constructed on geographically as well as facially close sections, for which the occurrence of certain ostracodes (actual or potential common taxa) is supposed. The taxa described mostly in open nomenclature and occurring in one section or even sample (see above) were excluded at once from further processings. In summary, that means to use only the distributional data on 242 ostracode species (taxa) from 85 Estonian, Latvian and Lithuanian sections, 42 of which were outcrops. Additionally some subsets of this dataset were formed and processed to test some hypotheses.

One scale constructed by means of the strategy 1 1 1 1 for the Baltic dataset above consists of 105 species arranged into a succession which gave 40 datum planes. As too many zonal species were not included into this scale, the demand to include 17 zonal species into the scale has been applied. As a result scale of only 92 species but with 41 datum planes was obtained. Although the latter scale (242/92/41) includes all the previously noted index taxa, the former one (242/105/40) is more operational: it can be applied easily in more sections.

The timescales based on ostracodes have been used just like the chitinozoan-based ones for checking the contemporaneity of the stage boundaries drawn by previous papers and for estimating sedimentation and evolution rates. In the given case, the use of ostracode timescale(s) was more productive in the Wenlock to Pridoli of the Baltic sections. The scales gave datings which in most cases supported the positions of the boundaries of the regional stages drawn in the previous work, although there were several notable deviations.

The studied ostracodes indicate at least one clear turnover (crisis) in their regional development. It falls into early Ludlow, *i. e.* into the time of episodic but remarkable regression of the Silurian sea in the Baltic. As this crisis concerns ostracodes from nearshore facies, the most favourite one for ostracodes, the increased extinctions, followed by simultaneous numerous appearances (origins) of new taxa may also denote a gap due to missing these taxa from more complete off shore sections but which were not included into the given scale according to contradictionary relationships. The case shows a very particular nature of the scales, their greater or smaller dependence on environmental factors.

But, if the scales constructed this way are indeed timescales, then both chitinozoan and ostracode scales should give the same results in estimating the positions of stage boundaries or sedimentation rates, especially gaps, in the studied sections. At the moment, both groups are studied from too few common sections to allow these comparisons. The small number of common sections excludes also the possible joint processing of the chitinozoan and ostracode data, especially, in limits of the Llandovery and Wenlock. But, as both groups support each other, dating the stage boundaries as well as gaps to the same places in these few cases, the scales selected above for time-rock correlations should be recognized as sufficiently well timescales. Or they are presumed to be as such up to the appearance of evidences which prove the opposite. It is also obvious that every such scale exists for a particular dataset (taxa and sections) and must be tested or even changed if new data on the distribution of taxa are added.

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EXPANSION OF BROAD-LEAVED TREES NEAR AND AT THEIR NORTHERN RANGE LIMIT (AN EXAMPLE FROM ESTONIA)

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The spread of broad-leaved trees (*Ulmus, Tilia, Quercus, Fraxinus, Carpinus* and *Fagus*) in Estonia is reviewed on the basis of 47 radiocarbon dated pollen diagrams (Fig. 1) of which 24 are analysed with high temporal and taxonomic resolution. To synthesise such an archive of pollen data is a rather complicated task, because the diagrams originate from various landscape types, lakes and mires of different size, and the material analysed at different levels. The diagrams used in our study have a radiocarbon dated empirical limit (start of a continuous pollen curve), rational limit (rise of the curve) and the maximum pollen occurrence of the described tax. In the case of the absence of radiocarbon dates, the age of these events was interpolated. Radiocarbon dates quoted in the text are uncalibrated radiocarbon years before present (BP). As the ^{14}C dates originate from different materials, the dates given here should be taken as approximations rather than precise estimates. Although the number of dated diagrams (47) and radiocarbon dates (450) is rather high, there are still areas left in Estonia, which need well-dated pollen diagrams.

To visualise the immigration and expansion of *Ulmus*, *Tilia* and *Quercus* isopoll maps for mentioned tree species were reconstructed for a 500-year interval. Such maps have been widely adopted, because they provide a good summary of palynological data (Huntley, 1988). However, several sources of error, as the pollen sum, long-distance transport and the size of the basin affect the validity of these maps (Davis & Sugita, 1997). *Ulmus* and *Quercus* are relatively good pollen producers and rational pollen limit was used to establish their spread. For *Tilia* the empirical limit was used for the same purpose. However, this does not exclude the possibility that the first arrival of these trees occurred earlier.

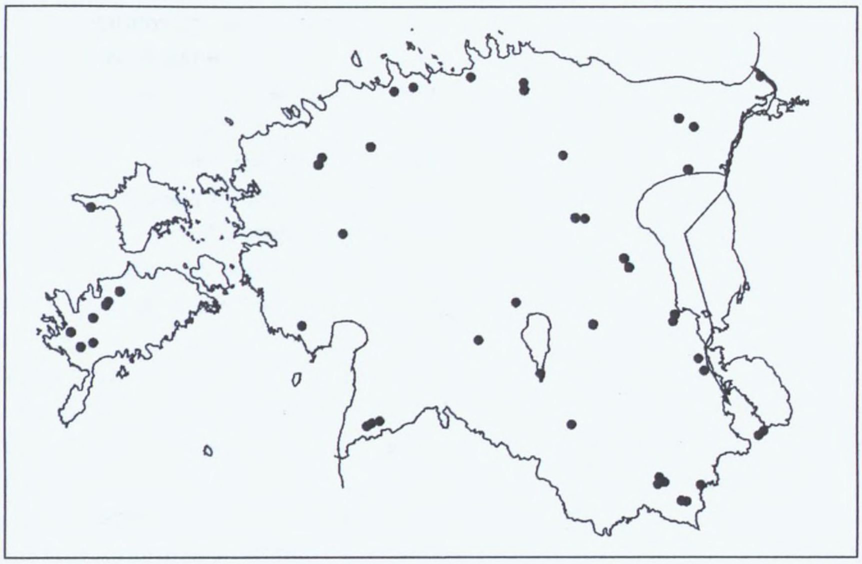

Fig. 1. Location of the examined biostratigraphical sites in Estonia.

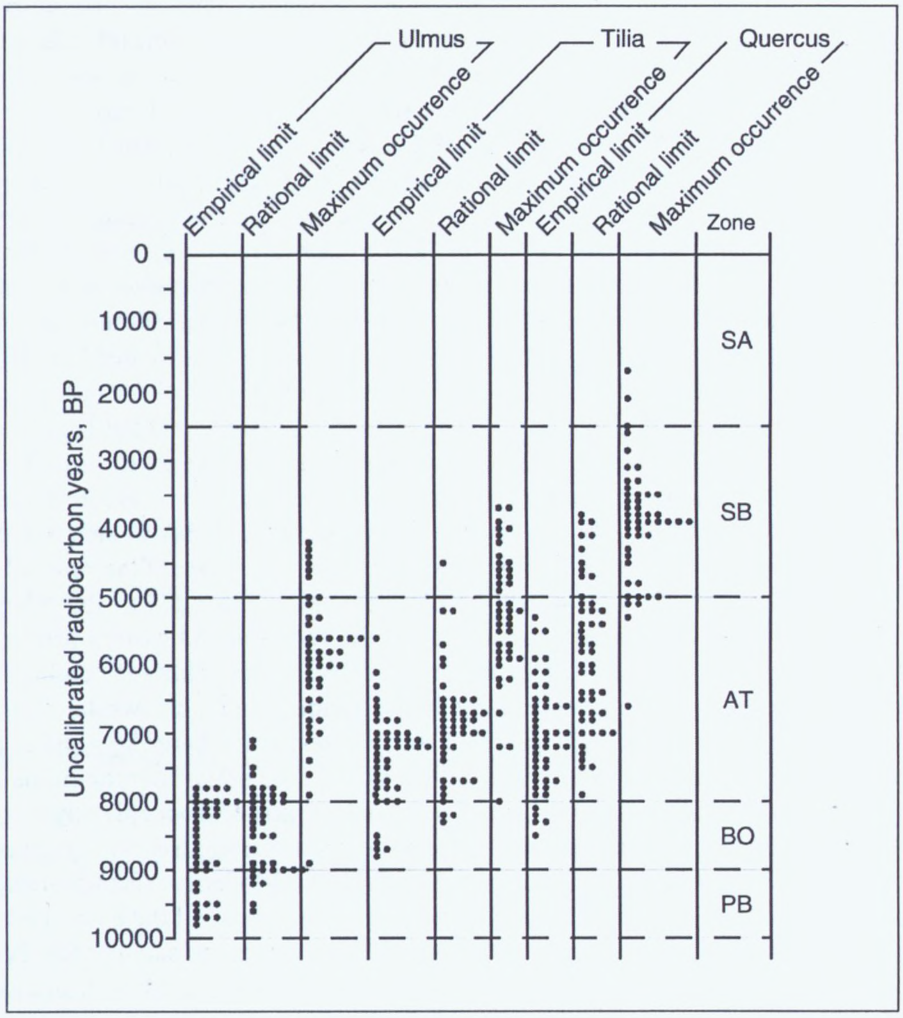

Fig. 2. **Time-distribution of the empirical and rational limits of elm *(Ulmus)*, lime *(Tilia)* and oak *(Quercus)* pollen and their maximum occurrence.**

Ulmus became locally established already at the end of the Preboreal in southern, western and northern Estonia (Saarse and Veski, 2001). Climate amelioration and available habitats have substantiated such rapid immigration and spread of *Ulmus*. Contrast to the quick spread maximum occurrence of *Ulmus* is dispersed over more than 3500 years, between 7900 and 4200 BP, being still more or less well pronounced between 6500–5500 BP (Fig. 2). During its major maximum *Ulmus* together with *Tilia* was the main forest-forming tree in broad-leaved stands on uplands, heights and archipelago. *Ulmus* pollen dropped since 5500 BP interpreted as *Ulmus* decline, caused by human activities and pathogenic attack as well (Peglar, 1993). In several sites *Ulmus* decline is associated with diminished frequency of *Tilia*, followed by new recovery and new declines, showing local clearances. The scattered increase in *Ulmus* representation *ca* 2000–1000 BP could be the result of increased pollen dispersal in more open landscape. *Ulmus* pollen decreased throughout the Subboreal and Subatlantic and by 1000 BP its pollen frequencies became extinct or decreased below the rational limit. *Ulmus* is a minor pollen type within the northern limit elsewhere in Estonia and its modern pollen percentage is generally below 1%. At present there are two native *Ulmus* (elm) species in Estonia (*Ulmus glabra* Huds. and *U. laevis* Pall.), whereas the range limit of *U. glabra* lies more northerly, along the Estonian coastline and the northern range of *U. laevis* crosses Estonia diagonally. The *Ulmus* stands are scarce, being present in protected locations in the Sakala Upland valleys and in the sheltered fore-*klint* mixed forest. Single planted elm trees are more common in farmyards.

Tilia. The rapid and dispersed spread of *Ulmus* is in contrast with the time-transgressive spread of *Tilia* in Estonia. The quite obvious north-western immigration route for *Tilia* can be confirmed, which coincides well with the routes presented by Huntley and Birks (1983). At about 8500 BP *Tilia* colonised the south-east and reached the north-western coastal area at about 7200–7000 BP, which makes the approximate range rate of 160 m yr^{-1} (Saarse and Veski, 2001). This is 2–3 times slower than the average for the European mainland (Huntley and Birks, 1983). The maximum distribution of *Tilia* pollen before 6000 BP resembles its immigration pattern, culminating earlier in eastern and later in north-western Estonia. The highest extent of *Tilia* pollen is recorded between 6000–4000 BP, first mostly in the sites it had reached first. After 5000 BP *Tilia* started to decline, which was more pronounced between 4600–4500 and after 3600 BP. Only on the Haanja Heights *Tilia* pollen maintained to keep high frequency up to 3000 BP. In several sites the *Tilia* decline is coherent with that of *Ulmus*, in others it is registered several hundred years later. At present the percentage of *Tilia cordata* Mill. pollen exceeds the limit of 1% only in the deciduous forest sites. Natural *Tilia* stands are locally present in the north-eastern and southern regions of Estonia, single lime trees are growing in the fore-*klint* forest and on the archipelago. Its northern limit lies in South Finland and its eastern limit is farther than that of any other broad-leaved trees treated here (Hultén, 1950).

Quercus was the next tree to arrive at about 7900–7500 BP, but it remained rather rare up to the beginning of the Atlantic. Single *Quercus* trees could have been present earlier, before 8000 BP, because already at that time *Quercus* pollen forms a continuous curve in several sites (Saarse and Veski, 2001). *Quercus* expanded slowly and by 7000 BP it had distributed unevenly over the studied area. According to new pollen records (Veski, 1998; Poska and Saarse, 1999), oak reached the northern coastal areas of Estonia at about 6000 BP. It took 3200 years for oak to reach the empirical limit and 4100 years to surpass the rational limit in all the examined sites (Fig. 2). Between 4200–3500 BP the oak population expanded as the other broad-leaved trees declined. The earlier maximum frequencies of oak in the southern and eastern profiles, and the late maximums in the north-western ones could be explained by differences in climatic conditions, which were more continental in the south-east and more maritime in the northern and north-western coastal regions of Estonia. Besides climate, a series of local factors, such as the topography, hydrology, soils, and conditions influencing seedling, also determined the distribution of oak, and the other thermophilous trees. *Quercus* culmination became to an end at about 3000 BP, when its main habitats were occupied by spruce. According to Huntley and Birks (1983), oak immigrated into our territory from the south-west, which is hard to prove on the base of the Estonian material only (Saarse and Veski, 2001). Modern surface pollen data show that unlike *Ulmus* and *Tilia*, *Quercus* pollen is present in low values in almost all the examined sites. *Quercus robur* L. at its northern boundary is native in Estonian oak stands. Oak forests account for about 0.5% of the total forest area in Estonia. Natural oak stands are rather common on the fertile calcareous soils of West Estonia and the Island of Saaremaa, and in the central Estonian fertile floodplains.

Fraxinus record starts at 7500 BP in several sites of Saaremaa and West Estonia and *ca* 6500 BP in the eastern part of Estonia. The pollen percentage of *Fraxinus* is low, commonly 1–2%, exceeding 3% level in a very few sites. *Fraxinus* started to decrease at about 3500 BP, almost simultaneously with *Quercus*. The geographic range of the *Fraxinus excelsior* L. (common ash) extends a little farther to the north than that of *Quercus*, being in Estonia also close to its northern range limit. *Fraxinus excelsior* L. is the only species of ash that grows naturally in Estonia, mainly on Saaremaa Island, West Estonia and around Tartu, forming 0.1 % of the total forest area.

Carpinus. Occasional grains of *Carpinus* were recorded more often since 3000 BP. It forms a low 0.3–0.6% continuous curve only on a very few high-resolution pollen diagrams. *Carpinus betulus* L. (hornbeam) is not native in Estonia today and the limit of its geographic range crosses Lithuania.

Fagus pollen is found only occasionally and it never formed a continuous pollen curve, *e.g.* surpassed the empirical limit. This does not exclude that single beech trees have grown natively in the western archipelago of Estonia.

So, the broad-leaved forest started to spread into Estonia already in the Preboreal and at its maximum between 6500–5000 BP was represented by lime, elm, ash and oak. Hornbeam appeared in the broad-leaved forest mostly since 3000 BP. Immigration and expansion of the thermophilous broad-leaved trees has a certain succession, caused mostly by climate, edaphic demands, local environmental conditions and to a lesser extent by human interference. By 6000 BP, elm, lime and oak had reached their present-day range limit or were near to it. Taking into account the correction for productivity and dispersal (Andersen, 1970), the elm-dominated woods grew mostly before 7000 BP. At 7000 BP, the share of elm and lime woods was almost equal, after that lime dominated in the broad-leaved forest everywhere in Estonia. The corrected pollen values show that lime could have made up to 20–30% of the total tree cover. Lime survived longest, for about 4000 years, on the Haanja Heights.

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STATE-OF-ART OF QUATERNARY STRATIGRAPHY OF LITHUANIA

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Introduction. Development of stratigraphic schemes is closely related with development of geological mapping as the mapping requires detailed, standartised and reasonable stratigraphic sudivision of rocks and deposits. On other hand, the mapping is main source of new geological information for the stratigraphic studies. Main results of the geological mapping of period of 1997–2002: the integrated geological mapping at a scale of 1:50 000 of has been completed the whole coastal region of Lithuania, a part of Norhern Lithuania (border area with Latvia) was mapped, the new revised Quaternary geological and geomorphological maps were compiled for the whole country at a scale 1:200 000 (author R. Guobytė). The maps at a scale of 1:50 000 are supported by relevant analytical data (lithological, biostratigraphical, etc.) sets. High quality Quaternary geological maps (scale 1:50 000) were available for the 30% of territory of country by August 2002.

For the geological mapping of Quaternary in Lithuania so far was in use the stratigraphical scheme (1994), assigned for practical geological investigations. Up to present time new assumptions for improvement of the Quaternary stratigraphic scheme became available. First of all, the new stratigraphic guide (Grigelis *et al.*, 2002) presented a new stratigraphic classification of Quaternary deposits. A new version of stratigraphic scheme should be based on this classification.

Below a brief outline of new data and interpretations of the Middle and Upper Pleistocene is presented.

The problem of the Snaigupėlė Interglacial. The Snaigupėlė Interglacial was introduced and so far regarded as Middle Pleistocene Interglacial (Kondratiene, 1996), however at first it was characterized as an untypical Merkinė Interglacial. So far has not been obtained more relevant palaeobotanical data supporting the stratigraphic position of the Snaigupėlė Interglacial as one of the Middle Pleistocene, separating Žemaitija and Medininkai Glacials (climatoliths). Despite during more than twenty year long investigations it has not been possible to find certain analogs to the Snaigupėlė site as to the independent interglacial in other Baltic countries (Liivrand, 1997). Recent studies of the palaeocarpological complexes (Kisielienė, 2002) led to conclusion that there is no big difference between the Butėnai and the Snaigupėlė interglacials florae. Moreover, some of correlative Snaigupėlė type florae are richer in exotic species than the Butėnai florae, and can be attibuted to older period than the Žemaitija Glacial (Kisielienė, 2002). The application of lithostratigraphic criteria to the Žemaitija and Medininkai tills showed the stadial rank of these both tills (Satkūnas & Bitinas, 1994). Besides that, practically all sections of the deposits of Snaigupėlė Interglacial has the signs of dislocations and therefore can be regarded as occuring not *in situ*. Taking into account all above mentioned conclusions and assumptions, the Snaigupėlė Interglacial must be abolished as independent interglacial located in between Žemaitija and Medininkai glacials.

The Žemaitija–Medininkai Interstadial. The research of the Antaviliai section (northern suburb of Vilnius) led to new interpretation of this section (Satkunas & Hütt, 1999). The thermoluminescence dating of the Antaviliai sands gave the age 175 000 yrs (TL 633) and > 250 000 yrs. (TL 631). According to European time division (Bowen *et al.*, 1986), the age of the Antaviliai section looks closest to the Drenthe–Warthe ice free stage, which comprises geochronological interval 198–252 ka B.P. The sediments of contemporaneus Odra–Warta interstadial interval are known from the the classical area of the margin of the Warta glaciation in central Poland and are characterised by pollen assemblages of interstadial type (Gozdzik & Balwierz, 1994). So, assuming similar TL age, conditions of occurrence and paleopalynological characteristics, the correlation of the Antaviliai section with the Odra–Warta interstadial can be concluded.

The studies of the Vilkiškės outcrop (approx. 15 km north of Vilnius) (Gaigalas *et al.*, 2002) resulted with similar TL dates, that allow the correlation of the the sandy complex, formed in interstadial climatic conditions, with the Drenthe–Warthe interval. If additional reseach of this sandy complex will confirm its chronostratigraphic position, the Vilkiškės Interstadial can be introduced as the new Middle Pleistocene Interstadial, most probably, separating the Žemaitija and Medininkai stadials.

The Pamarys Interstadial. Within the geological mapping of the martime region a widely spread inter-till deposits has been determined (Bitinas, 2002). The sediments were localy named Pamarys Interstadial and consist of 20 metres thick fine grained sand with organic interlayers. The climatostratigraphic rank of this unit was determined on basis of palaeobotanical analysis. According to the optically stimulated luminescence (OSL) dating, the sediments were formed in time span 140–160 ka BP, and can be attributed to the pre-Eemian ice free period, presumably to the interstadial of the Medininkai (Warta) Glacial (Climatolith), that could existed at the end of this glaciation. The Pamarys Interstadial tentatively can be correlated with the Zeifen Interstadial in the Kattegat Depression (Bitinas, 2002).

Upper Pleistocene. Significant progress can be stated in detailed subdivision of Upper Pleistocene stratigraphy. Studies including lithostratigraphic and palaeobotanical analysies, physical datings have been carried out in key sites of Eastern and Southern Lithuania – Jonionys, Mickūnai, Medininkai, Buivydžiai (The Late Pleistocene... 1997, Satkūnas *et al.*, 1998; Satkūnas & Grigienė, 2000). Especially great value have Eemian–Weichselian sequences, located outside the maximum limit of the Late Weichselian ice sheet, that provide possibilities of finding continuous sedimentary records encompassing the whole Last Interglacial/Glacial cycle.

Based on the new data the following stratigraphic units can be revealed within the Upper Pleistocene: the

Merkinė Interglacial, the first Lower Nemunas (Weichselian) stadial – Nemunas 1a, the interstadial Jonionys 1 (correlated with the Brörup Interstadial), the Nemunas 1b stadial, the interstadial Jonionys 3 (correlated with the Oerel Interstadial), the Nemunas 2b stadial, the interstadial Mickūnai 1 (correlated with the interstadial Glinde), the Nemunas 2c stadial, the Mickūnai 2 Interstadial (tentatively correlated with the Hengelo Interstadial), the Nemunas 2d stadial, the Mickūnai 3 Interstadial (tentatively correlated with the Denekamp Interstadial), the Nemunas 2e stadial, the Mickūnai 4 Interstadial, the Nemunas 3 stadial (the Nemunas (Weichselian) maximum) and the Holocene.

So far in Lithuania as well as in other Baltic countries there are still not found any site with reliably determined Upper Weichselian interstadial or interphasial deposits. Therefore, "stadials", "phases" and "interstadials" of Upper Weichselian have not proper climatostratigraphic background in the Baltic states and the Upper Nemunas at present can be subdivided only into lithostratigraphic units.

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THE SUCCESSION IN ACCUMULATION OF TILLS

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According to the opinion of author the accumulation of the upper tills in Latvia took place at the time of the retreat of the glacial advances or glacial stages, which interrupted common recession of the ice sheet. It is known that this proceeded in basal part (debris-rich ice) of moving ice cover. But the mechanism by which ready tills had been accumulated is not clear else.

At present, the formation of tills in Latvia is discussed based on two very similar to essence models only. The first of them is the model of Lavrushin (1976), which supplemented on example of Latvia by Āboltiņš (Āboltiņš 1975, 1989) and the second – the model of Dreimanis (Āboltiņš, Dreimanis 1995). Āboltiņš established different types of glaciodynamical structures in the tills. According to both models they reflect dynamic structures, which existed in the moving ice cover. Not disclaiming the active processes, which took place in environment of moving ice cover, author as controversy considers that the dynamical structures observed in ready tills had been formed in a different way.

The above mentioned both models not taken into account, that the beds of ready tills could be formed in result of pressing out of their initial fragmental material contained in the ice on subglacial surface in forward (in the direction to the margin of the ice sheet). It took place in narrow zones along marginal slopes of the ice sheet.

The mechanism of the accumulation of the ready tills in the ice relatively to the recession of ice margin is shown in Fig. 1. The width of elementary zone, where had been formed the ready till, is L_t. This zone constantly transferred according to displacement of retreating margin of ice cover. The vertical pressure of the ice in limit of such zones allowed: (1) proceeding of the pressing out of initial material from debris to reach ice on subglacial surface and (2) proceeding of the accumulation of this material into the ready till. Here the thickness of the ice cover decreased and therefore the force of vertical pressure decreased also. The thickness of the ice cover ($Ht_1 = Ht_2 = Ht_3 = Ht_4$) is a main factor, which determine the necessary conditions for the accumulation of the till.

According to observed model, the different deformations and dislocations (folds, overthrusts, *etc.*), which occur inside of the ready tills reflect the features of the dynamics of pressing out of their initial material.

The pressing out of fragmental material from the ice to ascending direction is typical for high hilly glacial forms in uplands, especially for the tills lying in the bases of plateau-like hills covered by clay. The pressing out of the material contained in ice on area of lowlands in comparison those in uplands had been realised mainly in horizontal directions.

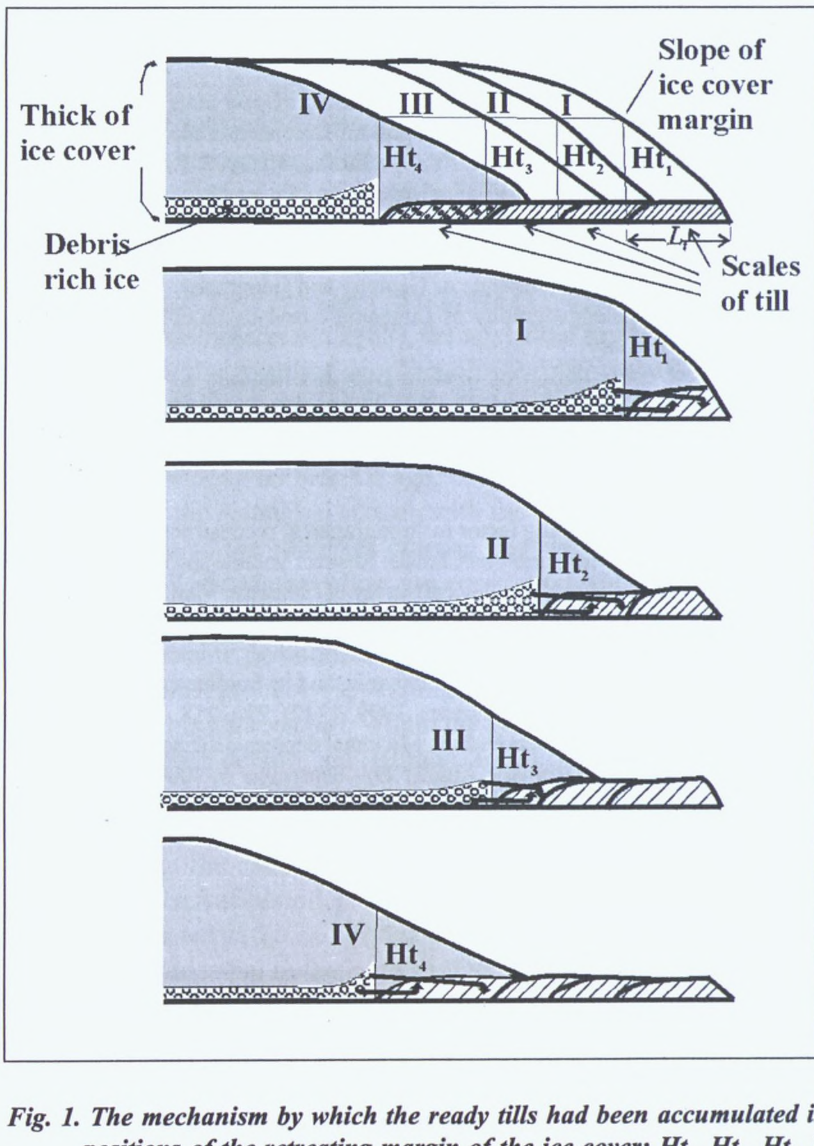

***Fig. 1.** The mechanism by which the ready tills had been accumulated in ice cover. I, II, III, IV – the different positions of the retreating margin of the ice cover; Ht_1, Ht_2, Ht_3, Ht_4 – the thickness of the ice, when the accumulation of the ready till is started ($Ht_1 = Ht_2 = Ht_3 = Ht_4$); L_t – the width of elementary zone, where the ready till is accumulated.*

Important information about the formation of tills could be obtained from peculiarities of the distribution of their thickness. The integral thickness of the tills clear is increased on the areas of recent uplands in comparison those in the lowlands (Savvaitov 1962). Later the increase of the thickness of the tills in uplands was established for each glaciations also and especially for tills of the Last ice sheet (Āboltiņš 1975, 1989, Āboltiņš, Straume, Juškevičs 1975, Meirons 1975). These data show that in limit of uplands, in comparison with lowlands, there existed the palaeogeographical conditions, which (1) preserved the glacial deposits from full exaration and (2) favoured to the accumulation of each till with increased thickness. Observed increase of the thickness of the tills is effect of often alternated of the advances and the retreats of the ice sheet. Such fluctuations were characteristic feature during common recession of the ice sheet within insular uplands. Therefore total sum of the thickness of the tills in the areas of the uplands is sufficiently larger.

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PALAEOENVIRONMENTAL INTERPRETATION OF ISOTOPE AND PALAEOBOTANICAL DATA FROM SECTION OF LAKE DŪBA, SE LITHUANIA

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Sediments formed during the Older Dryas – Subboreal in Lake Dūba, SE Lithuania, were studied by the means of stable isotope, pollen, diatom and ^{14}C analyses. Special attention was paid to the Preboreal and Boreal climatic periods when changes of the climatic conditions initiated alteration of the vegetation cover, fluctuations of the water level and development of the new environment.

As the authigenic calcite precipitates during the spring/summer photosynthetic bloom, the isotopic records reflect the equilibrium conditions (temperature and isotopic ratio) in lake water, when local evaporation and direct precipitation have their strongest isotopic effects (Punning *et. al.*, 2002). On the basis of the pollen survey the regional changes of the vegetation cover and mean climatic variables were determinate (Stančikaitė *et al.*, 2002). Diatom data shows the changes of the water level in investigated lake. Correlation of the palaeobotanical and stable isotope data suggests the possibility for the detailed evaluation of the environmental processes in particular area.

There are few levels of the $\delta^{18}O$ and $\delta^{13}C$ values changes that coincides with the environmental fluctuations registered on the basis of palaeobotanical data. The lowest value of $\delta^{18}O$ occurs in the sediments dated back to Alleröd period when water was cold and low in investigated lake. After that it's value gradually rises and reaches maximum at the Boreal/Atlantic transition. The increase of $d^{18}O$ value is related with a rise in the temperature of the water in the lake (Punning *et. al.*, 2002). The highest water temperature in Dūba Lake was at the Boreal/Atlantic transition, that could be related with the improvement of the climatic conditions. Abundant representation of broad-leaved trees in pollen spectra confirms the climatic amelioration.

Decreasing representation of the $\delta^{13}C$ was registered in the sediments formed during Alleröd and Younger Dryas periods as well as at Boreal/Atlantic transition. The background factors determining the formation of decreasing values of $\delta^{13}C$ of lake marl are the soil processes and development of organic matter in the lake catchment (Punning *et al.*, 2000; Punning *et al.*, 2002). Increasing amount of the organic matter in the sediments dated back to the Alleröd period could be related with the development of vegetation cover. High representation of *Pinus* and *Betula* pollen and their macroremains confirms the existence of light forest in the investigated area at that time (Blažauskas *et al.*, 1998). According to the isotope survey, the lowest amount of the organic material was fixed in the sediments dated back to the end of the Younger Dryas. Improvement of climatic conditions and changes in the vegetation cover initiated the increasing representation of the organic matter in Boreal/Atlantic transition.

Presented data suggest a good correlation between the results of isotope and palaeobotanical investigations. Significant changes in the carbon isotope content that took place in Alleröd ($\delta^{18}O$) and at the transition from Boreal to Atlantic ($\delta^{13}C$) coincide with the formation of the new vegetation cover, development of the soil layer and changes in whole palaeoenvironment.

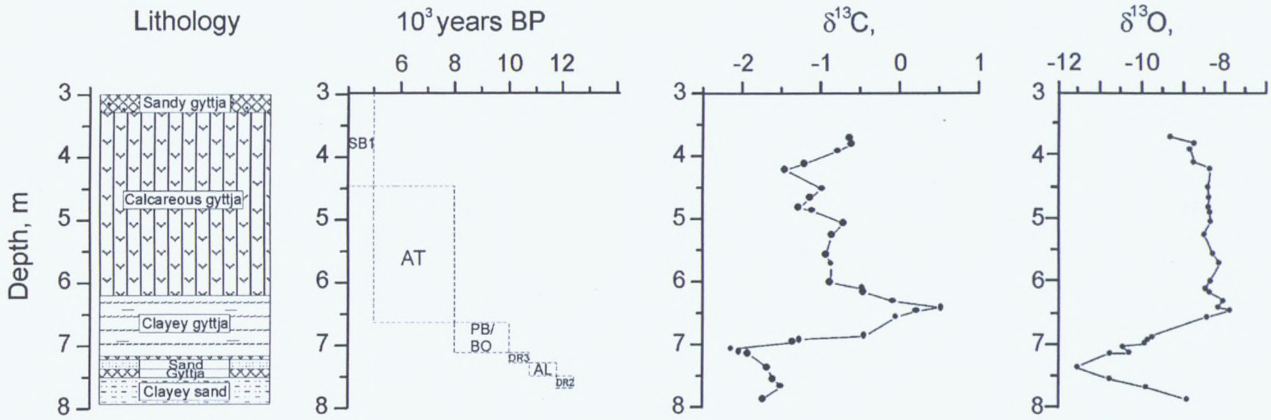

Figure. Correlation of the palaeobotanical and stable isotope data from Lake Dūba section.

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PALYNOSTRATIGRAPHICAL AND PALAEOGEOGRAPHICAL SIGNS FOR IDENTIFICATION OF INTERSTADIAL TERMS DURING RECESSION OF THE LAST ICE SHEET IN LATVIA

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The alternation both regional advances and accordingly following retreats of the ice cover had been typical phenomenon in dynamic of general recession of the Last ice sheet in Latvia (Savvaitovs, Veinbergs 1996). The firsts of them took place at the time of the coolings and reflected the stadial intervals, but the seconds – took place during of the warmings and reflected the interstadial intervals.

The terms of younger interstadials belonging to the Bölling and especially Alleröd are clearly established in Latvia. The basical palynostratigraphical signs of their are discussed by Stelle in principal (1996 a, b, 2000). The pollen of *Pinus* dominate in the composition of spores and pollen spectra, of the sediments of the Bölling (Be) and Alleröd (Al_1, Al_2, Al_3). The middle part of the Alleröd (Al_2) is distinguished by increase of *Betula*. It is the short period of the Lielauce cooling within the Alleröd, which existed about 11300±300 years ago. Note should be taken, that not easy to distinguish spores and pollen spectra of the Bölling from Alleröd, but between them are observed the stadial interval of the Older Dryas (Dr_2), the spores and pollen spectra of which are characterised by clearly increased values of *Betula and Artemisia*. Stratotypical site for the Bölling, Older Dryas and Alleröd is the Lielauce section, located in southeastern part of the East-Kurzeme Upland (Stelle 1966, Zobens, Putāns, Stelle 1969). The subdivision of the Alleröd into three stratigraphical parts was ascertained for bottom sediments in the Gulf of Rīga also (Stelle, Jakubovska, Savvaitov 1994, Stelle, Jakubovska, Savvaitov, Kalniņa, Juškevičs, Kovaļenko 1992).

Besides, older interstadials having differrent age are known in Latvia also. Their sediments lie between different beds of the upper moraine. The identification is based on as palynological data and available ^{14}C dates and as well as palaeogeographical evidences. The stratigraphical position of these interstadials in the curve of the Last ice sheet deglaciation on area of Latvia is shown in Fig. 1 and commented, as it will be said bellow. Separate glacial stages and their succession at the time of deglaciation were established by Āboltiņš, Veinbergs, Stelle, Eberhards (1972) and Veinbergs (1968, 1972). Possible two younger bands of marginal forms were established else (Savvaitov 2001; Savaitovs 2000). The accumulation of the bands of ice margin formations was due to the advances of glaciers and are supported by the facts, that the older sediments are covered by the beds of younger moraine.

The sediments refered to the Raunis Interstadial are found in several sites (Ceriņa, Danilāns, Dreimanis, Jakubovska, Stelle, Zelčs 1998, Ceriņa, Jakubovska, Savvaitov, Stelle 1998, Danilāns 1973, Meirons 1992, Savvaitov, Staume 1963, Segliņš 1986, Stelle, Savvaitovs, Jakubovska 1999). Their are known at stratotypical Raunis site (palynozones: Ra-1, Ra-2, Ra-3; optimal palynozone *Pinus* (Ra-2); ^{14}C dates 13390±500, 13250±160, 13320±250 BP), Līdumnieki site (palynozones are analogous those at Raunis site, ^{14}C dates reach to 13080±60 BP) and probably at Krikmaņi site (palynzone *Pinus,* but ^{14}C date – 12148±30 BP is rejuvenated ?). The Raunis Interstadial is a time span between the Pampāļi–Ranka (Pa) and Linkuva (Lk) Stages. Two beds of upper moraine are wide spread in central Latvia (Āboltiņš 1963; Savvaitov, Straume 1963) and are the formations of these Stages. Besides sites, which had been discussed for grounding of separate advances of the ice sheet earlier (Savvaitov, Straume 1963), at present there is known the section of Kraņciems Quarry. Here, two lithological different layers of upper moraine (Pampāļi-Ranka and Linkuva Stages) between which lie the intertill sand with gravel and pebble could be seen in Quaternary above of the dolomite of the Daugava Formations (D_3dg).

Next younger Plieņi (Pl) Stage is separated from the Linkuva Stage by the Mazsalaca Interstadial (clay and silt at Mazsalaca sites, lies under the upper moraine). Interstadial sediments between advance of the ice sheet of the Plieņi and Valdemārpils (Vl) Stages and as well between younger Staicele (St) and Lejassalaca (Ls) Stages are not found yet, but intermorainic clay at Bridagi, Upites, *etc.* sites located in lower reaches of the Vitrupe River (Danilāns 1973) probably belong to interstadial (Lejasvitrupe) between the Valdemārpils and Staicele Stages.

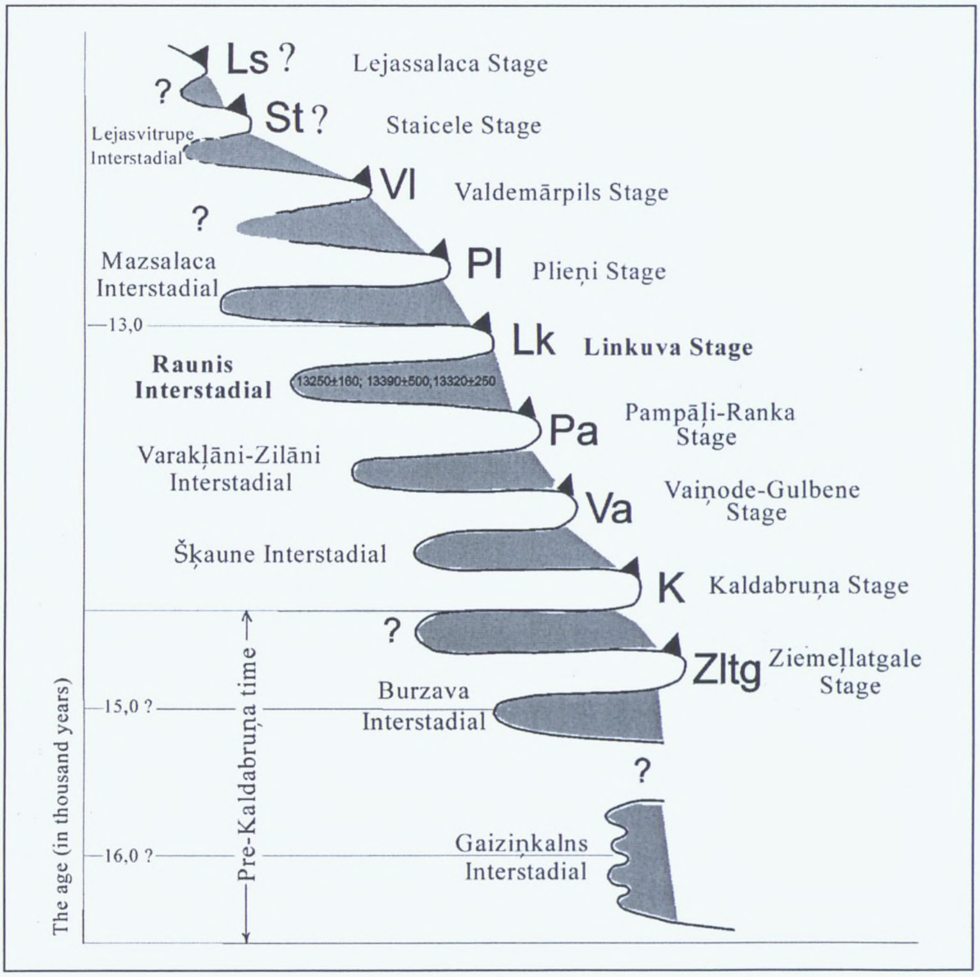

Fig. 1. Succession of older interstadials in curve of the Last ice sheet deglaciation on area of Latvia.

The reconstruction of stratigraphical positions of the interstadials older, than Raunis is the complicated problem (Veinbergs, Savaitovs 2001). The oldest of them existed at the pre-Kaldabruņa time, when in area of the central and eastern uplands of Latvia in small and open englacial basins had been accumulated clay, covering the tops of recent plateau-like hills. This process began firstly developed at Gaiziņkalns only, but afterwards took place in regions of other high plateau-like hills also. Therefore this interstadial is called by Gaiziņkalns name. The spores and pollen typical for interstadial conditions, small mollusc shells and decomposed remains of organic matter are found in the clay at Gaiziņkalns.

The younger interstadial of the pre-Kaldabruņa time is Burzava. The sediments of the Burzava Interstadial are characterised by cold spectra of vegetation (*Betula, Artemisia, etc.*), and lie "in situ" under layer of upper moraine in northern part of the Latgale Upland (Krūkle, Stelle, Veinbergs 1963). The sediments of Burzava site in spite of young radiocarbon dates (Meirons 1992) – from 7945±250 to 12970±120 BP (undoubtedly rejuvenation) belong to interstadial of the pre-Kaldabruņa time, but the moraine overlaping these sediments reflects the last advance of the ice sheet (Ziemeļlatgale Stage – Zltg) on area of the Latgale Upland. Interstadial sediments between the Ziemeļlatgale Stage and following the Kaldabruņa Stage (K) are not found yet. The clay composing cover on the tops of the plateau-like hills, which are known in northern part (Northern Elevation) of the Latgale Upland (Veinbergs, Krūkle 1965), probably could be connected with this interstadial. The Šķaune interstadial is the interval of retreating ice sheet, which existed between the Kaldabruņa and Vaiņode–Gulbene (Va) Stages. It is established due to palaeogeographical signs and was connected with the time, when the clay in Šķaune rampart had been formed. The Varakļāni–Zilāni Interstadial separates the Vaiņode–Gulbene and Pampāļi–Ranka (Pa) Stages. The sediments of the Varakļāni–Zilāni Interstadial are represented by clay of Krustpils ice dammed basin and as well as clay of englacial basins composing the ramparts within the East-Latvian Lowland. The palynozone *Betula* and palynozone *Pinus* are characterising the pollen spectra of clay of the Varakļāni–Zilāni Interstadial. The palynozone *Pinus* characterised the bed of peat within clay also, which was found and investigated by Dreimanis (1939) near Audze.

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CHEIROLEPIDS FROM THE MIDDLE AND UPPER DEVONIAN OF LITHUANIA AND KALININGRAD DISTRICT

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The remains, later named *Cheirolepis*, have been described first 1828 (Murchison, 1828) as the "second Gamrie ichthyolite". The name *Cheirolepis* was applied 1835 for the first time (Agassiz 1835–36). Two species were described in this issue: *Cheirolepis trailli ("Cheirolepis traillii")* and *Ch. uragus*. There are nowdays known 4 species: *Ch. trailli* Agassiz 1835, *Ch. canadensis* Whiteaves 1881, *Ch. gaugeri* Gross 1973 and *Ch. gracilis* Gross 1973. *Ch. trailli* and *Ch. canadensis* are biological taxa described by articulated specimens, *Ch. gaugeri* and *Ch. gracilis* – by the isolated scales.

The scales of *Cheirolepis* in Lithuania are defined from the Rezekne (Middle Devonian?) till the Pakruojis Regional Stage (Upper Devonian). About 300 samples from 64 Lithuanian boreholes and 1 of Kaliningrad District have been examined for our investigation. The most of studied localities are situated in the east and middle Lithuania. Boreholes from the western Lithuania have occured less perspective. Part of them (Nida-44, Salantai-59, Naujoji Akmenė-76, Šilutė-76) have not yielded palaeoniscoids, even the samples were taken in detail. The most part of samples belongs to the Middle Devonian rocks of Narva Regional Stage, less – to the Middle Devonian (?) Rezekne, Aruküla and Burtnieki regional stages, and the smallest part – to the Upper Devonian. The scales of palaeoniscoids have been found in 191 sample. Good preserved *Orvikuina* scales are very rare, mostly there are met only their fragments. On the contrary, *Cheirolepis* is represented by mostly good preserved microremains (scales).

77 cheirolepid scales were found in 18 boreholes. Their largest part falls to Middle Devonian Upninkai Group and Kernavė Formation (Fig. 1).

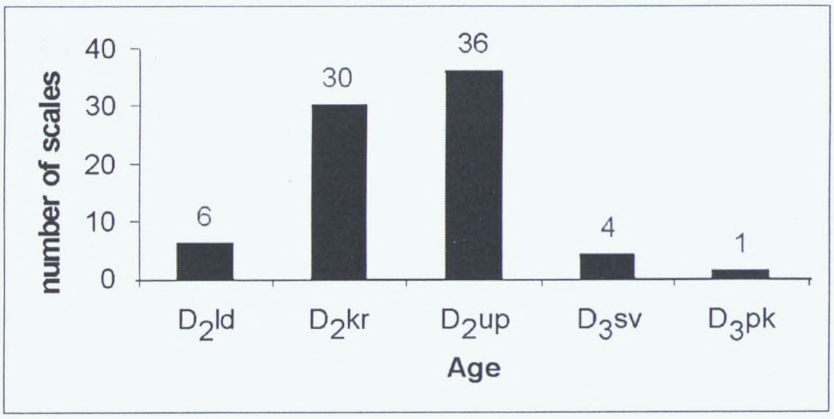

Fig 1. Amount of cheirolepid scales.

This material is insufficient to define the species, that is why the open nomenclature is used.

Class OSTEICHTHYES

Subclass Actinopterygii

Order Palaeoniscida

Family Cheirolepididae Pander 1860

Genus *Cheirolepis* Agassiz 1835

Type A

(Plate 1, figs. A–B)

Type locality. Riešutynė-321 borehole, depth 208.2 and 252.6 m.

Type stratigraphic unit. Middle Devonian, Eifelian, the Ledai Formation.

Stratigraphic range. Middle Devonian, Eifelian, from the Ledai Formation to the top of Kernavė Formation.

Material. See Table 1.

Description. Scales are rhomboid, rounded. Size 0.5–1 mm. Crown ornamented with the closelyset, fine ridges. The ridges are rectangular (surface is flat) in cross section. The interval between ridges is larger in the anterior part of the scale, posterior ward it narrows until flows together to the "sawing" branches. Ridges are straight, going slantwise through all the crown. Crown has to 10 ridges. They are parallel in posterior part of the scale. In the anterior part, ridges are grouped into two parts (lobes), seperated by the groove. The crown is rounded, in the anterior part of some scales depending on the position on the body. The peg and socket are absent. The crown is covered by thin ganoine layer.

Localities. See Table 1.

Type B

(Plate 1, figs. C–G, M; plate 2, figs. A, C)

Type locality. Dvoriki-2 borehole, depth 1009.2 m.

Type stratigraphic unit. Middle Devonian, Eifelian, the Kernavė Formation.

Stratigraphic range. From Middle Devonian, Eifelian, Kernavė Formation to Upper Devonian, Frasnian, the top of Pakruojis Formation.

Material. See Table 1.

Description. The scales are subrhombic, rounded. The base is larger than the crown. Size about 0.5–1 mm. The base can be wided anteriorly, posteriorly and/or laterally. Crown ornamented with 4–6 large ridges. Ridges are triangular (the edge is sharp) in the cross section. Ridges are unparallel, some are subradial, joint in the posterior part of the scale. One or two ridges look like stretched "S". These are the largest ridges and, possible, primary. Other ridges are shorter, converging into the primaries. Ridges are located disorderly. The crown is covered by ganoine layer.

Type C

(Plate 1, figs. H, K–L, N–O; plate 2, B, D)

Type locality. Drūkšiai-51 borehole, depth 178.2 m.

Type stratigraphic unit. Middle Devonian, Eifelian, the Kernavė Formation.

Stratigraphic range. From Middle Devonian, Eifelian, the Ledai Formation to Upper Devonian, Frasnian, the Šventoji Formation.

Material. See Table 1.

Description. Scales are rhomboid, rounded, high. Size about 0.5 mm. The base is larger than the crown. There are opened vascular canals on the neck of the scale. Crown ornamented with 7–9 large ridges, located accuratlly, stretching posteriorly. The cross section of ridge is rounded, semiarc in form. Ridges flow together into the sprout (often the sprout is broken) at the posterior part of the scale. Central ridges are almost straight, lateral – bended. The crown devides into lobes. The crown is covered by ganoine layer.

Localities. See Table 1.

Type D

(Plate 1, figs. I–J)

Typice locality. Drūkšiai-51 borehole, depth 178.2 m.

Type stratigraphic unit/stratigraphic range. Middle Devonian, Eifelian, the Kernavė Formation.

Description. Scales are rhomboid (almost rectangular). Size about 0.5 mm. Length exceeds width for 2 times. The base a little bit larger the crown. The ridges are not long, located slantwise, and orderly. The posterior part of the scale can be cuted into 2 branches. On the neck of the scale there are openings of vascular canals. It is possible, the scales are related of type C, but from another position on the fish body. The crown is covered by thick ganoine layer.

Localities. See Table 1.

Table 1. Distribution of cheirolepids according to morphotypes and stratigraphic units.

<table>
<tr><th rowspan="2">System</th><th rowspan="2">Series</th><th rowspan="2">Stage</th><th rowspan="2">Group, Formation</th><th colspan="4">Localities of cheirolepids</th></tr>
<tr><th>Type A</th><th>Type B</th><th>Type C</th><th>Type D</th></tr>
<tr><td rowspan="10">Devonian</td><td rowspan="7">Upper</td><td rowspan="7">Frasnian</td><td>Pakruojis</td><td></td><td>Lygumai-45 borehole, depth 375.5 m (1 scale)</td><td></td><td></td></tr>
<tr><td>Stipinai</td><td></td><td></td><td></td><td></td></tr>
<tr><td>Pamūšis</td><td></td><td></td><td></td><td></td></tr>
<tr><td>Įstras</td><td></td><td></td><td></td><td></td></tr>
<tr><td>Tatula</td><td></td><td></td><td></td><td></td></tr>
<tr><td>Pliavinai</td><td></td><td></td><td></td><td></td></tr>
<tr><td>Šventoji</td><td></td><td>Šešuvis-11, 525.3 m (1)</td><td>Kriūkai-146, 326.4 m (1); Šešuvis-11, 525.3 m (1), 534.0 (1)</td><td></td></tr>
<tr><td rowspan="3">Middle</td><td>Givetian</td><td>Upninkai</td><td></td><td>Drūkšiai-52, 162.0 m (1); Dvoriki-2, 987.0 m (1), 989.0 m (1); Ledai-179, 186.7 m (1); Lygumai-45, 543.0 m (1); Stačiūnai-8, 349.3 m (4), 352.2 m (2); Šešuvis-11, 573.5 m (8)</td><td>Dvorikai-2, 980.9 m (2); Kunkojai-12, 501.8 m (2); Lygumai-45, 553.5 m (1); Palanga-318, 660.1 m (1); Stačiūnai-8, 349.3 m (1), 352.2 m (1); Šešuvis-11, 573.0 m (1), 573.2 m (2), 573.5 m (6)</td><td></td></tr>
<tr><td rowspan="2">Eifelian</td><td>Kernavė</td><td>Ledai-179, 276.5 m (1); Likėnai-398, 297.4 m (1)</td><td>Dvoriki-2, 1009.2 m (20), 1011.4 m (1); Šaravai-427, 308.8 m (1)</td><td>Drūkšiai-51, 177.0 m (1), 178.8 m (1), 194.4 m (1); Dvoriki-2, 1006.8 (1); Kunkojai-12, 529.0 m (1)</td><td>Drūkšiai-51, 178.2 m (1)</td></tr>
<tr><td>Ledai</td><td>Ignalina-332, 192.1 m (1); Kunkojai-12, 591.2 m (1); Riešutynė-321, 208.2 m (1), 252.6 m (1); Rimšėnai-331, 200.7 m (1)</td><td></td><td>Kunkojai-12, 557.15 m (1)</td><td></td></tr>
</table>

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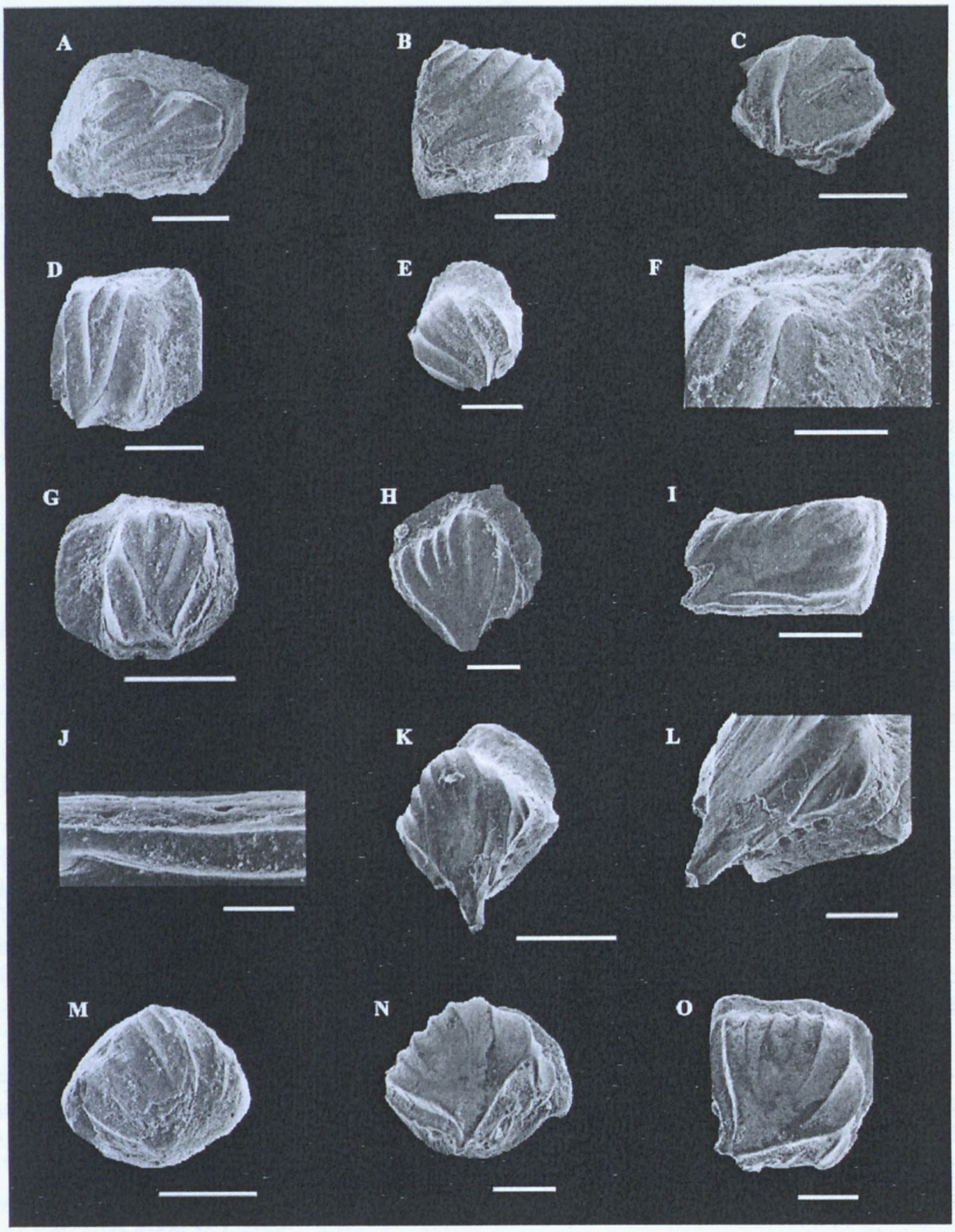
Plate 1.



A–B – *Cheirolepis,* type A. Riešutynė-321 borehole. Middle Devonian, Eifelian, the Ledai Formation. A – LGI 25-P/58, depth 252.6 m; B – LGI 25-P/59, depth 208.2 m.

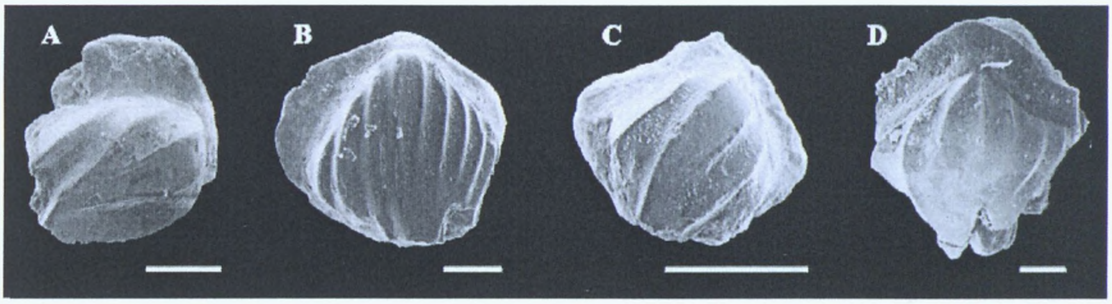
C–G, M – *Cheirolepis*, type B. Dvoriki-2 borehole, depth 1009.2 m. Middle Devonian, Eifelian, the Kernavė Formation. **C** – LGI 25-P/57; **D, F** – LGI 25-P/55; **E** – LGI 25-P/54; **G** – LGI 25-P/56; **M** – LGI 25-P/52. Stačiūnai-8 borehole, depth 349.3 m. Middle Devonian, Givetian, Upninkai Group.

H, K–L, N–O – *Cheirolepis* type C. **H** – LGI 25-P/40. Kunkojai-12 borehole, depth 529.0 m; **K–L** – LGI 25-P/62. Drūkšiai-51 borehole, depth 178.2 m. Middle Devonian, Eifelian, the Kernavė Formation. **N–O** – LGI 25-P/53. Borehole Stačiūnai-8, depth 349.3 m. Middle Devonian, Givetian, the Upninkai Group.

I–J – *Cheirolepis* type D. LGI 25 P/61. Drūkšiai-51 borehole, depth 178.2 m. Middle Devonian, Eifelian, the Kernavė Formation.

Bar: A–E, G, I, K, M–O – 200 μm; F, H, L – 100 μm; J – 50 μm. Plate 2 description.

Plate 2.



A, C – *Cheirolepis*, type B. **A** – LGI 25-P/51. Stačiūnai-8 borehole, depth 349.3 m. Middle Devonian, Givetian, the Upninkai Group; **C** – LGI 25-P/60. Šešuvis-11 borehole, depth 525.3 m. Upper Devonian, Frasnian, the Šventoji Formation.

B, D – *Cheirolepis*, type C. **B** – LGI 25-P/41. Utena-56p borehole, depth 209.0 m. Middle Devonian, Givetian, the Upninkai Group. **D** – LGI 25-P/39. Šešuvis-11 borehole, depth 525.3 m. Upper Devonian, Frasnian, the Šventoji Group.

Bar: A, C – 200 μm; B, D – 100 μm.

DIATOMS IN THE BIOSTRATIGRAPHY OF PLEISTOCENE SEDIMENTS IN LITHUANIA

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Diatom flora from different age interglacial sediments was studied to investigate the possibilities to use those data for the biostratigraphy.

The Lower Pleistocene (Turgeliai interglacial ~ 472–502 thousand years) diatom assemblage consist of about 165 taxa. The *Aulacoseira* Thw. and *Cyclotella* (Kutz.) Breb. genus species prevailed in the plankton. Periphyton species are represented by genus *Fragilaria* Lyng., *Gyrosigma* Hass., *Navicula* Bory, *Diploneis* Ehr., *Cymbella* Ag. and others. The *Aulacoseira* genus species have some characteristic morphological structure features: the frustules are thick-walled with coarse, large pores and dots, with contours of irregular form. The same morphological features are characteristic to resting spores of this genus, which predominated in some phases of Turgeliai interglacial. The high frequencies of spores could also serve as an age indicator for the assemblage, on account of the high numbers of spores encountered in the Late Pliocene and Early Pleistocene sediments. *Aulacoseira* genus forms curved along the vertical axis, the present occurrence of those is limited are very common in this assemblage. The species *Cyclotella radiosa var. lichvinensis* (Jous) Log, which became extinct in the Middle Pleistocene, has been noted as well.

It could be pointed, that the Turgeliai interglacial diatom complex has some archaic features, however the additional studies of this interglacial sediments are necessary for establishment of more reliable age criteria.

Turgeliai interglacial diatom assemblage is similar to diatom assemblage of Belovezh interglacial (Lower Pleistocene, Krasnaya Dubrova-13A borehole) in Belarus. However, these both assemblages are not absolutely identical (Khursevich et al., 1990).

More representative diatom assemblage consisting of about 125 species was found in Butėnai (Holsteinian) interglacial sediments. It is characterised by the diversity and abundance of plankton representatives of *Cyclotella* (Kutz.) Breb. genus, preferring transparent oligotrophic and mesotrophic waters. A group of extinct Pleistocene species was established as well. The Pleistocene relicts *Cyclotella radiosa* var. *pliocaenica* Krasske, *C. radiosa var. lichvinensis* achieved its flourishing during Butenai interglacial and became extinct in its end. *Stephanodiscus niagarae* var. *insuetus* Churs et Log. appeared and developed in the mass during this interglacial. Those species can be considered as biostratigraphically valuable markers. The group of Pliocene–Pleistocene exotic diatoms: *Cyclotella schumannii* (Grun.) Håkansson, *C. vorticosa* A. Breg., *Stephanodiscus niagarae* Her., *Navicula hasta* Pant., *N. cuspidata* Kutz., *Aulacoseira granulata f. curvata*

(Grun.) Hust., *A. italica f. curvata* (Pant.) Hust., *Aulacoseira ambigua f. curvata*, which are found only rarely in recent lakes, has been noted as well as some species with archaic features in a structure of frustule.

The most abundant diatom assemblage consisting of about 210 taxa was established in Merkinė (Eemian) interglacial sediments. Diatom composition of this assemblage is similar to recent one. Extinct species are absent and only a few relict species were identified. No morphological features in structure of frustules distinguishing them from recent ones were noticed.

The diatom flora of Butenai and Merkine interglacials is similar to the coeval diatom flora, established in sediments from Belarus, Russia, Poland, Latvia and Sweden (Loginova 1982; Marciniak 1984; Ambrosiani et al., 1998; Šeirienė 1998; Khursevich 1999).

The Nemunas (Weichselian) period diatom flora complex includes 173 taxa. Most of them are freshwater, alkaliphilous species. Planktonic species are prevailing by *Aulacoseira*, *Cyclotella* and *Stephanodiscus* Ehr. genera. Representatives of *Fragilaria*, *Achnanthes* Bory, *Eunotia* Ehr. and *Cymbella* genera are dominated among epiphytic species and *Navicula* – among benthonic species.

Diatom flora composition of different stadials and interstadials reflects the similar conditions of sedimentation. The cold (stadial) periods are characterized by a decrease of diatom concentration. During the Nm 1b stadial number of cold water species such as *Pinnularia borealis*, *Navicula scutelloides* increased. However, during the Nemunas 1a stadial warm water species *Cymbella ehrenbergii* prevailed.

Weichselian diatom flora is very similar to Eemian interglacial diatom flora. The diatom zones distinguished not always correspond with pollen zones. The most reliable is establishing of J1, J2 interstadials and Nm 1b stadial. The boundary between Nemunas (Weichselian) and Merkinė (Eemian) periods is indistinct in diatom diagrams.

It can be summarised that interglacial diatom assemblages of different age possess definite diagnostic features and can be used for the biostratigraphy and correlation of sediments. Nevertheless the biostratigraphy of lacustrine sediments is complicated, because of different size and depth of the lakes, determining the coeval diatom flora differences. The best results can be obtained by correlation of assemblages from the same facies and from the deepest part of the palaeobasin.

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CONODONT ASSEMBLAGES OF THE GĖLUVA (LOWER SILURIAN) REGIONAL STAGE

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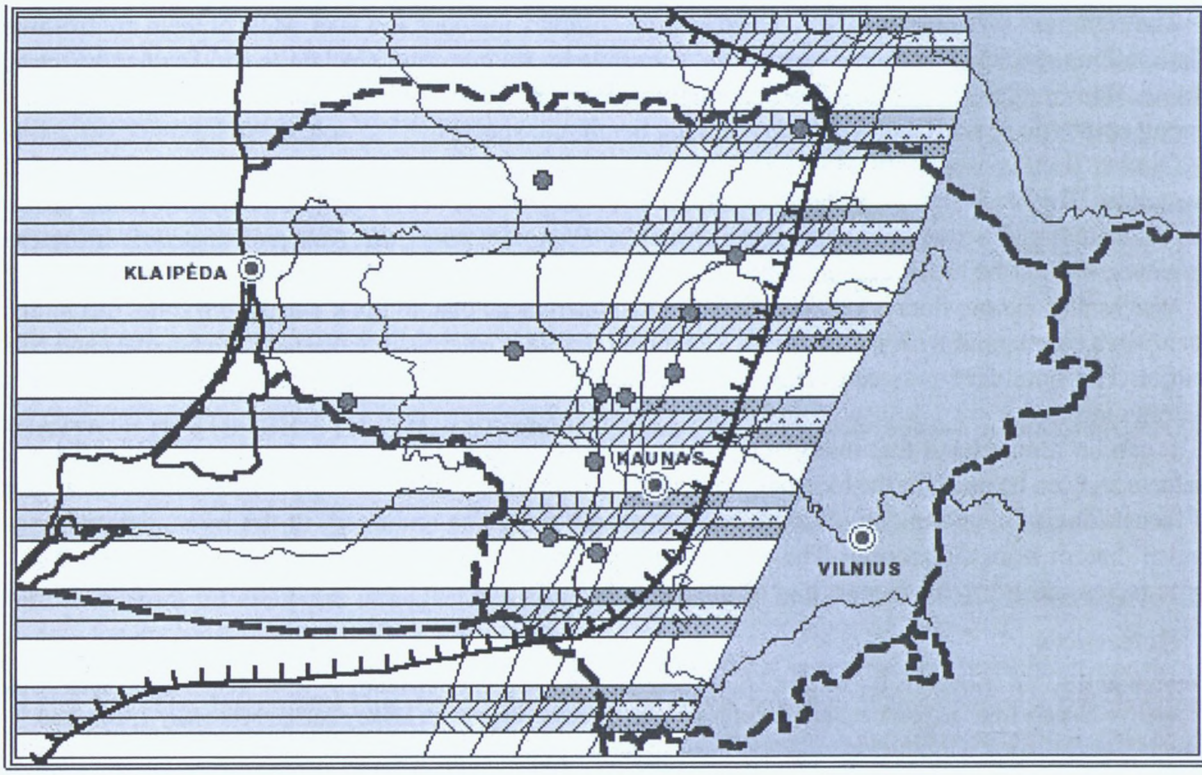
According to A. Brazauskas and P. Musteikis publications (Brazauskas, 1985, 1993; Musteikis, Paškevičius, 1999), the quantitative data of studied fauna to establish the conodont assemblages have been used. The amount of taxa in the sample have been calculated in per cent.

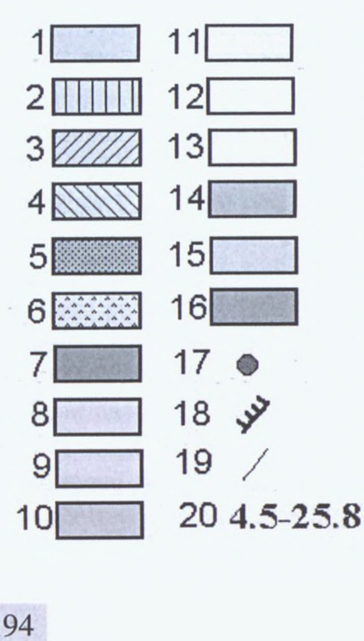
Based on the cluster analysis method, each section of 12 studied boreholes was divided into several units. Further the average values of each taxon were determined for each unit. These values have defined the conodont assemblages. Assemblages were named by one to three dominating species.

There were established 5 conodont assemblages and 1 assemblage was reconstructed in the sedimentary basin (Fig. 1): *Dapsilodus obliquicostatus-Ozarkodina e. excavata-Panderodus* sp. E, *Pseudooneotodus bicornis – Panderodus equicostatus, Panderodus equicostatus, Ozarkodina e. excavata – Panderodus equicostatus, Ozarkodina e. excavata – O. confluens* ir *Ctenognathodus* sp. A.

In Lithuania 5 lateral areas of assemblages have been distinguished. The eastern distribution area of assemblages was restored. Perhaps, this area has contained *Ozarkodina e. excavata – O. confluens* assemblage in the lower part of the section and *Ctenognathodus* sp. A assemblage in the upper one. These assemblages have been restored taking into account regularities of the lateral and vertical changes of sedimental environment.



A						
T	20.7-74.3	32.8	45.8-66.1	55.7-71.3	32.3-41.3	>60
T	15-29.3	30.6	15.7-36.2	11.8-23.0	6-27.9	<40
T	3.5-19.4	29.6	1-7.5	7-9.2	4-10.1	?



Conodont assemblages:

1 - Dapsilodus obliquicostatus - Ozarkodina e. excavata - Panderodus sp. E, 2 - Pseudooneotodus bicornis - Panderodus equicostatus, 3 - Panderodus equicostatus, 4 - Ozarkodina e. excavata - Panderodus equicostatus, 5 - Ozarkodina e .excavata - O. confluens, 6 - Ctenognathodus sp. A.

Predominant taxa of conodont assemblages:

7 - Dapsilodus obliquicostatus, 8 - Panderodus sp.n.E, 9 - Ozarkodina e. excavata, 10 - Pseudooneotodus bicornis, 11 - Panderodus equidendatus, 12 - Ozarkodina confluens, 13 - Ozarkodina bohemica ssp.P, 14 - Ozarkodina svetlanae, 15 - Ctenognathodus sp.A.

Other signs:

16 - denudation area, 17 - applied boreholes, 18 - boundary of the present distribution of the Gėluva Regional Stage, 19 - boundaries of conodont assemblages, 20 - value of taxon in per-cent in the assemblage.

Conodont assemblages stratigraphically overlap in separate sections, therefore the overlaping areas of 2 assemblages were reconstructed. The description of assemblages according to their distributional pattern is taken toward the shallowing direction of sedimentary basin. The names of the assemblages are offered for the first time, unless otherwise stated.

The *Dapsilodus obliquicostatus – Ozarkodina e. excavata – Panderodus* sp. E assemblage.

Composition: Polytaxonical assemblage; the dominating species in distinct samples reach 80–95%. Sometimes *Dapsilodus obliquicostatus* (Branson et Mehl) – 20.7–74.3%, *Ozarkodina e. excavata* – 15–29.6%, *Panderodus* sp. E – 3.5–19.4%.

Age: the Latest Wenlock – the Earliest Ludlow.

Geographical distribution: Western and Northwest Lithuania.

The *Pseudooneotodus bicornis – Panderodus equicostatus* assemblage.

Composition: Polytaxonical assemblage (6 species). In the older assemblages of other sections, the quantitative content of taxa slightly differs, and per cental value of predominant species is presented basing on the Sutkai-87 reference borehole. Here, *Pseudooneotodus bicornis* (Drygant) makes 32.8%, *Panderodus equicostatus* (Branson et Mehl) – 30.6%, *Ozarkodina e. excavata* – 29.6%.

Age: the Early and Early Late Wenlock.

Geographical distribution: Western Lithuania.

The *Panderodus equicostatus* assemblage.

Composition: Polytaxonical assemblage (to 10 species). Dominating taxa: *Panderodus equicostatus* – 45.8–66.1%, *Ozarkodina e. excavata* – 15,7 – 36,2%.

Age: Late Wenlock.

Geographical distribution: Middle Lithuania.

The *Ozarkodina e. excavata – Panderodus equicostatus* assemblage.

Composition: Polytaxonical assemblage (10 taxa). Dominating species: *Ozarkodina e.excavata* – 55.7–71.3%, *Panderodus equicostatus* – 11.8–23.0%.

Age: the Latest Wenlock – the Earliest Ludlow.

Geographical distribution: Middle Lithuania.

The *Ozarkodina e. excavata – O. confluens* assemblage (the name of assemblage is offered by V. Viira, 1982).

The assemblage is described by R. Aldridge & L. Jeppsson (1999). The complete characteristics basing on data from our region is as follows.

Composition: Polytaxonical assemblage (to 8 species). Dominating taxa: *Ozarkodina e. excavata* – 32.3–41.3%, *O. confluens* (Branson et Mehl) – 6.0–27.9%, *Panderodus equicostatus* – 4.0–10.1%, *Oulodus ziegleri* (Walliser) – 10.4%, *Ozarkodina bohemica* ssp. P to 6.7%.

Age: the Latest Wenlock.

Geographical distribution: Middle Lithuania.

The *Ctenognathodus* sp. A assemblage.

We can consider this assemblage as monotaxon. The assemblage consists of rare single specimens can be described as *Ctenognathodus* sp. A. Quite rare separate elements of *Ozarkodina confluens* apparatus are also found. The per cental amount is less than 40.0%.

The possible age is Latest Wenlock – Earliest Ludlow.

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PLEISTOCENE STRATIGRAPHY OF EAST LITHUANIA

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Pleistocene sediment sequence in East Lithuania is characterised by the dense net of boreholes. A number of them were used for the geological mapping at a large scale and have a core description of a high quality with detailed lithological characteristics. Together with the lithological characteristics of deposits, the results of plant fossil investigations and the analysis of spatial occurrence of beds they leaded the compilation of digital spatial (3D) model of Pleistocene sequence. The compilation of such digital model actually is the stratigraphical interpretation of the deposit structure. The model is based on the data basis of bed spatial occurrence obtained from deposit stratigraphy. The results of litho- and biostratigraphy and the geological information on entire complex of erosion and accumulation processes served as main criterions for its compilation. After the preliminary model was compiled, it was tested under the results of palaeogeomorphological reconstruction carried out for the main interglacials. The compilation of the vertical geological profiles of all the possible directions, the schemes and block-diagrams of bed spread, analysis of bed surfaces and deposit parameter distribution was used.

Till complexes of various age are notable for their different content of long distance transported crystalline rocks from Scandinavia, local Palaeozoic rocks (limestone, dolomite) and short distance transported Mesozoic rocks (marlstone, siltstone, sandstone, chalk). The till beds of the earliest Pleistocene glaciations are usually composed of a grey or brownish grey diamicton with a similar pebble composition. A greyish brown older till bed has, however, been observed in restricted areas in sheltered position in the bedrock depressions. The thickest sequences of glaciofluvial and glaciolacustrine deposits of this complex are associated with its uppermost part and marks the time of the glacial melting just before the forthcoming Butėnai (Holsteinian) interglacial.

The deposits of Butėnai (Holsteinian) interglacial are comparatively well presented in East Lithuania. The lake-margin sandy deposits, deltaic lithofacies and lacustrine bottomsets can be distinguished in a large palaeobasins of this age indicated by pollen grain spectra from silt and clay beds. However the best palaeobotanical evidence is obtained from the organic rich sediment sequences of small lakes or bogs.

The largest middle part of the Pleistocene deposit sequence is composed of a very hard, brown diamicton of till beds with very similar physical properties and composition. The deposits formed by ice-melt water are mostly associated with the uppermost part of the complex, however the till beds are numerous. The question of subdivision of this glacial lithocomplex usually is related with the idea of the existence of Snaigupėlė (Drente–Warthe) interglacial. Some interglacial deposit sequences according to the results of pollen analysis are attributed to this time interval, however the lithostratigraphical position of these interglacial deposits and other investigation results rise doubt on its position.

Lake sediment sequence of Merkinė (Eemian) interglacial comprises a big variety of lithofacies including siliciclastic sediments of various grain-size, silty clay, lake marl as well as organogenic sediments with abundant mollusc shells. In south-eastern Lithuania these interglacial deposits occur on present day surface in the depressions of the peat bogs. However the rest territory is cowered by the Late Glacial till beds composed of reddish brown and yellowish grey diamicton forming the relief together with the deposits of ice melt water.

STRATIGRAPHY AND LITHOFACIES OF VENDIAN, SOUTHEASTERN LITHUANIA

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The Vendian deposits represent the basal portion of the sedimentary cover of the Baltic basin. The sedimentation was restricted to the eastern part of Lithuania and further east relating to the post-rift widening of rift-related structures (Volynian rift). Due to particular structural and sedimentation features the Vendian–Earliest Cambrian (Baltic Series) succession is attributed to the Baikalian structural stage. It was characterised by relative stability of local structures that were identified in south-eastern Lithuania on the basis of thickness and lithofacies variations. The thickness of the Vendian deposits reach 170 m in the east.

Data of more than 100 wells were inspected to map lithofacies distribution of different stratigraphical units of Vendian in the southeastern Lithuania. An emphasis has been made to standard logging data, because local stratigraphy of the Vendian deposits is based on lithostratigraphic criteria. Various authors presented different interpretation of the lithofacies variations that resulted in a non-unique stratigraphical subdivision of well sections. The presented study allowed unification of the stratigraphical subdivision of the southeastern Lithuania.

Five formations, i. e. the Merkys, Jašiūnai, Rūdininkai, Skynimai, Vilkiškės Fms. are defined up the Vendian succession of Lithuania. They compose distinct sedimentation cycles. The sedimentation trend indicates gradual levelling of the relief and increasing degree of the chemical weathering of the provenance rocks. This has associated with gradual cessation of the local tectonic movements as it is inferred from lower differentiation of thickness in the Vendian basin.

The lithofacies and isopach maps of each formation were compiled that revealed sourcing and sedimentation trends as well as the sin-sedimentary tectonic activity of local structures. The latter show rather persistent character. Three first-order structures were documented in the study area, i. e. the Žiežmariai–Maišiagala and Merkys depressions separated by the South Lithuanian elevation striking west-east (Fig. 1). Against this background structural pattern the smaller-scale structures were recognised that also affected thickness and lithofacial variations (Fig. 1), the latter pointing to syn-depositional activity of structures. The South Lithuanian uplift is characterised by deposition of the coarser grained sediments compared to the adjacent Žiežmariai–Maišiagala and Merkys depressions. Similarly, the Šalčininkėliai, Šalčininkai, Pabarė, Kalesnikai, Barčiai, Varėna structures are well defined. Majority of these structures were controlled by north-south and west-east striking faults reactivated in the crystalline basement. The diagonal fault family was of much less activity, except scarce NE–SW-oriented faults that frame narrow (few kilometres width) graben-like depressions established during the Volynian time (Fig. 1). The small-scale Kernovelė drape structure was documented in SE Lithuania, which is composed of the basement block uplifted as high as 80 m (Fig. 1). This block is overlain by upper Rūdininkai–Vilkiškės sediments showing formation of the uplift prior sedimentation. The South Lithuanian elevation is separated from the Žiežmariai–Maišiagala depression by W–E trending Birštonas–Vilnius fault of the crystalline basement. In the south the uplift is controlled by the Alytus fault. The Lower Merkys depression and Upper Merkys depression are separated by the Rūdiškės fault of the north-south orientation.

Despite the persistency of majority of defined structures, some evolutionary trends are discernible. A partial inversion of some Merkys–Rūdininkai structures has been stated during the Skynimai–Vilkiškės time that was preceded by break in sedimentation.

The lithofacies distribution is sensitive to the structural grain of the basin bottom (Fig. 1). The lithofacies distribution reflects not only the local features, but also are indicative of activity of local uplifts located beyond the basin that were eroded during Vendian time. A discordant trend of lithofacies belts with respect to thickness changes pursued a presence of uplifted structure further south that can be related to the Grodno–Mosty basement uplift in the north-western Belarus.

The Merkys Fm. is represented by dark-brown quartz-feldspar (Q–Fsp) conglomerates, sandstones, siltstones and rare claysontones. The two latter are abundant in the Žiežmariai depression in the northeast (Fig. 1). The lithosfacies distribution shows rather strong differentiation controlled by basin bottom morphology. The thickness ranges from a few dozens of centimetres to 22–47 m in the Zavišonys graben-like structure in the south-east (Fig. 1).

The Jašiūnai Fm. is dominated by Q–Fsp conglomerates that pass to coarse-grained sandstones in the northeast and east. The Žiežmariai–Maišiagala and Upper Merkys depressions are separated by west-east trending Vokė and Taučionys elevations. The thickness attains 65 m in the east.

The overlying Rūdininkai Fm. is composed by brown Q–Fsp conglomerates in the southwest, they give way to coarse-grained feldspar-quartz sandstones cemented by clay and iron hydroxides in the northeast and east. The thickness exceeds 50 m in the southeast.

The Skynimai Fm. marks significant changes in the sedimentation. The brown coarse and fine-grained sandstones dominate the section. The siltstone and claysonte layers grow in abundance eastwards. Conglomerates prevail in the western periphery of the Vendian basin. The thickness attains 50 m in the east.

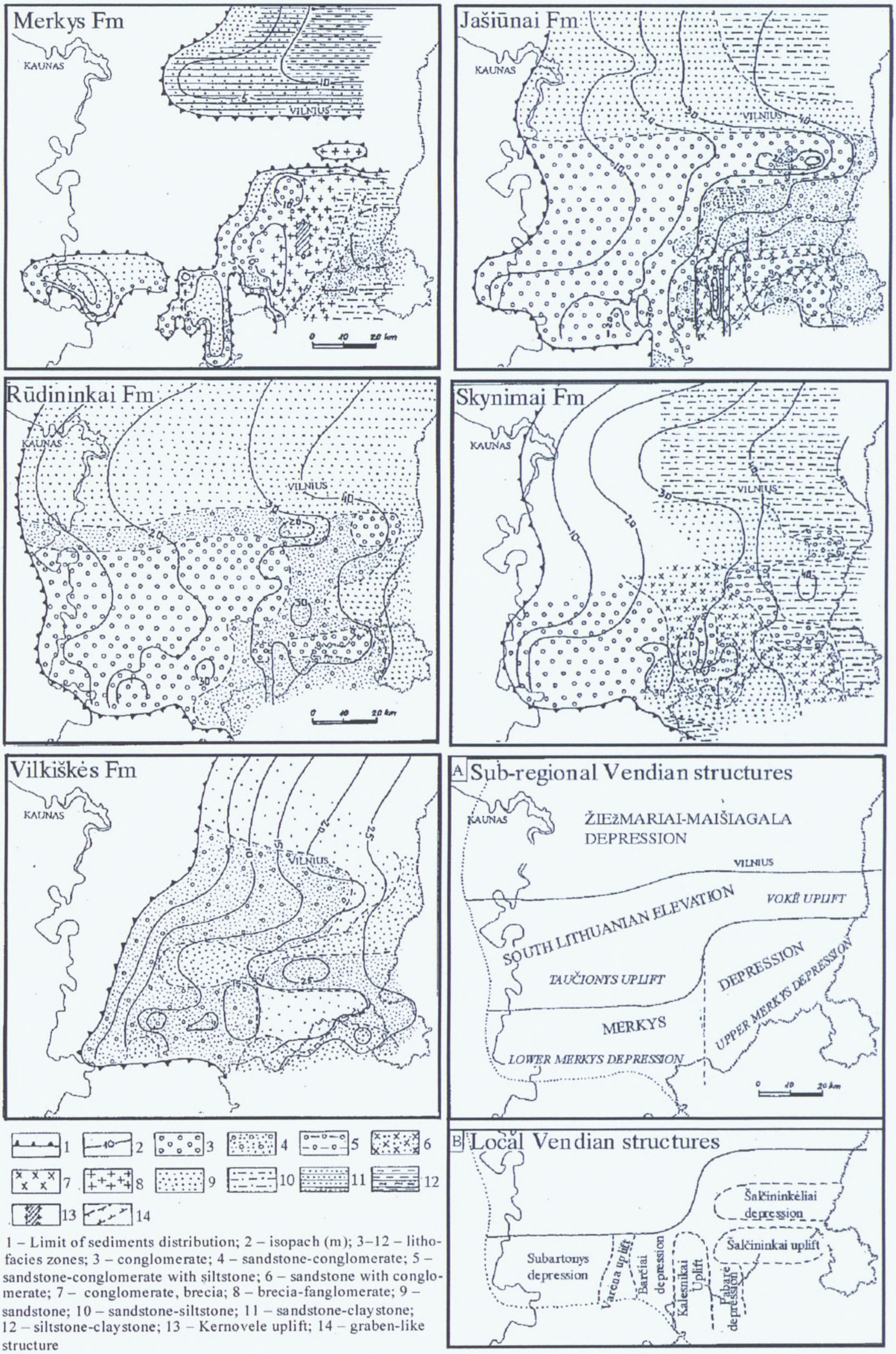

Fig. 1. The lithofacies and thickness distribution of the Merkys, Jašiūnai, Rūdininkai, Skynimai, Vilkiškės formations (Vendian) in SE Lithuania. The lower right hand figure shows sinsedimentary tectonic structures of sub-regional (A) and local (B) order.

The Vilkiškės Fm. deposited after some break in sedimentation as it is indicated by 0.2–2 m thick zone of weathering at the top of he Skynimai Fm. A protracted break in sedimentation is also evidenced by considerable change in deposits composition compared to that of the underlying Vendian rocks. The Vilkiškės Fm. deposits are characterised by high maturity. The quartz sandstones prevail showing fining to the east, where feldspar occurs at increased amount. The thickness is up to 27 m.

GEOMETRY OF MIDDLE PROTEROZOIC ANOROGENIC INTRUSIONS OF THE BALTIC AREA: IMPLICATIONS FROM 2¾D MODELLING OF POTENTIAL FIELDS

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INTRODUCION

The crystalline basement of the Baltic region was consolidated during Early–Middle Proterozoic and is represented by igneous and high-grade metamorphic rocks. The cratonization of the crust has finished with intrusion of numerous anorogenic bodies at ca. 1.6–1.5 Ma that show different composition and geometries. The shape and tectonic control of the intrusions in plan view are reflected in gravity and magnetic fields, combined with the scarce wells. The extent and geometry to the depth as well as internal structure of the intrusions were modelled by means of $2^3/_4$D modeling program of the potential fields. A group of different-scale bodies were studied in the south-western Lithuania (Lazdijai area), the regional-scale profile crossed the Riga massif the largest intrusion in the Baltic region.

LAZDIJAI AREA

Five anorogenic bodies, referred to as the Veisiejai Complex, were mapped in SW Lithuania (Fig. 1). The composition ranges from diorite to granite. They are compatible to the Mazury rapakivi-like granites which constitute the wide W–E-oriented belt breaking its way along the north-eastern border of Poland. The Veisiejai intrusions represent the eastern extreme of this belt. They are of increased density and high magnetic susceptibility causing distinct gravity and magnetic anomalies (Fig. 1). A high contrast of density and magnetic properties of the granidoids to hosting rocks makes the modelling consistent. Four profiles were modelled.

Profile I. It strikes NW–SE across the granodiorite Veisiejai massif of W–E elongated shape (Fig. 2). Following the modelling, the thickness of the body is 3 km (Fig. 3).

Profile II. It crosses the Lazdijai and Akmena massifs. Lithological and petrophysical characteristics are close to the Veisiejai massif. Modelling reveals some internal differentiation of the bodies. The well No. 6 has penetrated the upper more felsic portion of the Lazdijai massif. To explain the high-intensity magnetic anomaly, more magnetised granodiorites are suggested at a depth (Fig. 3). The intrusion is of cone shape.

Profile III. The modelling was run along the line west-east crossing the Akmena and Viesiejai intrusions. It shows that they constitute a common body with two centres. Most likely the granitoids intruded west-east striking shear zone. The shape of the intrusion suggests dextral strike-sleep of the controlling W–E oriented fault.

Profile IV. The profile IV strikes across the large Kapčiamiestis intrusion of (Fig. 2). It shows strongly differentiated lithologies ranging from dominating granodiorites in its periphery to prevailing granites in the centre. Petrophysical and lithological composition of the peripheral part of the massif resemble the afore-described Veisiejai Complex bodies. In the centre, the granites prevail. The petrophysical boundaries are inclined to the centre. The intrusion is as of a tabular shape, it is as thick as 4–5 km.

RIGA MASSIF

The Riga massif is the largest igneous body identified in the Baltic region. It is of 200–250 km diameter. Rapakivi granites subcrops in the north, they grade to monconites it the centre. Gabbro-anorthozites subcrop in the southern periphery of the massif. The rocks are dated ca. 1.58 Ma.

The gravity and magnetic fields are also variable within the Riga massif. The northern mafic belt associates with high anomaly, while centre shows gravimagnetic low. The limits of the intrusion are well traced on the

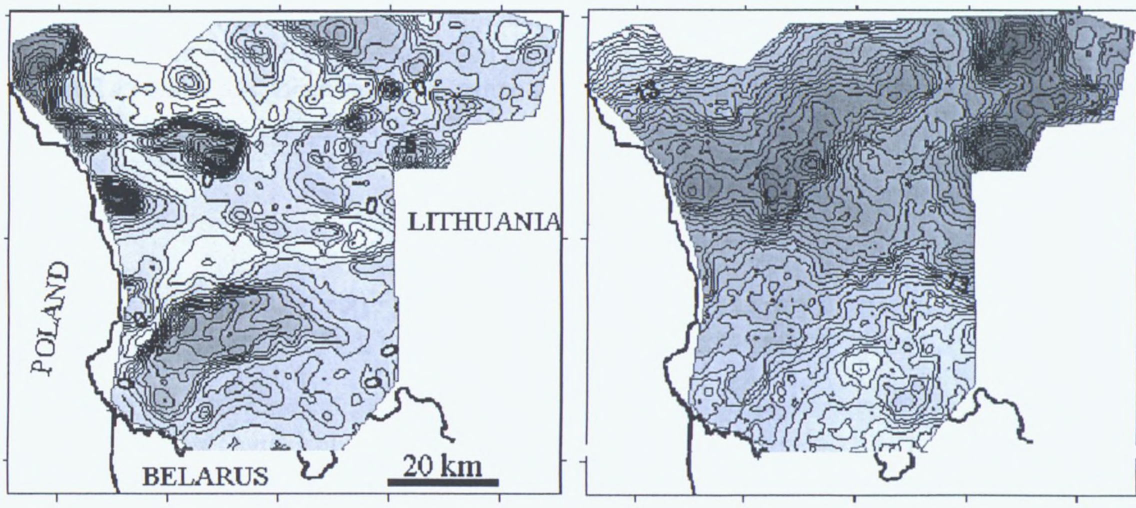
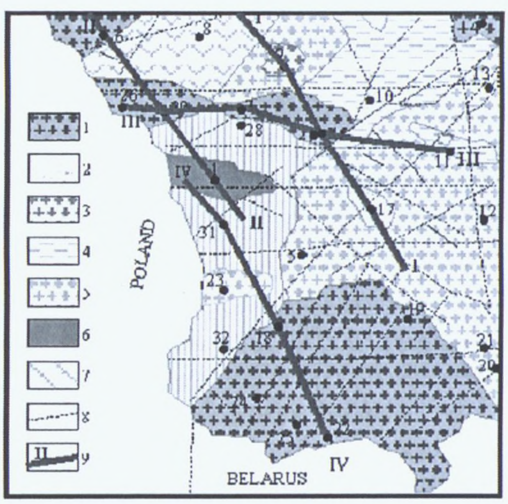


Fig. 1. Magnetic (left) and Bouger anomalies (right) maps of SW Lithuania.



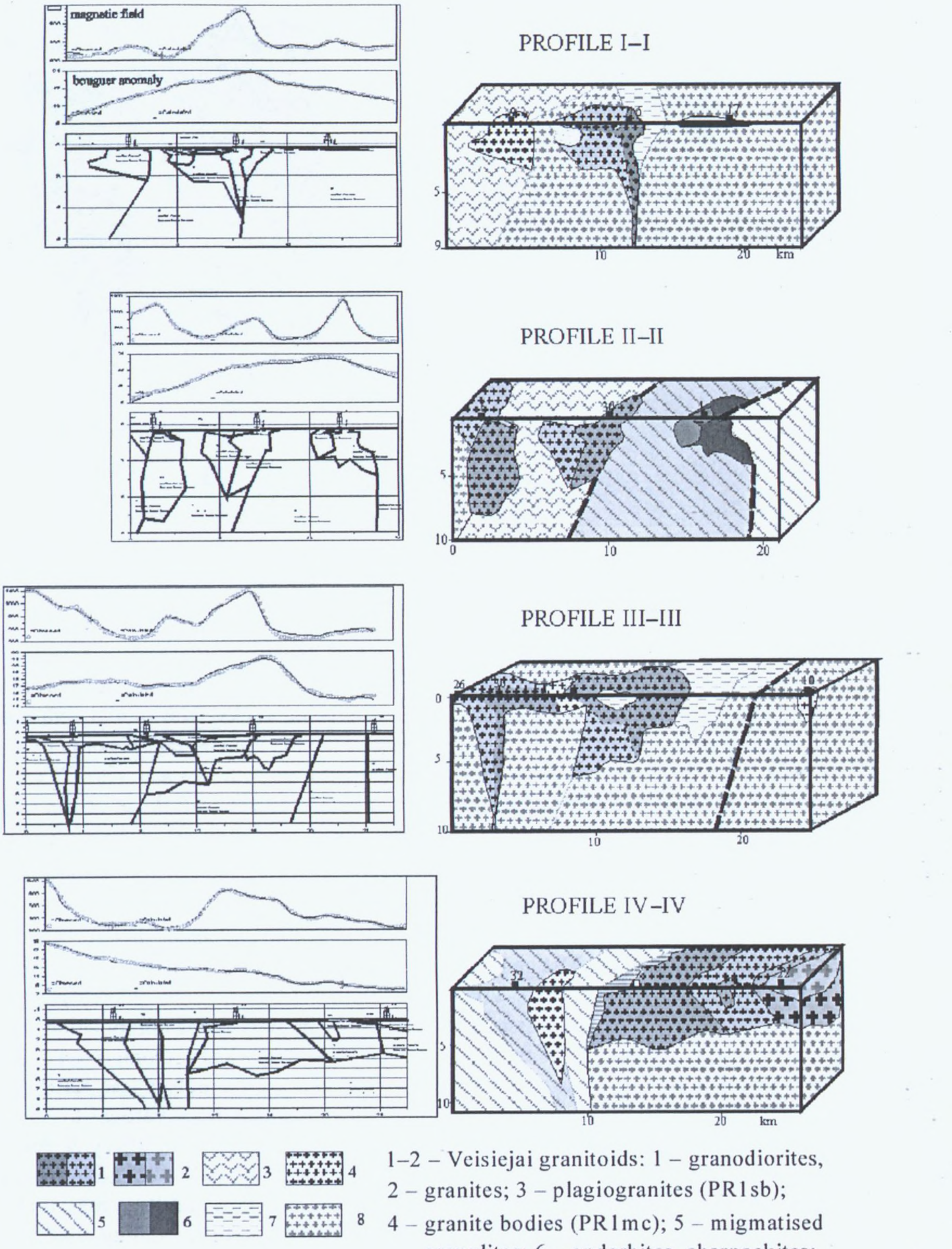
1 – Veisiejai Granitoid Compex (PR1vs)
2 – plagiogranites of Subartonys Compex (PR1sb)
3 – granites of Marcinkonys Compex (PR1mc)
4 – supracrustal amphibolites, gneisses of Balninkai Complex (PR1bl)
5 – migmatites, granites with relicts of (4) (PR1mc)
6 – enderbites, charnockites (Kuciunai block)
7 – migmatites, granites with relicts of granulites (PR1mc)
8 – fault
9 – line of modelled profile

Fig. 2. Geological map of Lazdijai area, SW Lithuania (according to Marfin et al., 1994).

magnetic and gravity maps, except the western part that does not show any distinct geophysical features (Fig. 4). It can be interpreted in terms of plunging of the intrusion to the west under the hosting basement rocks.

The gravity and magnetic fields were modelled along the 400 km long NNE–SSW profile. Modelling of the geological bodies based on potential field and drilling data only is not consistent, the backward modelling is invariant. The main petrophysical crustal boundaries were calibrated according to the Sovietks–Kohtla-Jarve DSS profile that parallels the modelled profile.

The geothermal field provides an important information on the crustal lithologies. The Riga massif associates with the 30 mW/m^2 heat flow minimum (lowest in the Baltic basin), showing depletion of the crustal lithologies in K, Th, U, that goes into conflict to high heat production potential of the rapakivy granites. For example, assuming 3 mW/m^3 heat production and 2 km thickness of the Riga granites and 12 mW/m^3 of the mantle heat flow, it leaves only 12 mW/m^3 for the rest 40 km thick earth's crust part, implying very low average heat production 0.3 mW/m^3. Thus, heat flow data provides rather strict limitations for the modelling of the potential fields. Firstly, the granitic part of the intrusion could not be thicker than 6 km, but in this case the rest crust is represented by mafic rocks that would lead to abnormal gravity anomaly. Contrary, we observe the gravity low compared to adjacent areas which indicates comparatively decreased density of the crustal rocks. The most likely solution of this conflict is assumption of the anorthosites comprising considerable part of the intrusion. They are of rather low density (2.71 g/cm^3),



1–2 – Veisiejai granitoids: 1 – granodiorites,
2 – granites; 3 – plagiogranites (PR1sb);
4 – granite bodies (PR1mc); 5 – migmatised granulites; 6 – enderbites, charnockites;
7 – supracrustal aphibolites, gneisses (PR1bl);
8 – migmatites, granites (PR1mc)

Fig. 3. Gravity and magnetic 2.75-D models along I–I, II–II, III–III, IV–IV profiles** (see Fig. 1. for locations) **and their geological scetches.

magnetic susceptibility (530x10^{-6} CGSM) and heat production. Still, the middle- and lower crust should show higher amount of mafic rocks than that in the surrounding areas. The Sovietks–Kohtla-Jarve DSS profile indicates anomalous thick crust (exceeding 60 km) close to the intrusion. The thermal balance makes it hardly to be believed the thick crust presented under the Riga massif.

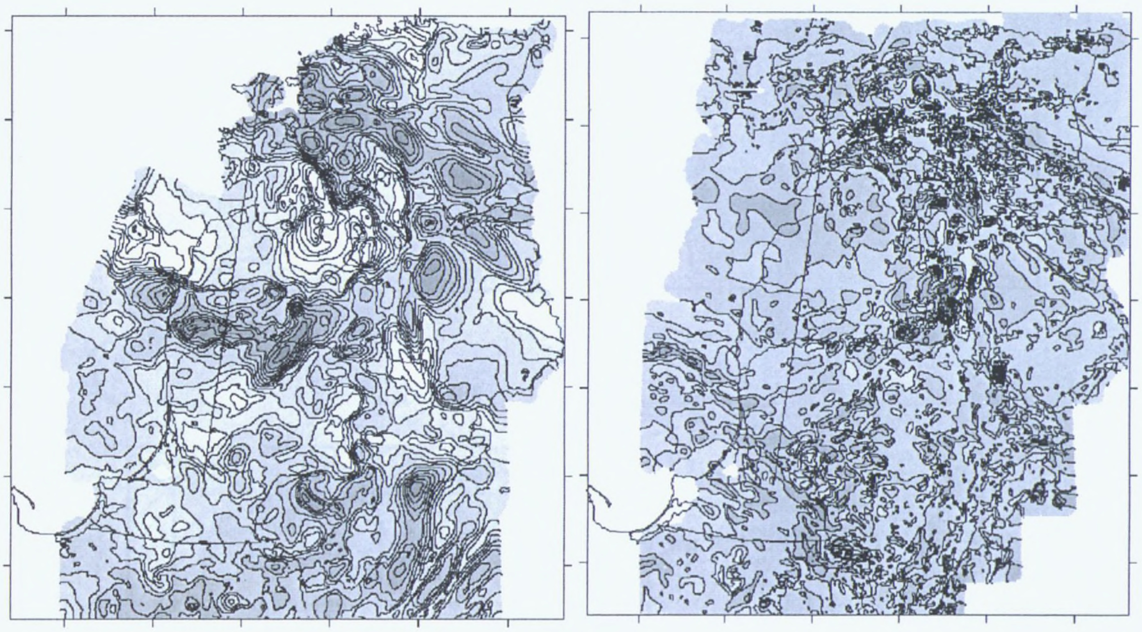


Fig. 4. Gravity (left) and magnetic (right) field of the Baltija region. The line indicates the modelled profile location.

Based on afore-described implications, the magnetic and density model was calculated to fit the observed gravity and magnetic fields (Fig. 5). Also, the Werner deconvolution operator was employed to document internal structural features of the intrusion and hosting rocks. The model indicates distinct lithological and petriophysica differentiation. The thickness of the top granitic layer is only of 2–2.5 km. It is underlain by of ca. 5 km thick anortohosites, thus, the total thickness of the intrusion is of 7 km. The mafic rocks accumulated on the flanks. The middle and lower crust are denser than those of the neighbouring blocks. The thickness of the crust increases from the massif centre (44 km) to its periphery (47 km).

DISCUSSION AND CONCLUSIONS

The modelling of the gravity and magnetic fields combined to other available information revealed two types of the anorogenic granitoid bodies, i. e. cone-shaped and tabular-shaped, that actually reflect two different kinematic types of intrusions. The geometry of the intrusions depends firstly on the interplay of the tectonic and intrusion-induced stresses (e. g. Vigneresse et al., 1999). The cone-shaped geometry of the small Veisiejai, Akmena etc. bodies is consistent to their rather homogeneous composition showing one phase intrusion. The small amount of the melt was not enough to reorient tectonic stress filling in the tectonic pathways oriented normal to the minimum stress. By contrast, the poly-phase Kapčiamiestis and Riga massifs were large enough to reorient the minimum tectonic stress from vertical to horizontal that caused the tabular shape of the intrusion. Based on empirical correlation by McCaffrey & Petford (1997) $T = 0.29 L^n$ ($n = 0.8$+-0.2, and T and L are respectively the thickness and radius of the intrusion) one can expect 4 km thickness of the Kapčiamiestis intrusion ($R = 25$ km, $n = 0.8$) and Riga massif should be as thick as of 11.5 km (assuming $n = 0.8$ and $R = 100$ km), which is close to the modeling results.

The Riga intrusion originated in the prior thickened earth's crust due to melting of the lowermost part of the crust that explains modelled thickened periphery and shallowing of the Moho in the centre of the massif. Moreover, the lower and middle crust should have been subjected to parial melting that led to depletion in Fe, K, Th, U and higher densities as it is implied from the thermal and potential field modelling data. The melt accumulated in the uppermost crust, the anorthosites concentrating in the lower half of the chamber whereas granitic material accumulated in the top part. The role of the mafic part in this process remains not clear. Judging from the reduced heat flow (30 W/m^2 against background 40 W/m^2), 3 km of the granitic top layer were eroded, thus the primary thickness of intrusion was around 10 km.

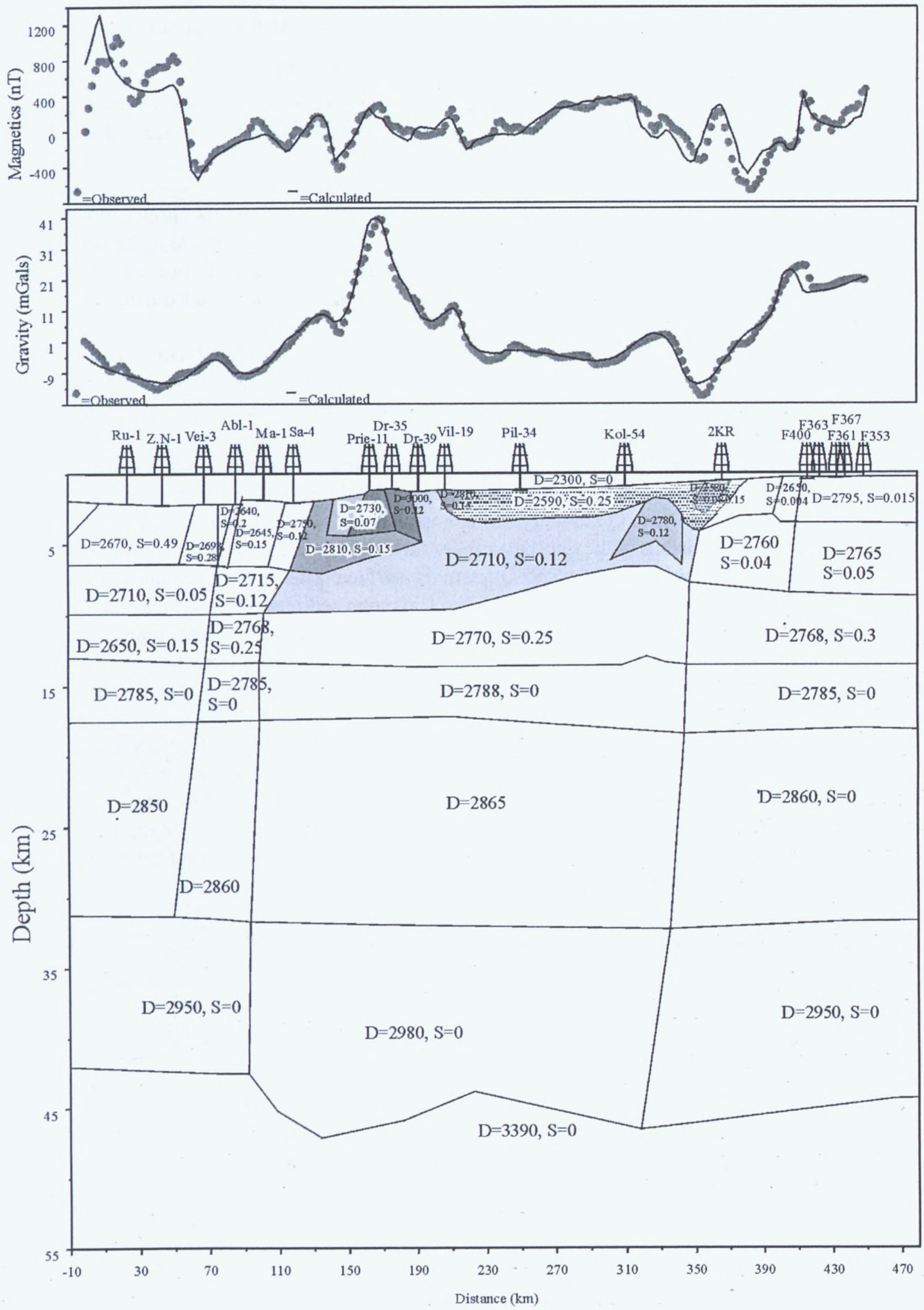


Fig. 5. Gravity and magnetic model of the crust along profile indicated in Fig. 4. The Riga massif is greyed. The doted polygon indicates granites, light grey is anorthosites and dark grey shows mafites. Density (D) and magnetic susceptivility (S) are indicated.

STRATIGRAPHY OF THE SUB-QUATERNARY SURFACE OF LITHUANIA: A NEW EDITION OF THE PRE-QUATERNARY GEOLOGICAL MAP AT A SCALE OF 1:200 000

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A set of Pre-Quaternary geological maps of Lithuania at a scale of 1:200 000 was published since 1968 to 1984. Due to the long period, the maps differ in accuracy, the different stratigraphical legends were used. A great number of new wells was drilled, the stratigraphical studies much advanced since than. Therefore, the new Pre-Quaternary Map of Lithuania, based on unified legend, was compiled during 1995–1999 in Geological Survey of Lithuania.

The digital data base "Borehole" of Geological Survey of Lithuania, containing location and stratigraphic information on 27 000 wells, was employed. This much improved the compilation of the maps. Due to rather schematic description of some hydrogeological wells (that compose the main part of the wells available in the territory of Lithuania) they were corrected according to nearby mapping wells.

A supplementary Map of the Relief of the Sub-Quaternary Surface of Lithuania at a scale of 1:200 000 was compiled. The relief is highly dissected that controls distribution of subcropping Pre-Quaternary layers. The topography the of Sub-Quaternary surface vary from 90–100 m above sea level to 30–70 m below sea level. The paleoincisions of the Sub-Quaternary surface are someplace as deep as of – 150 m b. s. l.

Strata of different age crop out on the Sub-Quaternary surface. The oldest Silurian rocks are directly overlain by Quaternary deposits east of Vilnius city. The Devonian sediments crops out in the north-eastern half of Lithuania. The truncation level increases to the south-east from Upper Famennian to Eifelian. The south-western half of Lithuania is covered by Upper Permian–Mesozoic rocks, the subcropping layers younging to the south-west. The Palaeogene and Neogene sediments are mapped in southern Lithuania as the isolated patches.

The same general tendencies were indicated in the former maps. However, a larger number of borings used allowed to define the limits of strata more precisely. Furthermore, the newly compiled Pre-Quaternary geological maps of Lithuania were compiled in digital form by means of GIS MapInfo. It allows easy access to the data and supplementation of the map with new data (e.g., of the geological mapping at a scale of 1:50 000). The maps can be easily plotted in different size and used as a background for hydrogeological, environmental and other scientific and applied maps.

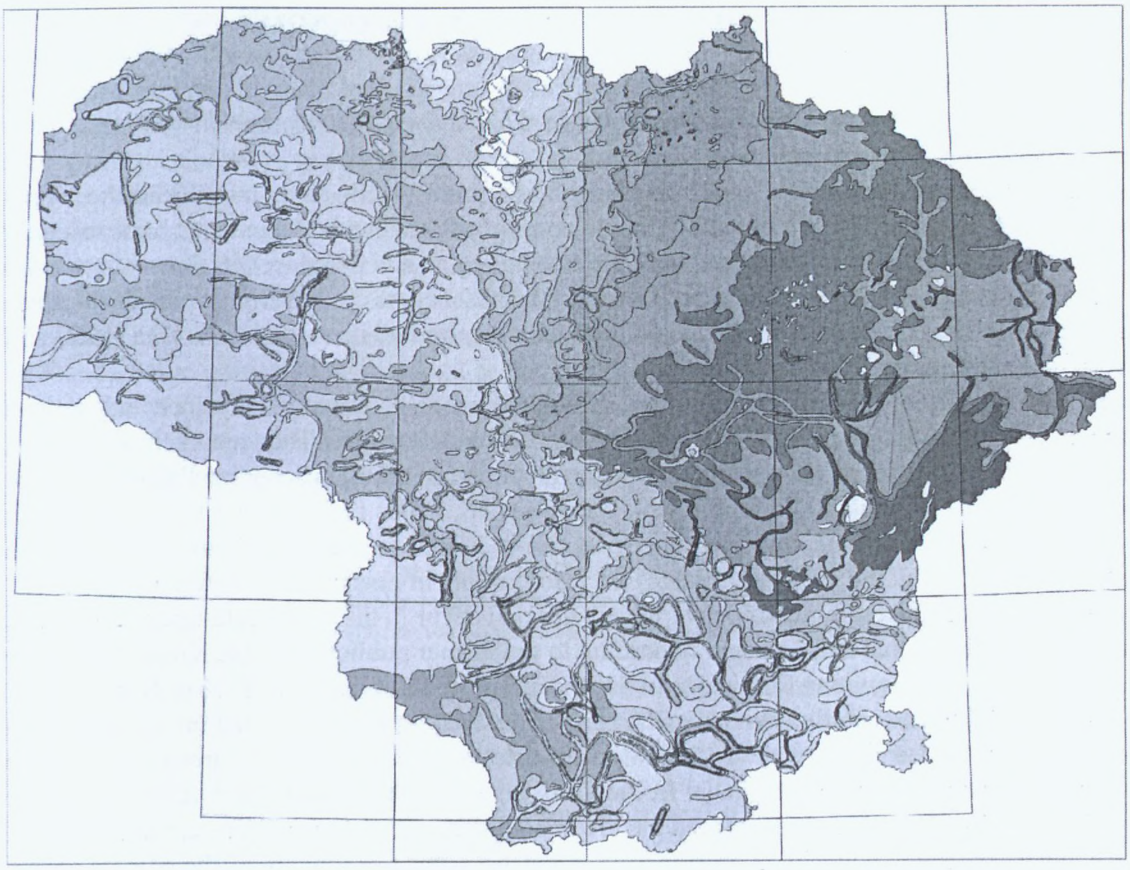

Figure. Pre-Quaternary Geological map of Lithuania (above) and map of Sub-Quaternary relief of Lithuania (bellow). Original maps are in colour.

LOWER BOUNDARY OF THE KUNDA STAGE IN BALTOSCANDIA

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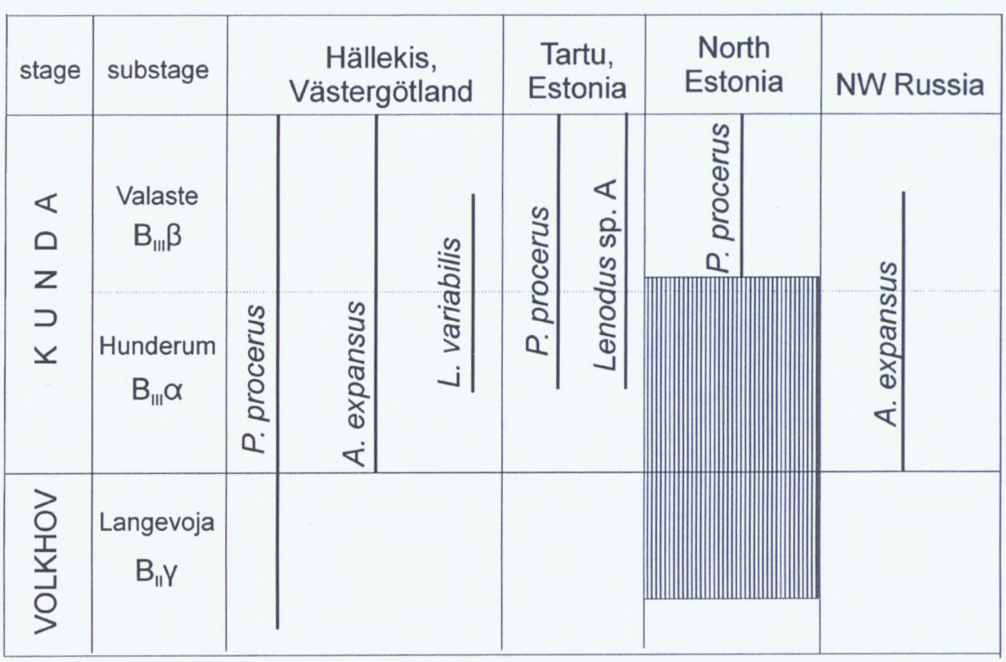
The first descriptions of the Kunda Stage was given by Schmidt (1881), who introduced the Vaginatum Limestone in NW Estonia. The present stratigraphic concept of the Kunda Stage dates back to Lamansky (1905) who defined this stratigraphic interval in St. Petersburg region and introduced its three-part subdivision, based on trilobite zonation. He was the first to assume the absence of the lowermost part of the stage, the *Asaphus expansus* Biozone, in northern Estonia. The latter viewpoint was accepted by Orviku (1960), Männil (1966), Männil and Meidla (1994) and Meidla (1997).

The lower boundary of the *Asaphus expansus* trilobite Biozone has been widely accepted as the lower boundary of the Kunda Stage. Judging from the absence of the A. expansus Biozone in North Estonia, this level is thought to mark a regressive interval in the development of the Baltoscandian Palaeobasin (Fig. 1). Recent investigations (Puura and Tuuling, 1988; Dronov *et al.*, 2000) indicate, that the regression started already in the late Volkhov time, and the upper part of the Volkhov Stage (the Langevoja Substage and substantial parts of the middle, Vääna Substage) are also missing in northwestern Estonia.

More complete sequences of the Volkhov / Kunda interval occur in the Central Baltoscandian Confacies Belt in Latvia, South Estonia and Sweden. According to the former publications, the Kunda Stage is represented by the Šakyna and Baldone formations in South Estonia and Latvia. Since Paškevičius (1976), judging from the graptolite and trilobite evidence, the base of the Šakyna Formation (the unique grey-coloured limestone interval in predominantly red-coloured limestone succession of the area) is considered as equivalent to the base of the Kunda Stage. In the Tartu core, Meidla (1998), interpreting the ostracode, chitinozoan and conodont distribution, placed the lower boundary of the Kunda Stage 2.4 m below the grey limestone, where the zonal boundaries of *Pinnatulites procerus* and *Lenodus* sp. A (the species considered as the index of *Eoplacognathus ? variabilis* Biozone by Stouge and Bagnoli, 1990) are recorded (Fig. 1).



In the Hällekis section (Tinn and Meidla, 2001), the same part of the Middle Ordovician sequence is cropping out and the sequence is largely similar to the Latvian one. In predominantly red-coloured limestone sequence (Lanna and Holen limestones according to Jaanusson, 1982), a grey-coloured 2 m thick limestone unit (locally termed as 'Täljsten') occurs at nearly the same stratigraphic level. In the Ordovician correlation chart of the East European Platform (Resheniya..., 1987) the base of the Holen Limestone (and the "grey interval" in its lowermost part) is tentatively correlated with the base of the Šakyna Formation and this correlation became widely accepted. However, in the Hällekis section, the aforementioned zonal boundaries are drawn differently. The first occurrence of *Pinnatulites procerus* is recorded at the level of -12.50 m. The base of the Biozone of *Lenodus variabilis* (Zhang, 1998; ? = *Lenodus* sp. A in Stouge and Bagnoli, 1990) was established near the reference level (depth 0 in the Fig. 1, representing the lower boundary of the grey interval), while the appearance of *Asaphus expansus* in the Hällekis section is preliminarily established at about 3 m below the reference level (J. Villumsen, pers. comm.). The boundary of the *A. expansus* Biozone coincides with the major rise in ostracode diversity and gradual increase in concentration of bioclastic material, representing rather a facies boundary than an immigration or speciation level (Tinn and Meidla, 2001). The conodont faunal logs (Stouge in Põldvere *et al.*, 1998; Zhang, 1998) demonstrate a complete faunal representation in the studied intervals of the Tartu and Hällekis sections, suggesting that both sections are fairly complete.

The Kunda Stage was originally defined as beginning with the *Asaphus expansus* Biozone. In Hällekis section both the "Täljsten" layer and the lower boundary of the *Lenodus variabilis* Biozone lie higher up of this level and tentatively could be correlated with the lower part of the Kunda Stage. If this is correct, the boundary of the *Pinnatulites procerus* Biozone is drawn in the upper Volkhov Stage.

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CONTINUOUS SEISMIC REFLECTION PROFILING – A TOOL FOR INVESTIGATION OF THE PALAEOZOIC SEQUENCES IN THE BALTIC SEA

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A large part of the Cambro–Silurian sequences of Baltoscandia is located under the Baltic Sea. Thus, they are not accessible to the exploration techniques normally applied for lithological, stratigraphical and sedimentological investigations on land. From a geological point of view, the Baltic Sea conceals an area of key importance for correlation of the Cambro–Silurian successions between the Swedish and Estonian mainlands, and also for reestablishment of the lithofacies distribution. So far, the only systematic data concerning the geology of the submarine Palaeozoic sequences of this area is based on continuous seismic reflection profiling (Flodén 1980; Tuuling 1998).

THE METHOD OF CONTINUOUS SEISMIC REFLECTION PROFILING

As echosounders and seismic systems in general, the continuous seismic reflection profiling method is based on velocity alterations of elastic body waves passing through a layered rock sequence, i.e. on their ability to provide seismic energy reflections at sharp lithological boundaries. Recording the reflected energy is a means to determine the depth to each successive reflecting horizon, thus fixing the major levels of lithological changes in a stratified rock sequence. Hence, this technique works firstly as a lithostratigraphical tool, separating distinctive lithological boundaries in a vertically layered rock body.

THE SEISMIC FREQUENCY – A KEY PARAMETER IN DECIDING BETWEEN QUALITY AND DEPTH OF THE GEOLOGICAL INFORMATION

This method always contains a discrepancy between accuracy, namely vertical resolution, and depth of penetration. Both parameters depend on the frequency band of the produced seismic waves. Low frequencies generate greater wavelengths, thus providing greater investigation depths. On the other hand, great wavelengths always decrease the vertical resolution in the recordings. In this context, vertical resolution means the ability to resolve between two adjacent reflecting horizons. In optimal cases the resolution is about 1/4 of the wavelength. In practice, however, it remains on the order of 1/3 – 1/2 of wavelength.

THE FREQUENCY SPECTRUM OF THE CONTINUOUS SEISMIC PROFILER

Depending on the purpose of a seismic investigation, different kinds of seismic transmitters are used. Apart from signal strength that normally follows with the size of a transmitter, different transmitters provide different pulse lengths and different frequency spectrums. In order to achieve major penetration into layered bedrock sequences, a frequency range below 1000 Hz is needed. Air guns of different configurations are most commonly used to fulfil this requirement.

The seismic profiling performed in the northern Baltic during recent decades has been performed using an air gun, namely a PAR–1600. The frequency spectrum of the air gun, ranging from less then 50 Hz to more than 2000 Hz, greatly exceeds the useful frequency band with respect to penetration and resolution in the Palaeozoic sedimentary bedrock. Normally two frequency bands, namely 100–200 and 250–500 Hz, were recorded as a compromise between penetration and resolution.

SEISMIC P-WAVE VELOCITIES IN SEDIMENTARY BEDROCK – A KEY PARAMETER FOR DEPTH CALCULATIONS

In order to calculate depths and thicknesses in seismic reflection recordings, reliable P-wave velocities of the layers in concern are needed. According to laboratory data, the seismic P-wave velocity in the Palaeozoic sequence of the northern Baltic Sea varies from c. 2700 m/s in loose Cambrian sandstones to c. 6000 m/s in cryptocrystalline Ordovician limestones. In most cases the lithological changes in layered bedrock sequences are gradual rather than contrasting, thus providing no distinct seismic reflectors. Therefore, for thickness calculations in the Baltic Palaeozoic, mean P-wave velocities for larger lithological complexes have been used. These mean velocities were determined by refraction soundings with strict control of the layers in the sequence (Flodén 1975). Hence, the mean seismic velocities used in our investigations are 2725 m/s, 3500 m/s and 3000 m/s for Cambrian, Ordovician and Silurian rocks, respectively.

SEISMIC RESOLUTION AND DEPTH OF PENETRATION

Based on the recorded frequency bands, and the mean seismic P-wave velocities, as presented above, the vertical resolution in the recordings is in the magnitude of 4–6 m. The depth of penetration in the Palaeozoic bedrock is normally limited to the magnitude of 200–300 m. The seismic penetration is, however, remarkably reduced in areas of enhanced acoustic impedance contrast between the seawater and the sea bottom, i.e. in areas where hard limestone layers are exposed at the seafloor. Furthermore, the depth of investigation is dramatically reduced in shallow water areas, where reflection multiples overprint the primary reflections.

THE PRIMARY REFLECTOR CONFIGURATION AS AN INDICATOR OF THE INTERNAL BEDDING STRUCTURE – A TOOL IN THE ASSESSMENT OF THE SEDIMENTARY STRUCTURES AND PROCESSES

Layers of similar lithological composition, and of horizontal attitude that can be followed across large areas, are normal for relatively quiet and stable sedimentary basins, as the Paleobaltic Basin. In seismic profiles, this is expressed as more or less continuous and parallel reflectors. The deviations from these kinds of normal images are usually induced by the presence of different syn- and postsedimentary structures. Thus, nonparallel, strongly undulating, or interrupted reflectors point either to an instable and rapidly altering sedimentary environment or to postsedimentary tectonic movements and erosional processes. In addition to lithologial subdivisions, continuous seismic reflection profiles always contain some information about the general changes in lithofacies along the profiles.

SEISMIC INVESTIGATIONS OF THE PALAEOZOIC SEQUENCE IN THE NORTHERN BALTIC SEA

It is pertinent that only a sedimentary basin that developed during calm tectonic conditions can guarantee steady precipitation conditions and facies distribution across large areas. Such basins provide good possibilities for reliable seismic correlations.

Geological investigations in the Baltic countries and in Sweden have proved that the most favourable geological conditions for the formation of an extensive and steady sedimentary sequence in the Palaeobaltic Basin existed during the Ordovician period. Due to a complicated palaeogeographic development, most of the Cambrian sequences on the two sides of the Baltic Sea are diachronuous. The progressing Caledonian orogeny towards the end of the Silurian period increased gradually the tectonic instability in Baltoscandia. This is distinctly expressed in the inconsistent facies distribution of the Silurian sequence in the Basin.

SEISMIC CORRELATION OF THE TRANS-BALTIC CAMBRO–SILURIAN SEQUENCE

The most recent correlation attempts between Sweden and the Baltic countries have proved that there is a general lack of seismic correlation both of the Cambrian and of the lowermost Silurian (Wenlockian) sequences (Tuuling et al. 1997; Flodén et al. 1994). So far, only two seismic reflectors in the Ordovician sequence have been traced across the northern Baltic. These reflectors have been correlated with the sharp lithological boundaries between the Rakvere–Oandu and the Pirgu–Vormsi stages (Flodén et al. 1994). The latter one is the most prominent and strongest reflector in the northern Baltic, having in places the shape of a distinct double reflector. This character is probably attributable to its interference with an additional reflector, being located to the lower boundary of the Pirgu Stage.

Numerous Upper Ordovician carbonate buildups severely complicate the normal reflector configuration northeast of Gotland. This caused major difficulties for the correlation of the Ordovician–Silurian boundary. According to the mainland sections, the latter boundary was expected to be one of the clearest reflecting horizons in the region. Furthermore, in the western part of the northern Baltic Sea, the Ordovician Silurian boundary irregularity is enhanced by erosion, which has caused clear undulations in its vertical position.

SYN- AND POSTSEDIMENTARY STRUCTURES DISTORTING THE PARALLEL BEDDING STRUCTURE

The most remarkable structures distorting the parallel bedding configuration in the Baltic are the carbonate buildups. They usually occur in the form of patch reef-like bodies, marking the nearshore shallow water zones in the Palaeobaltic basin during various Ordovician and Silurian time intervals. In places, however, they appear in large quantities, forming elongated barrier reef-like structures with distinct lithofacies differences along their land- and basinward sides (Tuuling & Flodén 2000; Flodén et al. 2001).

The geology of the reef-like structures, and their surrounding sediments, are best explored northeast of Gotland, where analogous structures of Upper Ordovician age reveal downslope variations in quantities, morphologies and dimensions. Overwhelming deviations from the normal reflector configurations occur within and around the large swarms of carbonate buildups. This is, first of all, due to the rapidly varying content and thickness of the inter-reef sediments. Some strongly chaotic bedding features around the carbonate buildups are probably due to variations in syn- and postsedimentary compaction, accompanied by gravitational collapse and spalling of the reef structures. In places, the tilted bedding configuration is probably caused by clinoform structures that developed along the flanks of the carbonate buildups.

In a local area northeast of Gotland, the primary sedimentary structure of the Cambrian and lowermost Ordovician sequences are occasionally replaced by an image of unlayered rocks of crystalline basement. This area, located about midway between Gotska Sandön and Hiiumaa, conforms to a zone of tectonic disturbances. The fading of the primary stratification is obviously due to solutions that have percolating through the crashed and smashed rocks, causing dolomitization of the sequences along the zone.

TECTONIC DEFORMATIONS

Folds and faults of reasonable extensions and amplitudes are easily identified in the seismic profiles where they appear as abrupt undulations of the reflectors, or as interruptions in reflector continuity. Moreover, the continuous image of a reflector provides the best opportunity for general morphology and structure investigations of tectonic deformations being usually inaccessible in mainland studies.

The main north-south direction of the seismic profiles shot in the northern Baltic Sea coincides with the prevailing orientation of the fault structures (Tuuling & Flodén 2001). This parallelism has remarkably reduced the number of discovered faults. Nevertheless, small interruptions of the reflectors, marking minor displacements of layers, occur frequently. Faults with amplitudes of 10 m and above occur rarely. The large faults mostly occur as solitary structures, although in places a few closely spaced faults can form a step-wise dislocated fault system.

Fold-like structures of tectonic origin are very rare in the northern Baltic. Such folds have merely been identified along a tectonically active zone some 50–60 km northeast of Gotland (Tuuling et al. 1997). They can reach several hundreds of metres in diameter and up to 10 m in vertical direction.

In a general way, the quantity, extension and amplitude of the tectonic deformations in northern Baltic increase westwards, reaching their maximum around northern Gotland. As well known from the mainland areas, the tectonic deformations in the northern Baltic are largely restricted to numerous linear zones (Tuuling & Flodén 2001). These tectonically active zones, with crashed and smashed rocks, are often distinctly imprinted in the seismic profiles as vertically extensive stripes of chaotic and blurred reflectors. In the present bedrock relief, the zones are frequently expressed by erosional depressions filled in with Quaternary sediments.

EROSIONAL FEATURES

Paleoerosional features in the sedimentary bedrock are usually visible in the seismic recordings as undulating disconformities or as reflectors that cut through the horizontally bedded layers at acute angles. They can furthermore be decipherable along the profiles as seismostratigraphical units that decrease regularly in thickness, or as reflectors that are truncating primarily arch- or conical-shaped tops of carbonate buildups.

Obvious features of erosional origin in the submarine Ordovician sequence appear at two stratigraphical levels (Tuuling & Flodén 2000). The lower one corresponds to the boundary of the Vormsi and Nabala stages. It appears in the central part of the northern Baltic and is expressed as a regular channel-like erosional depression. Another distinct disconformity of erosional origin coincides with the Ordovician Silurian boundary within a large area closely northeast of Gotland. Although in wide areas it is expressed as an ordinary reflecting horizon, its unambiguous erosional origin is in many places recognizable due to vigorous undulation or slope-like configuration of the reflector. Along some profiles, this reflector marks a regular thickness decline in the underlying rock unit, i.e. in the uppermost Ordovician sequence.

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ASTEROLEPIS (PLACODERMI): BIOGEOGRAPHY, BIOSTRATIGRAPHY AND HABITAT

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BIOGEOGRAPHY AND BIOSTRATIGRAPHY

Placoderms are largely restricted to the Devonian and are often well preserved due to their dermal armour. The antiarchs have been much used in biostratigraphical zonations of the Middle and Upper Devonian sediments on and peripheral to the Old Red Sandstone continent (Karatajūtė, 1958a; Young, 1974; Blieck et al., 1988). The antiarch zonation is best developed in the East Baltic, where species lineages within single genera are established (Lyarskaya, 1981; Esin et al., 2000).

The genus *Asterolepis* has been intensively studied for correlating and assessing the age of rock sequences in E-European platform (Gross, 1942; Karatajūtė, 1958a, 1963) and in E-Greenland (Marshall & Astin, 1996). *Asterolepis* is used in the vertebrate zonation and correlation of the Givetian and Frasnian sediments in the Baltic area, NW-Russia (*A. dellei, A. ornata, A. radiata*), Scotland (*A. orcadensis, A. thule, A. maxima*), N-Timan (*A. radiata*).

Asterolepis has been recorded with full certainty from the Baltic States, Russia (NW-Russia and Timan), Scotland, Spitsbergen, East Greenland and U.S.A., e.g. the area of its distribution coinsides with the Euroamerica paleocontinent. It is being represented by 13 named species and 11 unnamed species (see Table), from that 7 and 4, respectively, are restricted mainly to the Main Devonian Field (E-Baltic area: Estonia, Latvia, Lithuania, NW part of Russia and Belarus). The widest distributed species of the genus appears *A. radiata*: it remains occurs also in the N-Timan and Central Devonian Field.

The first appearance of the genus *Asterolepis* is known from the Emsian of Belarus (Mark-Kurik, 2000). Starting the middle Eifelian (Narva RS) *A. estonica* appears in the Main- and Central Devonian Fields (Karatajūtė-Talimaa, 1963). The genus its widest distribution reaches in the Late Givetian and Early Frasnian: a number of species are recognized in the Main Devonian Field, Scotland, E-Greenland, France, Belgium, Spitsbergen, U.S.A. (Table).

The upper limit to *Asterolepis* range overlaps with *Bothriolepis,* which replaces it in Frasnian in East Greenland, Scotland, East Baltic (Young, 1974; Lyarskaya, 1981; Marshall & Astin, 1996;) and U.S.A. (Denison, 1978); in Famennian in Belgium (Lelievre & Goujet, 1986). Eight species of *Bothriolepis* are used in the vertebrate zonation of the Frasnian and Famennian sediments of the Main Devonian Field (Esin et al., 2000).

East Baltic Middle and Late Devonian sediments containing the *Asterolepis* remains are shallow-marine and deltaic genesis (Kuršs, 1975, 1992 a, b). In marine sediments they occur in Belgium, France, and Iran (Schultze & Cloutier, 1996). In Orkney and E-Greenland *Asterolepis* is believed to be restricted to fluvial and marginal/shallow lacustrine environments (Marshall, Astin, 1996).

HABITATS

Antiarchs were benthic forms (Gross, 1931; Karatajūtė, 1958b) with weak, toothless jaws and small mouths capable of eating only small food (Denison, 1961, 1975). They are believed to have been bottom feeders. Mud fillings of intestines are known in *Bothriolepis* (Denison, 1941); the preserved stomach content of detritious material is recognised in a number of juveniles of *Asterolepis* (Upeniece & Upenieks, 1992; Upeniece, 2001). *Asterolepis* appears to have been a detritus feeder and benthonic scavenger.

***Table. Asterolepis* localities, stratum and age.**

Asterolepis species	Country, province, locality	Stratum	Age	References
Baltic craton + Timan				
A. sp.	Belarus	Vitebsk RS Gorodok RS, Kastyukovichy RS Polotsk RS	Emsian late Eifelian Gv	Mark-Kurik, 2000
A.estonica Gross, 1940	1. S-Estonia (lake Võrtsjärv, Tartu) 2. Latvia (boreholes) 3. NW-Russia, St.-Petersburg district (river Luga) 4. Belarus (borehole Starobin) 5. C-Russia, Kaluga district	Narva, Aruküla Fms.	m/l Eifelian	Gross, 1940 Karatajute-Talimaa, 1963
A.dellei Gross, 1940	1. N-Latvia (rivers Salaca, Roja, Peldanga, boreholes) 2. NW-Russia (rivers Luga, Yoglina) 3. NW-Russia, Pskov distr.(river Yoglina, Bez'va, Stolbi) 4. Belarus (borehole Bogushevskaya)	Burtnieki Fm. Polotsk RS	Gv	Gross, 1940 Karatajute-Talimaa, 1963 Lyarskaya, 1981
A. sp. 1	S-Estonia, Karksi	Burtnieki Fm.	Gv	Karatajute-Talimaa, 1963
A. sp. 2	NW- Russia (river Yoglina, Bez'va)	Burtnieki Fm.	Gv	Karatajute-Talimaa, 1963
A. sp. 3	W-Latvia (river Abava, Muižarāji)	Gauja Fm.	early Fr	Karatajute-Talimaa, 1963
A.essica Lyarskaya, 1981	SE-Estonia (river Võhandu, Essi)	Gauja Fm.	early Fr	Lyarskaya, 1981
A.ornata Eichwald, 1840	1. Latvia (rivers Gauja, Brasla, Amata) 2. S-Estonia (river Piuza, Jõksi) 3. NW-Russia, Pechori, Pachovka	Gauja Fm.	early Fr	Gross, 1931, 1933 Lyarskaya, 1981
A.radiata Rohon, 1900	1. N-Timan 2. E-Baltic 3. NW-Russia 4. C-Russia (Kursk, Lipeck) 5. Belarus	Kumushka Fm. Amata, Pļaviņas Fms. Shchigry RS Lan` RS	early Fr	Karatajute-Talimaa, 1963 Lyarskaya, 1981 Ivanov & Lukševičs, 1996
A.syasiensis Lyarskaya, 1981	NW Russia (river Syas', Stolbovo)	Dubniki Fm.	early Fr	Lyarskaya, 1981
A.? amulensis Lyarskaya, 1981	W-Latvia (river Amula, Kalnamuiž a)	Ogre Fm.	m/l Fr	Lyarskaya, 1981
Belgium, France				
A. sp.	Belgium, Namur, Hingeon	Assize de Mazy	Gv	Gross, 1965 Blieck & Lelievre, 1995
A. sp.	France, Boulonnais, Carriere "la Parisienne"	Blacourt Fm.	Gv	Lelievre et al., 1988
Scotland				
A.orcadensis Watson, 1932	Orkney Islands	Upper Rousay series	Gv	Watson, 1932; Denison, 1978
A.thule Watson, 1932	Caithness, John O'Groats	John O'Groats sandstones	late Gv	House M. et al., 1977 Schultze & Cloutier, 1996 Watson, 1932
A.maxima (Agassiz), 1844	Nairnshire, Kingsteps	Nairn sandstones	late Gv	Miles, 1968 Schultze & Cloutier, 1996
	Morayshire, Boghole	"Boghole beds", Nairn sandstones	early Fr	Miles, 1968 Denison, 1978
A.? sp.	Morayshire, Whitemire	Edenkillie beds	early Fr	Miles, 1968
Greenland				
A.säve-söderberghi Stensio & Säve-Söderbergh, 1938	East Greenland, Canning Land and Wegener Peninsula	*Asterolepis* series	late Gv	Stensio & Säve-Söderbergh, 1938 Marshall & Astin, 1996
Spitsbergen				
A.scabra (Woodward), 1891	Mimer Valley	"Fish Cleft"	Gv	Nilson, 1941
A.cf.orcadensis	Wijde Bay, Andredalen	Wijde Bay Series	Gv	Nilson, 1941
*A.*sp.	Wijde Bay, Andredalen	Wijde Bay Series	Gv	Nilson, 1941
U.S.A.				
*A.*sp. (2 species)	Nevada, Red Hill	Denay Fm.	early Fr	Murphy et al., 1976 Gregory et al., 1977
A.chadwicki Wells, 1964	New-York, Parksville	Lower Katsberg Fm. (Senecan)	early Fr	Wells J., 1964
Iran				
? A. sp.	Central Iran, Kerman, Bidou 1 limestone	*Cyrtospirifer verneuili*	e/m Fr	Janvier, 1979 Schultze & Cloutier, 1996

Abbreviations: Gv – Givetian, Fr – Frasnian; Fm – Formation, RS – Regional Stage

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THE ACANTHODIAN STANDARD FOR THE LOWER AND MIDDLE DEVONIAN OF THE OLD RED SANDSTONE CONTINENT

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The calibrated scale on the left defines a base of Devonian 410 Ma and the top of Middle Devonian 372 Ma (Blieck & Turner, 2000). The first almost corresponds to the Devonian boundary age of the global geochronometric standard 89 (Harland *et al.*, 1990), whereas the second differs in references by about 8 Ma, e.g., 369 Ma by G. C. Young (1995) and 377 Ma in the GGS 89. The stage boundaries are coincident with the GGS 89 except for Givetian. They indicate the longest duration of Lochkovian (14 Ma), shortening Pragian (6 Ma) and Emsian (4 Ma), and Givetian (8 Ma) longer than Eifelian (6 Ma).

The stage boundaries adopted by the SDS and defined by the standard conodont zones (CZ) are applied to estimate duration of acanthodian zones (AZ). Quantifying duration of CZ, there are used scores referred to for the Australian series (Burrow *et al.*, 1993; Young, 1995), where the CZ are grouped into negligible (*partitus*), small (*patulus, australis, hemiansatus*), small+ (*eurekaensis, inversus*), medium (*woschmidti/hesperius, pesavis, kindlei, pireneae, hermanni-cristatus, disparilis*), big (*costatus, kockelianus*) and very big (*dehiscens, perbonus-gronbergi, serotinus, varcus*). The main difficulties concern the Lochkovian CZ, because the stage boundaries are recorded differently, 404–410 Ma by G. C. Young and 396–409 Ma in the GGS 89. Thus the *eurekaensis* and *delta* CZ became supposed a distinctly longer duration and have to be qualified as very big.

The basic information of the ORSC acanthodians is obtained from the regions of the eastern part, the Baltica terrain (Figs. 1–2), whereas the west part, Laurentia (Canada, Spitsbergen, Greenland), has provided little applicable datasets due to the incompleteness of studied stratigraphic interval or taxonomic scarcity of referred fauna. The Ibero–Armorican acanthodians are not included, as the ORSC margins along the south-west Europe remain ill- defined, and both regions reconstructed differently, either near to the northern Europe or at the northwestern margin of Gondwana.

The lowermost Lochkovian *Nostolepis minima* AZ is best grounded in the Baltic and Podolia, conditioned by dominating and age diagnostic nostolepids (Valiukevičius & Kruchek, 2000) and some characteristic ischnacanthids, e.g., *Poracanthodes subporosus.* The Timan–Pechora series, composed of terrigenous--carbonate facies, contains native *Nostolepis zinaidae, N. kozhymica, N. adzvensis, N. terraborea,* etc. (Valiukevičius, 2002). The supplementary zonal units are the *Poracanthodes menneri* AZ of Severnaya Zemlya (defined on articulated acanthodians) and *Nostolepis* sp. cf. *N. tcherkesovae* AZ of Timan–Pechora, concurrent to the lower part of the *N. minima* AZ. The age of the zone is grounded by the *remscheidensis* conodont assemblages of Belarus (Golubtsov *et al.*, 1981) and Timan–Pechora (Deulin, 1991), and the *woschmidti-postwoschmidti* CZ in Podolia (Drygant, 1988).

The other regions for correlations are the Central Urals, yielded the supposed *Poracanthodes menneri* (Märss, 1997) and nostolepids (recognisable *N. zinaidae, N. kozhymica, N. athleta*), and the Welsh Borderland, at the base of Ditton Group beared an ischnacanthid dominated assemblage (Vergoossen, 2000) corresponding to the *leathensis* Zone of pteraspids and the basal part of the *micrornatus-newportensis* miospore zone.

The *Lietuvacanthus fossulatus* AZ, tentatively attached to the *delta* CZ in the Baltic and the *delta-pesavis* in Podolia on grounds of the *Icriodus serus* local zone age, is defined by the key fossil and *Ectopacanthus pusillus, Tareyacanthus dissectus* and *Cheiracanthoides nativus.* On the Severnaya Zemlya, it corresponds to the *Lietuvacanthus-Tareyacanthus* fauna entered by *Diplacanthus poltnigi,* the another characteristic taxon. In Timan–Pechora, the *Ectopacanthus-Tareyacanthus-Diplacanthus* AZ comprises the *delta-pesavis* and the *sulcatus* CZ of Pragian, as it includes at the top the concurrent *Nostolepis taimyrica* AZ, aged by the correlation with the Taimyr and Salair carbonate series yielded *Vjaloviodus marinae* CZ assemblage (Sobolev, 1990) defining the standard *sulcatus.* The AZ is also supposed to develop in the Arctic Canada, Member A of the Red Canyon River Formation, basing on the representatives of *Lietuvacanthus, Ectopacanthus* and *Diplacanthus.*

The *Nostolepis watsoni* and *Gomphonchus tauragensis* AZ, originated respectively in Timan–Pechora and the Baltic, supposedly cover the *kindlei* and *pireneae* CZ of Pragian and the *dehiscens* of early Emsian, as the last one correlates with the *Amaltheolepis baltica* thelodont zone (Blieck *et al.*, 2000) and is supported by the *Emphanisporites annulatus* miospore assemblage. The boundaries are ill-defined because they are grounded by the single index fossils.

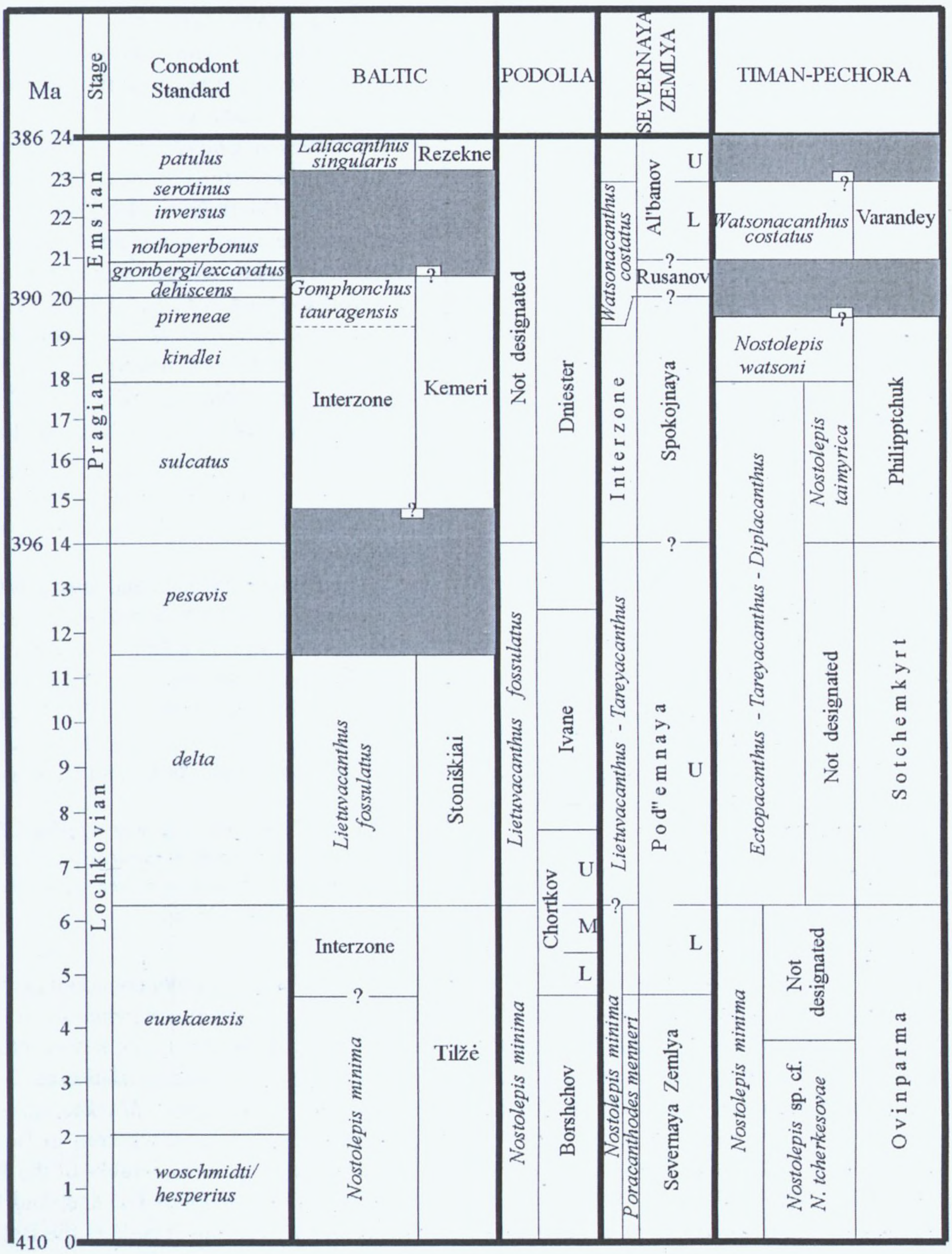

Fig. 1.A.

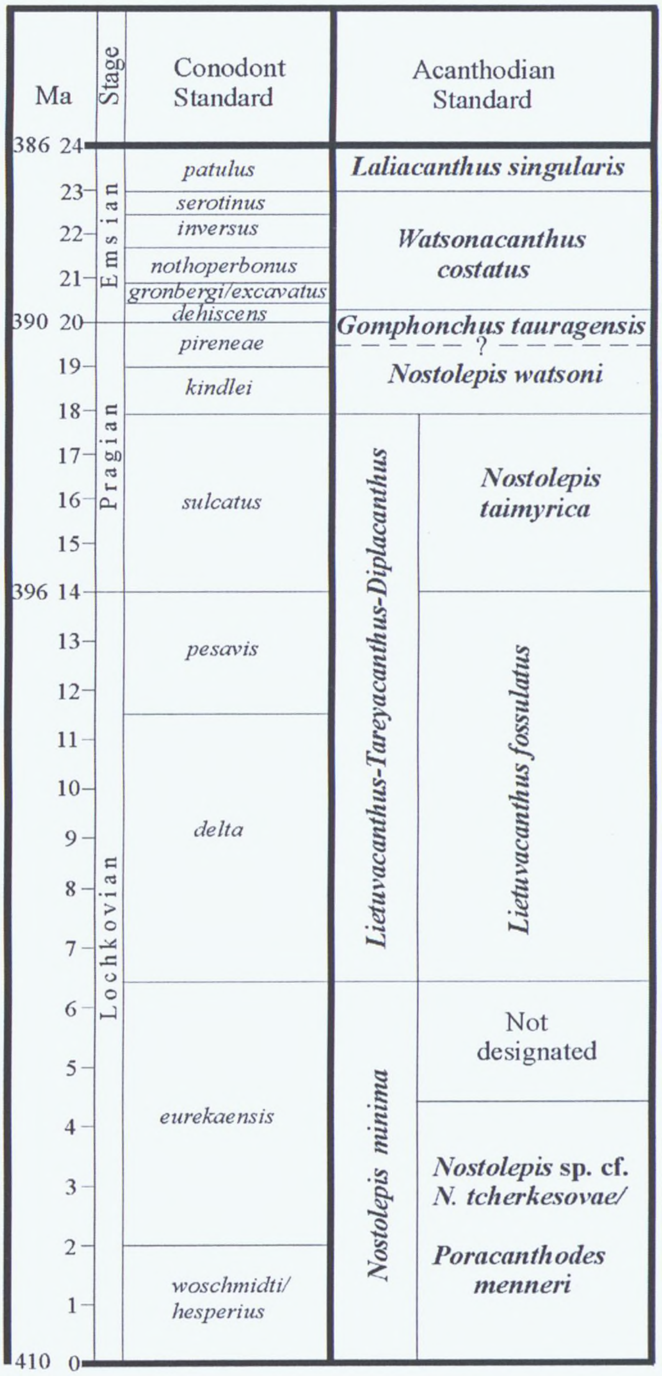

Fig. 1.B.

The *Watsonacanthus costatus* AZ is dated by the *dehiscens* to *serotinus* standard CZ of Emsian on grounds of conodont association with predominant *Pandorinellina exigua exigua, P. e. philipi* and *P. e. expansa*, obtained from the Severnaya Zemlya and Timan–Pechora. The keys are *Nostolepis curta, N. multangula* and *Cheiracanthoides comptus.*

The Lower/Middle Devonian boundary is not certainly grounded by acanthodians, as the *Laliacanthus singularis* AZ is correlated with the *patulus* and *partitus* standard CZ. The acanthodians define a larger closeness to Middle Devonian because the characteristic Middle Devonian genera, *Rhadinacanthus, Ptychodictyon, Cheiracanthus, Markacanthus* and *Haplacanthus* occur. Contrary, the placoderms (Mark-Kurik, 1991) and miospores indicate the late Emsian age basing on the *Hystricosporites elegans* Zone in the Baltic (Valiukevičius *et al.*, 1986) and the *Diaphanospora inassueta* in Belarus (Avkhimovitch *et al.*, 1988). The another region for correlations is Spitsbergen with a close acanthodian assemblage of the Grey Hoek Fm. yielding characteristic *Watsonacanthus oervigi, Ptychodictyon ancestralis, Rhadinacanthus primaris, Cheiracanthus gibbosus,* etc.

The Eifelian *costatus* and *australis* CZ age is given for the *Cheiracanthoides estonicus* (the second key is *Cheiracanthus crassus*) and *Ptychodictyon rimosum* AZ (representatives of *Rhadinacanthus, Ptychodictyon, Cheiracanthus*) grounded by the identical assemblages from the Baltic and Belarus.

Fig. 1. Principal regions applied for the acanthodian zonality standardisation (A) and a proposed acanthodian standard (B) for the Lower Devonian of the ORSC.

The *Nostolepis kernavensis* AZ is aged latest Eifelian by the *Polygnathus parawebbi* conodont association in the Baltic, Belarus and the Central Devonian Field, and *Eognathodus bipennatus bipennatus* in Timan–Pechora (Valiukevičius & Kruchek, 2000). It correlates with the standard *kockelianus*. The key acanthodians are *Nostolepis kernavensis, Cheiracanthoides mosolovicus, Ch. proprius, Diplacanthus solidus,* etc.

The Givetian *Diplacanthus gravis* AZ (*D. gravis, Rhadinacanthus multisulcatus, Markacanthus alius* for keys) is best grounded in the Baltic from the Aruküla and Burtnieki horizons, whereas in Belarus it attaches to the Goryn' beds only, as the *Devononchus concinnus* AZ loweres considerably in the Polotsk Horizon. Thus for the *D. gravis* AZ is supposed the *hemiansatus* to middle *varcus* CZ age, and for the *D. concinnus* AZ – a scope of standard CZ beginning with the *hermanni-cristatus* (Gauja in the Baltic and Timan horizons in Timan–Pechora) and continuing into Frasnian.

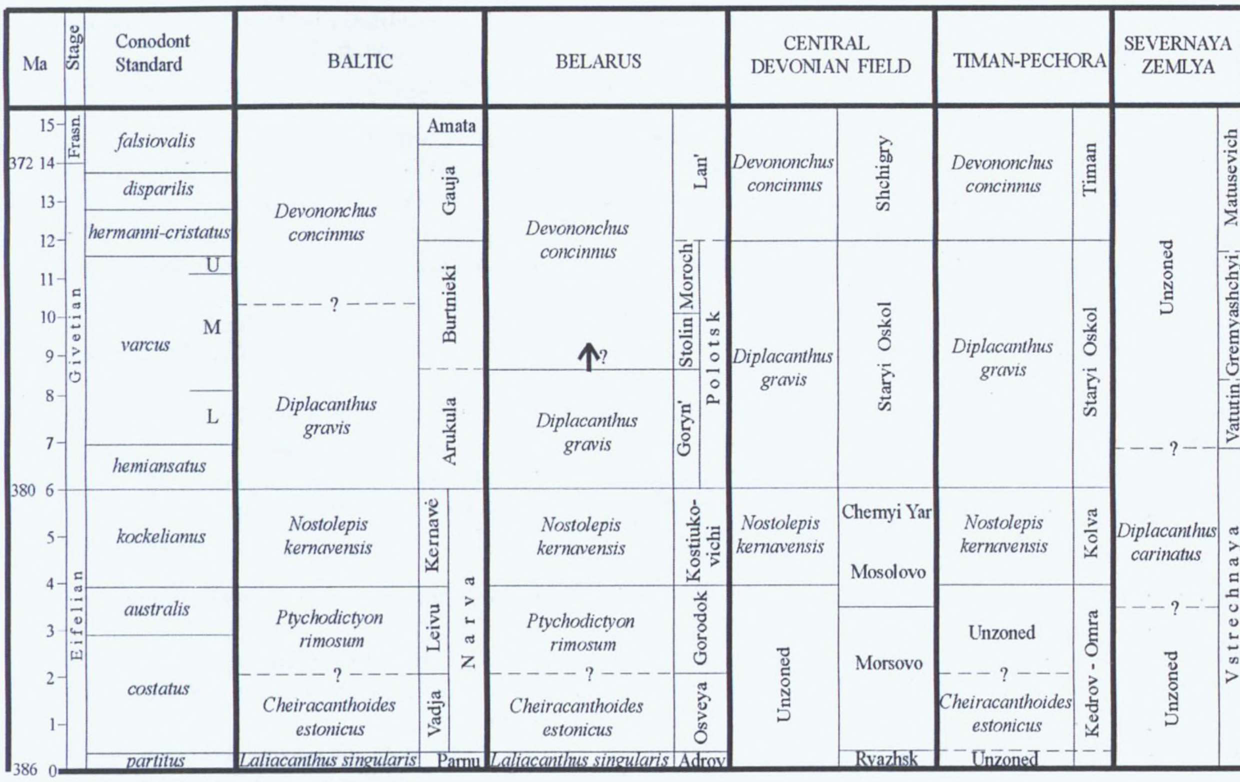

Fig. 2.C.

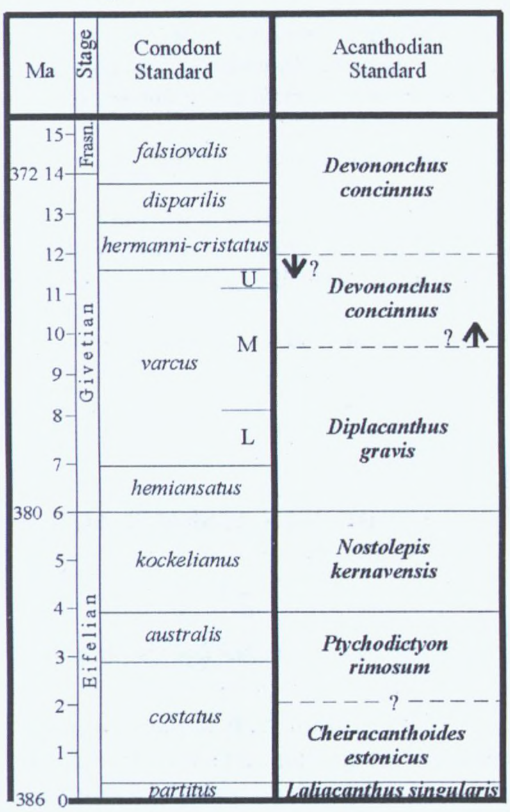

Fig. 2.D.

Fig. 2. Principal regions applied for the acanthodian zonality standardisation (C) and a proposed acanthodian standard (D) for the Middle Devonian of the ORSC.

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REAPPRAISAL OF THE *WATSONACANTHUS OERVIGI* ACANTHODIAN ASSEMBLAGE OF THE GREY HOEK FORMATION FROM THE WEST SPITSBERGEN

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1979 author has published a paper on acanthodians of the Grey Hoek Formation with the description of some new taxa and brief stratigraphic attachment of samples according to the smaller units, the Elsvikfjellet, Tavlefjellet and Forkdalen subformations. The data need a correction now, particularly after the Emsian and Eifelian fossil assemblages of Taimyr, Timan–Pechora and Severnaya Zemlya have been studied.

T. Ørvig (1969) was first to refer "indeterminable fish scales" from the Verdalen Member of the Wood Bay Formation, which were later named *Watsonacanthus oervigi* (Valiukevičius, 1979). In this publication, three further new acanthodians, *Ptychodictyon distinctum, Isodendracanthus ramiformis* and *Ectopacanthus cristiformis* have been described. Concerning the stratigraphic distribution of taxa, there were supposed different associations for the lower (Elsvikfjellet Sfm.) and upper (Forkdalen Sfm.) parts of the Grey Hoek Fm.

Watsonacanthus oervigi and *Isodendracanthus ramiformis* have been defined characteristic of the first one, whereas the second was proposed dominated by *Ptychodictyon distinctum, Cheiracanthus longicostatus, Ectopacanthus cristiformis* and representatives of *Acanthoides*.

Partly revised acanthodians of the Grey Hoek Fm. were multiply discussed in the publications concerning the Emsian/Eifelian boundary problem (Mark-Kurik, 1991; Valiukevičius, 1988, 1994; Valiukevičius & Kruchek, 2000; Blieck & Cloutier *et al.*, 2000), but never have been represented completely by the corrected version.

As it is recently followed from the personal communications with V. Karatajūtė-Talimaa, who has participated the field work, the samples for acanthodian test distinctly fall to the lower part of the formation. This perhaps has caused, that throughout the sampled stratigraphical interval, the only *Watsonacanthus oervigi* assemblage can be defined. It is composed of 14 acanthodian taxa. Of them only five, *Acanthoides?* sp. C, the second commonest *Watsonacanthus oervigi* and a little rarely met *Ectopacanthus cristiformis, E.* sp. cf. *E. flabellatus* and *Ptychodictyon distinctum,* make the principal content of the assemblage (Fig. 1). The other species are identified sporadically distributed. *Cheiracanthus gibbosus* occurs only in the Watson and Tavlefjellet Hill localities, *Ch.* sp. cf. *Ch. longicostatus* – in the last one, *Ch.* sp. cf. *Ch. gibbosus* is met in all three localities of the Dickson Land, and *Diplacanthus kleesmentae* – in the Watson and Tavlefjellet Hills and Austfjord. The Watson Hill and Austfjord areas show the largest closeness of acanthodian taxonomic composition. The samples from the Tavlefjellet Hill stand out by an almost absent *Watsonacanthus oervigi,* the key fossil of the assemblage.

Locality			Sample	*Ectopacanthus* sp. cf. *E. flabellatus* VALIUK.	*Ectopacanthus cristiformis* VALIUK.	*Watsonacanthus oervigi* VALIUK.	*Cheiracanthus gibbosus* VALIUK.	*Acanthoides?* sp. C VALIUK.	*Ptychodictyon distinctum* VALIUK.	*Acanthoides?* sp. B VALIUK.	*Isodendracanthus ramiformis* VALIUK.	*Cheiracanthus* sp. cf. *Ch. gibbosus* VALIUK.	*Diplacanthus kleesmentae* VALIUK.	*Acanthoides?* sp. D VALIUK.	*Cheiracanthus* sp. cf. *Ch. longicostatus* GROSS	*Ptychodictyon ancestralis* VALIUK.	*Diplacanthus?* sp.	Formation
Dickson Land	Watson Hill		97 F	✖	✖	●	✖	●	●	✖	✖	✖	✖					Grey Hoek
			97-1	✖			✖	●	✖									
			109 F	✖	✖	✖		●	✖									
			110 F			✖		✖	✖									
	Mimerdalen		6/4-A						✖									
			11/2					✖										
			12/4	✖	✖	✖		●				✖						
	Austfjord		29 F	✖		●												
			35 F	✖	✖	●		●	✖		✖	✖	✖					
Andre Land	Wijdefjord	Tavlefjellet Hill	B-1					✖						✖				
			B-2	●	✖			●	✖				✖		✖	✖	✖	
			B-2 (1)	✖	✖		✖	●										
			B-2 (2)		✖			✖	✖									
			B-6	✖			✖								✖			
			B-6 (1)					✖										
			1326			✖	✖	✖										
		Heintz Hill	153-A	✖	✖			✖										
	Skamdalen		B-5(talus)			✖		✖										
			86	✖	✖	✖		●	●		✖							
			420	✖	✖	✖		●	✖		✖							
			1331-228			✖								✖				
Woodfjord			1142 (talus)			✖												
			73/13	✖		✖		✖										
			83/8			✖			✖									

Fig. 1. Distribution of acanthodians in the localities of West Spitsbergen. The predominant taxa are shown in black circles.

What is the correlational and age dating potential of acanthodians from the Grey Hoek Fm.? The zonal indices and other most important acanthodian species of the Emsian and the lowermost Eifelian are taken in Fig. 2. The earliest occurrence of watsonacanthids (*W. costatus*) is attached to the Rusanov Fm. of Severnaya Zemlya, where it is correlated with the *dehiscens* standard conodont zone. In Taimyr, the species appears in the Dolgan Beds (*gronbergi* CZ) and in Timan–Pechora – in the Varandey Fm., tentatively attributed to the *nothoperbonus-inversus* CZ. In all regions, the taxon disappears in the *inversus* Zone of conodont standard, which is indicated basing on the representatives of *Pandorinellina* and *Steptotaxis,* grounding regional conodont zonations (Sobolev, 1990, 1999). *Watsonacanthus costatus* is closely associated with the predominant nostolepids in all regions. *Watsonacanthus sibiricus,* the zonal fossil of the overlying *Wijdeaspis arctica* Beds of Taimyr, is accompanied by *Nostolepis infida,* having no related species in other regions. The zone is supposedly correlated with the *serotinus* CZ.

The acanthodians of the Grey Hoek Fm. are most similar to the Rezekne and Pärnu assemblage of the Baltic. *Ptychodictyon ancestralis, Cheiracanthus gibbosus, Diplacanthus kleesmentae* and *Ectopacanthus flabellatus* are characteristic taxa for both regions. Particularly notable, that almost nothing is in common with the earlier Lower Devonian acanthodians of the Baltic and other regions. The only exception is *Ectopacanthus flabellatus,* occurring first in the Saunoriai Fm. of Lithuania, attributed to the *Emphanisporites annulatus-E. schultzii* miospore zone (Vaitiekūnienė *in* Valiukevičius, 1994). The lower part of Rezekne Fm. contains spores of the *Hystricosporites elegans,* whereas the upper part together with the Pärnu Fm. is assigned to the *Periplecotriletes tortus* Zone considered Middle Devonian in age. These regional zones are correlated with the considerable part of the *apiculatus-proteus* zones of the European palynomorph scale (Streel *et al.*, 1987).

Summing up palaeontologic datasets, the Rezekne and Pärnu formations of the Baltic are dated by the *patulus* and *partitus* CZ age. The lower part of the Grey Hoek Fm. on the studied area might be correlated with the *patulus* Zone, corresponding to the Rezekne of Baltic, but is younger than *Wijdeaspis arctica* Beds of Taimyr, contrary to the opinion of Mark-Kurik (1991), who considers these units contemporaneous on the basis of placoderm fishes. Besides, most of placoderms, indicated as the age dating taxa, including *Wijdeaspis arctica,* range stratigraphically whole Emsian and thus are not useful as precise age determinants.

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Conodont Standard	TAIMYR: Formation, Beds	TAIMYR: Significant acanthodian taxa	TAIMYR: Local conodont zonation (Sobolev, 1990)	SEVERNAYA ZEMLYA: Formation, Subformation	SEVERNAYA ZEMLYA: Significant acanthodian taxa	SEVERNAYA ZEMLYA: Local conodont zonation (Sobolev, 1999)	TIMAN-PECHORA: Formation	TIMAN-PECHORA: Significant acanthodian taxa	MAIN DEVONIAN FIELD: Formation	MAIN DEVONIAN FIELD: Significant acanthodian taxa	MAIN DEVONIAN FIELD: Miospore zonation (Vaitiekūnienė *in* Valiukevičius, 1994)	SPITSBERGEN: Formation	SPITSBERGEN: Significant acanthodian taxa
partitus				Vstrech-naya			Kedrov	?↑	Pärnu	***LS*** *ChG DK RP*	*Peripleco. tortus*	Wijde Bay	
patulus				Al'banov, Upper					Rezekne	*LS DB AL PA EF ChK ChG DK RP*	*Hystrico. elagans*	Grey Hoek	***WO*** *PA EF ChG DK*
serotinus	*Wijdeaspis arctica*	*NI* ***WS***		Al'banov, Upper								Wood Bay	↓?
inversus	Dolgan	***WC***	*Pand. expansa Steptotaxis* sp.	Al'banov, Lower	***WC***		Varandey	***WC*** *NTi*				Wood Bay	
nothoperbonus	Dolgan	*NTm*	*Steptotaxis? furnishi*	Al'banov, Lower	*NW NTm*	*Pand. e. exigua Pand. expansa Steptotaxis?* n. sp. UYENO	Varandey	*NMu NW NTm NC*				Wood Bay	
gronbergi/excavatus	Dolgan	*NMu NC*	*P. dehiscens* (late form)	–?–	*NMu*				Saunoriai	***GT*** *EF*	*Emphani. annulatus E. schultzii*	Wood Bay	
dehiscens	Taribigai	***NMu*** *NT TM NC*	*Polygnathus dehiscens*	–?– Rusanov ↓	***WC*** *TD NTi*				Saunoriai	***GT*** *EF*	*Emphani. annulatus E. schultzii*	Wood Bay	

Fig. 2. Correlation of Emsian and the lowermost Eifelian (*partitus* CZ) basing on selected acanthodian taxa with the local conodont and miospore zonation and the conodont standard.
AL – Acanthoides? latgalica, ChG – Cheiracanthus gibbosus, ChK – Cheiracanthus krucheki, DB – Diplacanthus berziensis, DK – Diplacanthus kleesmentae, EF – Ectopacanthus flabellatus, GT – Gomphonchus tauragensis, LS – Laliacanthus singularis, NC – Nostolepis curta, NI – Nostolepis infida, NMu – Nostolepis multangula, NT – Nostolepis tareyensis, NTi – Nostolepis timanica, NTm – Nostolepis taimyrica, NW – Nostolepis watsoni, PA – Ptychodictyon ancestralis, RP – Rhadinacanthus primaris, TD – Tareyacanthus dissectus, TM – Tareyacanthus magnificus, WC – Watsonacanthus costatus, WO – Watsonacanthus oervigi, WS – Watsonacanthus sibiricus.
The zonal species are taken in bold. Line from below of the species index marks appearing taxa, line from above – disappearing taxa. The shadowed intervals concern to stratigraphic gaps.

OBSERVATIONS ON THE CORRELATION OF MICROVERTEBRATE FAUNAS FROM THE LATE SILURIAN ÖVED SANDSTONE FORMATION /ÖSF (SKÅNE, SOUTHERN SWEDEN) WITH THE MICROVERTEBRATE STANDARD, AND ON THE INFLUENCE OF WATER DEPTH ON THE FAUNAL COMPOSITION

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Rock, residue and slide samples from the Swedish Museum of Natural History (SNMH) from Öved Sandstone outcrops (at Helvetesgraven, Klinta, Rinnebäcks, Ramsåsa sites C, D, E, H) mostly yielded microvertebrate faunas that on the whole are typical of *Thelodus sculptilis* zone assemblages (Late Ludlow–Early Přidoli), with the zonal thelodont in most faunas. Species were listed for each rock residue, plus some of the (published) form variants within the species, or new form variants. For one Ramsåsa C rock, and three Ramsåsa E rocks such faunal lists were also made for several size fractions (0.106, 0.212, 0.355, 0.425 and 0.5 mm sieve width). Faunal lists are dominated by thelodont and acanthodian species and composed of different species and form variants per rock. Especially particular form variants proved to be restricted to particular size fractions. Per rock, the most contrasting differences in faunal composition were observed in the 0,106–0,212 size ranges, which also turned out to have the most complete species lists. Data on faunal composition, or on the morphology of species and their form variants, within such small size ranges are not available from the Late Silurian Microvertebrate Standard (the East-Baltic sequence). Another correlational problem is that some of the ÖSF faunas with *T. sculptilis* (Helvetesgraven, Ramsåsa D/SW24) could be older than the *T. sculptilis* zone faunas of the East Baltic succession where a hiatus exists before the first appearance of the zonal fossil.

Judging from newly published time scales (Williams et al., 2000), the studied ÖSF faunas cover a time range of approximately 2–3 million years, which gives them a high biostratigraphical resolution potential. Some of the faunal differences can be attributed to (relatively small) differences in age within the Late Whitcliffian, or Late Ludlow part of the *T. sculptilis* zone. Other differences may be attributable to other, partly environmental, factors.

Age-related differences can be associated with the presence of the older osteichthyan zonal fossil *Andreolepis hedei* (Helvetesgraven), with the presence of the thelodont (and zonal equivalent) *Thelodus admirabilis* and its formae: either in combination with *A. hedei* (Helvetesgraven) or *Paralogania ludlowiensis* (Ramsåsa C), with the regular rather than incidental presence of *P. ludlowiensis* (Ramsåsa SMNH-Q 607). In the Baltic region the latter species is known from Whitcliffian levels both older and younger than the lowermost Tahula beds, whereas in the Welsh Borderland it is also known from levels dated as Přidoli (Miller and Märss, 1999). *Thelodus carinatus* is as old as, or older than, *A. hedei* and the presence of *T. carinatus* (-like) scales (Ramsåsa D/SW24) might also be age-related, although most of the scales figured by Vergoossen (2002c) will be referred to a new *Thelodus* taxon. Among the thelodonts, other potential biostratigraphical markers are several new form groups from the smallest size fractions, at first interpreted as form variants within known species (e.g Vergoossen 2002a: fig. 49; 2002c: fig. 81) but soon to be described as new *Thelodus* taxa. Among the acanthodians in the faunal lists no taxon can at present be linked up with age-related differences in the composition of the faunas because the species associated with *T. sculptilis* in the reference successions must be revised. The most important of these is '*Poracanthodes porosus* Brotzen', which was presented as an acanthodian zonal fossil and deep shelf equivalent of *T. sculptilis* by Märss (1997, 2000), but '*P. porosus*' in its current sense comprises scales from several taxa with different geographic distributions within Laurussia. The porosiform scales originally described as such were obtained from a Lower Devonian erratic conglomerate. *Radioporacanthodes porosus* (Brotzen) s.s as defined by Vergoossen (1999) was not observed in the ÖSF faunas, but some of the form variants grouped together in the oldest occurrences of '*P. porosus*' have been found and redescribed as *Radioporacanthodes biblicus* and forma *bifurcatus*, partim a new *Radioporacanthodes* species (Vergoossen, in press). It is likely that the zonal fossil '*P. porosus*' also comprises eroded scales of the presumably punctatiform *Poracanthodes*? *lehmani*, which is present both in the lowermost Tahula beds and in some of the ÖSF faunas. Punctatiform morphs, partim published as cf. *Poracanthodes punctatus*, were also listed from the ÖSF faunas, but have not yet been studied, because in the rocks from Ramsåsa their preservation is poor; the best material for such study appears to be present in the Helvetesgraven fauna, which needs to be described in detail. *Gomphonchus volborthi* is a potential biostratigraphic marker that disappears towards the Mid Přidoli. It is absent from the Helvetsgraven fauna. When *G. volborthi* and *G. sandelensis* occur together in the Ramsåsa faunas, the first is dominant and represented in all size ranges, whereas the second was chiefly collected from the smallest size ranges and difficult to identify (except in the fauna from Ramsåsa SMNH-Q607). The morphological analysis of trunk scales of *Nostolepis striata* and allied species (Vergoossen, 2002b–c; in press) will contribute to the revision of *N. striata*, which in the ÖSF rocks includes forms (esp. in the 0.106 and 0.212 fractions, and from Ramsåsa E) that show strong morphological resemblance to the Lochkovian acanthodian zonal fossil *N. minima.* Scales restricted to Ramsåsa C and described as Acanthodii indet., morph 5 (Vergoossen, in press) might also be suited as markers.

Faunal dominance of thelodonts over acanthodians, or vice versa, has been attributed to water depth, with the deeper shelf as the realm where the acanthodians dominate, and shallower waters for the dominance of the thelodonts (cf. e.g. Märss 1997). Such faunal interpretations might help explain some of the faunal changes listed for the ÖSF assemblages. In the fauna from Helvetesgraven thelodonts and acanthodians are in balance, with abundant *T. admirabilis*, common *Loganellia cuneata,* regular *Katoporodus tricavus* (new record), a fair number of the punctatiform poracanthodids (esp. *Poracanthodes lehmani*) and with *Gomphonchus sandelensis* but without *G. volborthi*. In the faunas from Ramsåsa (C, D, E, H) the thelodonts dominate, with fewer or no *T. admirabilis,* very scarce, small, poorly preserved loganellids (both *Loganellia cuneata* and *Paralogania ludlowiensis*), no *Katoporodus tricavus*, with *Gomphonchus volborthi* outnumbering rare, small, and poorly preserved *G. sandelensis*, and very scarce, small, poorly preserved porosiforms and punctatiforms. The Peyel residue and Ørvig slides, which hold robust, well-preserved, large acanthodian remains from Ramsåsa E, might indicate a phase with deeper water when acanthodians flourished. A most interesting fauna in this respect was obtained from Ramsåsa rock SMNH- Q607 (precise locality unknown) with dominance of thelodonts, but yielding more and better preserved acanthodians than the rocks from Ramsåsa sites C, D, E, H. This fauna might take up a position intermediate between those from Ramsåsa (C, D, E, H) rocks and that from Helvetesgraven, with dominance of *T. parvidens* s.s., regular loganellids (both *L. cuneata* and *P. ludlowiensis*-the latter rather well-preserved but chiefly in the 0.106–0.212 size ranges), with occasional *Katoporodus tricavus*, with *G. volborthi* outnumbering *G. sandelensis* (which is better preserved and more frequent in Q607 than in any other Ramsåsa rocks), with higher numbers of porosiforms (esp. *Radioporacanthodes biblicus*) and punctatiforms. The most striking observation, however, was that the increase in *P. ludlowiensis* scales is accompanied by a sharp decrease in the scales and form variants of the zonal fossil *T. sculptilis*: Q607 holds the relatively lowest number of *T. sculptilis* scales of all the Ramsåsa faunas examined by the author so far. Faunal composition of the ÖSF rocks (proportion of thelodont to acanthodian scales; proportion of porosiform scales) suggest deeper sea water for the Helvetesgraven fauna than for most of the Ramsåsa faunas. Within the faunas this difference (probably in combination with other factors) might also be reflected at the thelodont and acanthodian species levels: for example, the proportion of *Katoporodus tricavus*, *Loganellia cuneata*, *Paralogania ludlowiensis* to other thelodont scales, and in particular the proportion of *Thelodus sculptilis* scales to *P. ludlowiensis* scales; absence or relative frequency of *T. admirabilis* scales; proportion of *Gomphonchus volborthi* to *G. sandelensis* scales etc.

Quick comparison with *Thelodus sculptilis* faunas from Baltic-derived erratic algal limestones (*Sphaerocodium/Girvanella*, a shallow water indicator) shows that they share the similarities and differences between the studied Scanian faunas to a greater or lesser extent, except in the following aspects: 1) At the species level *Andreolepis hedei, T. admirabilis, T. carinatus*-like scales, *T. traquairi* and *Poracanthodes? lehmani* (and probably punctatiform taxa) are absent. Osteostracan remains were not observed either: these are known from all the Scanian sites. 2) *T. sculptilis* is never as frequent as it can be in the Scanian faunas. 3) Particular form variants can be restricted to either the erratic or the Scanian faunas. The erratic faunas with *T. sculptilis* were interpreted as *T. sculptilis* zone faunas, judging from the component taxa (unpublished).

A morphological feature restricted to forma *Thelodus trilobatus* scales from the Ramsåsa C fauna is the presence of double median basal spurs, which might reflect a particular environmental response (e.g. to high energy waters).

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APPLICATION OF STATISTIC METHODS TO LITHOLOGICAL CORRELATION OF GLACIAL TILLS

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In many cases lithological features of Quaternary sediments – first of all of tills – are used as helpful criteria for correlation of lithostratigraphical units. Among the methods applied to tills, petrographic analysis of gravel and boulders (erratics) has a principle meaning. Establishment that ice sheets deposited material (as tills) transported from various regions of Scandinavia (alimentary area) and results in variety of petrographic spectrum of erratics make these features useful for correlation of tills. The differences in petrographic spectrum of gravel have more quantitative then qualitative meanings, so application of statistic methods for analysis of these differences is necessary. Quantitive differences are the base for creation of the so-called lithotypes of tills. Lithothypes, which are characterised by similar petrographic spectrum and occur in logical superposition can be correlated with one another (applying the method used in the Polish Geological Institute, the petrographic features of gravel 5–10 mm are presented by the so-called petrographical coefficients: O/K, K/W, A/B).

The presentation contains description of petrographic features of tills occuring in the Brown Coal Mine Konin (KWB Konin) at outcrops: Pątnów, Lubstów and partially Jóźwin. In this area there are 3 to 5 till horizons (T-1, T-1A, T-2, T-3, T-4), commonly correlated with different glaciations. This tills are characterised by different petrographic spectrum of gravel in fraction 5–10 mm and in fraction > 20 mm (Stankowski, Krzyszkowski 1991; Czubla 2001).

Petrographic spectrum of gravel (5–10 mm) was statistically analysed in some aspects:

- analysis of distribution frequency of main rock groups and spectrum of value of petrographical coefficients in tills belonging to the same and different stratigraphical units (*histograms*);
- relations between main rock groups (*statistic tests*).

This type of presentation gives better comparison of petrographical features of tills and presents a lot of aspects of petrographic analyses.

The comparable features of petrographic spectrum of gravel are noted for samples from the till T-2 (the main till in study area), which occurs in the same superposition in the examined outcrops. Distribution frequency of main rock groups (for ex.: Kr – crystalline rocks, Wp – palaeozoic limestones; Dp – dolomites) and value of petrographical coefficients are comparable. There are no significant differences between proper petrographical features of gravel (Fig. 1). The same petrographic features are not comparable between samples from three separate tills (for example: T-3 and T-4 – lithotypes belonging to different glaciations) at the Lubstów. Differences in this case have signified statistic meaning – it is very important for stratigraphical conclusions (Table).

Table. Test U Manna–Whitneya for tills T-2 and T-4 at Lubstów.

$\alpha = 0.05$	Sum of ranges		test U	Z (N1+N2>20)	level p	Z corrected	level p corrected	nr of obs.	
	T-4	T-3						T-4	T-3
KR%_rang	135	141	36	1.7	0.08898	1.70084	0.08898	9	14
WP%_rang	94	182	49	-0.9	0.37783	-0.88192	0.37783	9	14
DP%_rang	45	231	0	-4.0	0.00007	-3.96863	0.00007	9	14
O/K_rang	58	218	13	-3.1	0.00164	-3.14970	0.00164	9	14
K/W_rang	162	114	9	3.4	0.00067	3.40168	0.00067	9	14
A/B_rang	52	224	7	-3.5	0.00042	-3.52767	0.00042	9	14
Dp/Wp_rang	45	231	0	-4.0	0.00007	-3.96863	0.00007	9	14
for level p < 0.05 no correlation between groups									

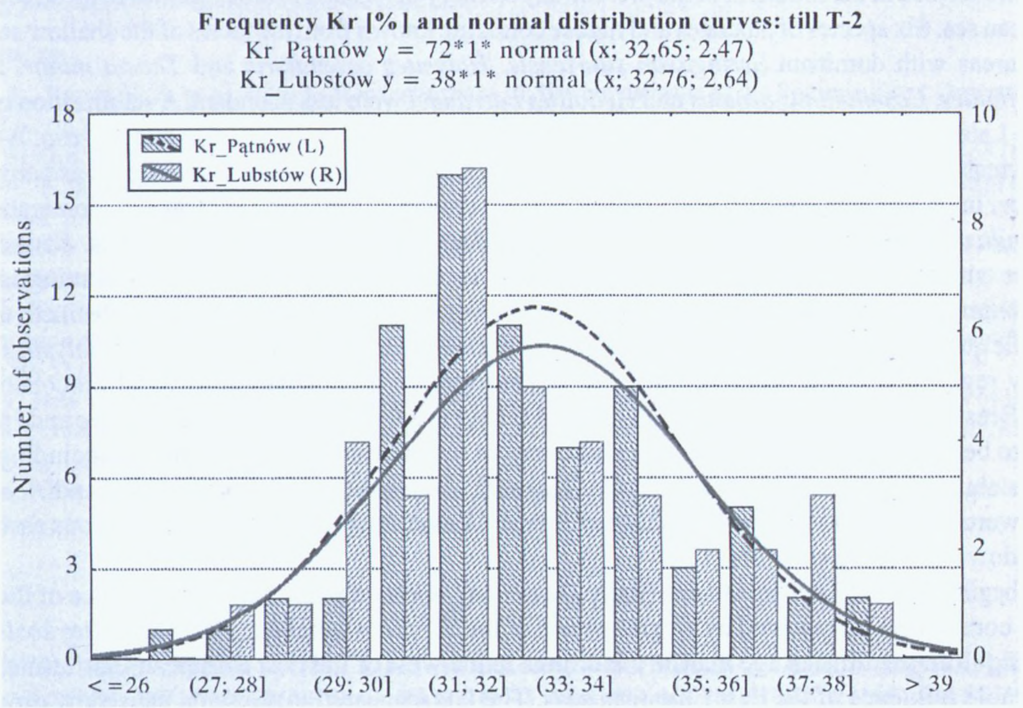

Fig. 1. Frequency and normal distribution curves for crystalline rocks: till T-2; outcrops: Pątnów [Kr_Pątnów] and Lubstów [Kr_Lubstów].

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MONOFACIES COMMUNITIES OF EARLY AND MIDDLE PALEOZOIC CORALS FROM BELARUS

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During the Ordovician, Silurian and Devonian marine basins situated within the present-day territory of Belarus had been repeatedly populated with coral communities. Epochs of extensive distribution and disappearance of coral associations are well corellated with paleogeographic changes in sedimentation basins.

The first appearance of corals took place in Oandu time (the Middle Ordovician, Caradoc) in the north-western part of Belarus. This was associated with an eastward transgression in the so-called Baltic structural bay of the European marine basin. A community of rugose corals and tabulates immigrated there and inhabited the littoral shelf zone. Their taxonomic structure has yet to be studied.

The basin shallowing developed later on (in Ashgillian time) stimulated a stormy growth of corals, that was indicated by their active participation in creating bioherms. Species predominating in Ashgillian complexes are as follows: *Acidolites medialis* and *Wormsipora miranda* both from the Wormsi beds and *Pseudocosmiolitus varians* from the Porkuni beds. All of them belong to the *Heliolitida* subclass. Tabulates of *Sarcinula organum* were also met in the Pirgu formation (Upper Ordovician).

Fundamental changes of benthic associations took place in the Early Silurian as a result of transgession. The corals became extinct in the Belarussian part of the Baltic bay, but during the Late Wenlockian some monofacies communities formed in the southeast of the present-day territory of Belarus in the Podlasie–Brest structural bay of the European sea. Six species of tabulates and rugose corals are known from the facies of the shallow sea zone and open sea areas with dominant *Syringaxon siluriensis, Halysites catenularis* and *Thecia minor. Phaulactis cyathophylloides, Cosmiolithus ornatus* and *Heliolites barrandei* were less abundant. A stabilization of the basin during the Late Ludlowian caused an essential increase in diversity of both *Tabulata* and *Rugosa.* A number of species increased four times to as many as 26 species and a number of families was doubled as compared with their number in Wenlockian time. An appearance and rapid drowth of favositid *Tabulata* with dominant *Favosites subgothlandicus* as the main species is the charactaristic feature of Ludlowian communities. *Favositidae* with some other tabulates, mainly *Syringopora affabilis* and *Stromatoporoidea* were of rockforming importance. Another community of the same age which inhabited the deeper parts of the basin had some distinctive properties shown in the quantitative prevalence of halysitoid *Tabulata: Halysites catenularis* and *Cystihalysites mirabilis.*

A slow regression of the basin in the Přidolian resulted in a degradation of coral biocoenoses in the Podlasie–Brest bay. A number of families decreased to four involving 7 species. *Halysitidae* and *Heliolitidae* appeared to be mostly able to survive among tabulates. A new family of *Multisoleniidae* including 3 species was rather abundant. Some of *Ptenophyllidae (Rugosa): Spongophylloides nikiforovae* and *Acmophyllum armatum* were also found sometimes in this community. These formed a rather homogeneous association as coral meadows overlying soft silts [1].

At the begining of the Lochkovian age (Early Devonian), during the last stages of existence of the Podlasie–Brest bay, corals were represented by single species of the *Tabulata* subclass*: Pachyfavosites kozlowskii* [2].

Begining from the Emsian age marine conditions in the west of the East European Platform came under a considerable influence of the East European seas. The late Emsian transgression caused by downwarping of the western part of the Moscow syncline involved eastern and central regions of the present-day territory of Belarus, where sediments of the Middle and Upper Devonian are widespread in the west of the Main and the Central Devonian Fields and within the Pripyat Trough.

In addition considerable changes of sedimentary processes occurred, and conditions favourable for coral existence appeared at the Late Eifelian (Middle Devonian) only. Single corals which species belonging has not been determined are known from limestones of Kostjukovichi horizon. Later on, under the influence of persisting transgression and as a result of river water freshening the normal marine conditions were disturbed and were not reestablished until the Middle Frasnian (Upper Devonian). Carbonate silts were deposited at that time in a well aerated and warm shallow water with normal salinity within the Orsha Depression and the Pripyat Trough. Corals made up the major portion of benthic communities and formed two monofacies groups. The first group involved massive and incrusting tabulates: *Crassialveolites* and

Alveolites, forming biostromes and bioherms with some stromatoporoids [3]. *Alveolites suborbicularis, Crassialveolites domrachevi* and *Cr. Multiporosus* [4] contributed mostly to these structures. Some representatives of *Aulopora* were determined as symbiotic organisms of stromatoporoids.

Ramose tabulata of *Thamnopora cervicornis, Th. polyforata, Th. reticulata, Scoliopora denticulata* and some other coral species inhabited all the space between bioaccumulated structures. Rugose corals of *Peneckiella minima, Thamnophyllum monozonatum* and some rarer representatives of this subclass inhabited similar places, but were less abundant [5].

New transgressive processes and an increased tectonic activity, especially within the Pripyat Trough in the Middle Frasnian resulted in a disappearance of the *Rugosa* and *Tabulata* community. It was replaced in Voronezh time by some new species migrated from the central part of the East European Platform. Genera of *Thamnopora, Alveolites, Crassialveolites, Scoliopora* and *Tabulophpyllum* became completely extinct. Tabulates poor in species, but reach in specimens, such as: *Aulopora soshkinae* and *Aulocystis tikhyi* became dominant [6]. *Rugosa* genera somewhat increased in number, and *Thamnophyllum monozonatum* and *Peneckiella minima* preserved from the previous association became their dominant species.

A decline of the benthic biocoeneses was associated with a transgression in Late Yevlanovo–Liven time, which resulted in lagoon formation and salt accumulation. The last Late Devonian corals in the Pripyat Trough were represented by rare *Aulopora soshkinae (Tabulata)* and not numerous Rugosa, mainly of *Thamnophyllum trigemme* [7].

A conclusion may be drawn that coral faunas can serve as important indicators of evolution of sedimentation basins, and when compared with changes in the taxonomic structure of corals from adjacent areas, may provide a good support to stratigraphic corellation.

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SEDIMENTARY CONDITIONS OF LATORPIAN–KUNDAN STAGES (LOWER–MIDDLE ORDOVICIAN) OF THE CENTRAL PART OF LADOGA KLINT (NW RUSSIAN PLATFORM)

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Though the Ordovician system of the North-Western part of the Russian Platform has been studied since the beginning of the XIX century [5], there are still many points to be studied and discussed. One of them is depositional environments and the paper is focused on this problem.

In the Early Paleozoic this territory has been covered by epicontinental basin with a flat bottom the land was rather distant. There are many controversial questions: the bathymetry of the basin, its hydrodynamics, temperature, salinity, the position of the coastline, etc. Newly received integrated data on the lithology, sedimentology, and paleoecology allow to solve some of these problems. For the current study we used the stratigraphical scheme by A. Yu. Ivantsov [4].

The material was collected in the sections of the central part of the Ladoga Klint. The succession of the Upper Latorpian, Volkhovian (Lava River) and Kundan (Lava River, Putilovo quarry) stages were studied bed-by-bed. Over 170 thin sections had been studied, about 100 samples of the clay fraction were analysed by the X-ray method.

11 lithotypes of carbonate rock in the sections of Volkhovian and Kundan stages were recognised by the differences in the content of the bioclasts and pelitic material; bioclast composition, texture and structure and by the presence of different types of glauconite and limonite microconcretions. The changes in lithotypes corresponds with the transgression conditions upwards through the section. It is expressed in changes in the glauconite and bioclast types. There are four of them: I – Syngenetic glauconite of rounded, elongated or "palmate" shape. The conformity of grains and bioclasts and their hard cracking demonstrates the authigenetic origin and formation in the sediment. This glauconite type is widespread in the Volkhov and Lower Obukhovo formations. II – Fragmental redeposited glauconite with angular grains. It occurs in the lower and upper parts of Volkhov Stage and the lower part of Kunda Stage. III – Glauconite connected with the Fe redistribution in the slightly lithified sediment and represented by pseudomorph after bioclasts. It was found in the Volkhov Stage. Its origin connected with the echinodermal skeletons, which are highly porous and can maintain an organic matter. IV – Glauconite, with incrusts hardground (HG) surfaces developed in Volkhov Stage and to a lesser extent in the lower part of Kunda Stage.

The clay fraction in the carbonate rocks is represented by a hydromical mixed-layered association with a small chlorite and caolinite content. The regularity of its composition demonstrates constant basin width during the studied interval. The absence of terrigenous minerals shows that the coast line was distant.

Integrated analysis of the new data allows characterise the style of sedimentation in the basin during the Volkhov–Kunda time. It was the period of the Baltic paleobasin transgressive cycle [1, 6, 7], during which the transgressive deeper-water Volkhov and Kunda deposits were interrupted by a short-term shallowing. Both phases are well-defined by lithological and mineralogical peculiarities.

The beginning of the Latorpian–Volkhovian time was characterised by the shallowest conditions with active hydrodynamics. The latter is revealed by the presence of gradational structure in the "Zeleny" Bed [3] and upwards in the "Dikari" Member [2] and by a presence of terrigenous material (Paite Member) in the top. The period corresponds to 3 transgressions represented in the succession by three members of limestone/clay alternation that normally [3] replace each other.

The presence of oblique sloping microlayering in combination with graded structures and II-type glauconite indicates sediment rewashing but this is rather an exception. In other cases the rock is hardly bioturbated, bioclasts are randomly oriented and the primary structures are poorly preserved. The ability of glauconite is more significant in "Dikari" Member. Here all types of glauconite are present. The lower part of the succession was formed in a shallow conditions at the basis of wave-cut and far-off the shoreline. The accumulation prevailed and rarely interrupted by the HG formation. Bioturbation of the carbonate sediment led to its degasation and rapid subsequent cementation formed the HG. Followed period of HG exposition led to its destruction (dissolution, bio- and mechanical erosion). During the new sedimentation phase HG has been covered by friable sediment and IV-type glauconite grew on the newly formed geochemical barrier.

In the late Volkhovian time the sea-level rise effected in the carbonate accumulation/dissolution with frequency decreases downwards the section. The mature HG did not form. At the end of the period thin carbonate and clayey material sediments in accordance with the maximum of the basin depth. Fragmental glauconite and mature HG-forming demonstrates shallowing of the basin on the Volkhovian/Kundan boundary.

The development of the paleobasin in the Kundan time includes 2 phases: (1) Sillaoru–Obukhovo phase and (2) Sinjavino–Simankovo phase. Both of them have transgressive character and are bordered by a short-term sea-level drop. It is confirmed by the presence of limonite microconcretions and numerous HG surfaces incrusted by brachiopods. Each phase is characterized by a gradual sea-level rise revealed by decrease of glauconite content, increase of pelite material content, gradual change of the composition and increasing of the bioclast size. The beginning of the Kundan time is characterized by insignificant quantity of echinoderms (Cystoidea, Crinoidea, Eocrinoidea, etc. – 10%) and large quantuty of ostracodes (50%). Afterwards the content of echinoderm bioclasts is shortly increased and then gradually decreased during the Kundan time.

Such dependence is connected with the changes in the substrate conditions, on which the echinoderms settled: their maximum corresponds to HG development levels.

At the beginning of the Kundan time a large quantities of Fe colloids were transported from the dry land to the axial part of the basin. In intensive red-coloured rocks were formed in front of the sourceland (the Belorussia Land?). Fe is more oxydised here than in the far-off parts, where limonite microconcretions have formed. We accept the Belorussia Land as a possible sourceland since the character of the lithofacies distribution implies the marine basin existence on the south-eastern borderland of Baltic Shield, where the Ordovician deposits are not preserved now. In the Leningrad Region near the Ladoga Klint the basis of the storm waves could reach the bottom. Westwards Saint-Petersbourg and on the greater part of Estonia this interval is characterized by "limonitisied" bioclasts abundance, with core the evidence for the shallowness of the basin and active hydrodinamics. On the eastern borderland of Baltic–Ladoga Klint (Volkhov and Lava rivers) terrestrial Fe content was very low because of greater distance from the sourceland. Here the interval doesn't contain any limonite concretions, which appear above and can be traced only in some places. The conditions were the same in the middle of Kunda Stage (Sinjavino Formation), but it was less spread and limonite concretions were extend only up to the central part of Ladoga Klint.

Thus the Lower–Middle Ordovician carbonate deposits were formed during several transgression phases. The absence of terrestrial input demonstrates the distant coastline and basin width stability during that time.

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LOWER SILURIAN (UPPER LLANDOVERIAN) THELODONTS (AGNATHA) OF NW MONGOLIA

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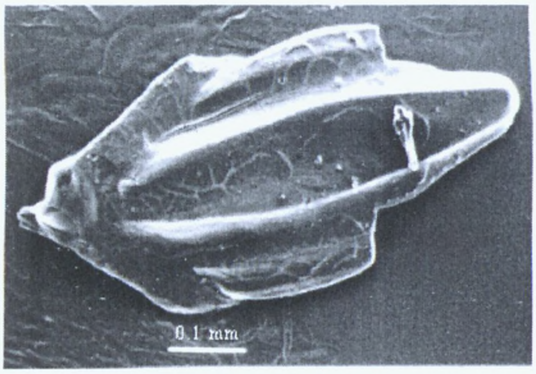
Silurian shallow water sediments of internal North Mongolia paleoland continental shelf are widely distributed facies of North-West Mongolia [2]. The microremains of vertebrates yielded in these mainly terrigenous sediments are presented of Thelodonti, Heterostraci, Anaspida (Agnatha), and Mongolepidida, Elasmobranchii and Acanthodii scales and tesserae.

Two samples of the same Upper Llandoverian age were analysed. The first sample explored by K. S. Rozman, have more than 500 microremains [1]. The second one, gathered by P. Kosbayar & I. Sodov, contains a great number of microremains (over 10 thousand exemplars) and is discussed in this paper.

Thelodonts in the latter sample are presented in two genera – *Loganellia* and *Helenolepis*. The main part of thelodont scales belongs to *Loganellia* sp. nov. (at least 400 scales). There are scales from different parts of the body among them. Head, trunck scales (see Fig. 1 A, B), thin and prolongated fin scales are chosen, and quite full morphological set can be prepared. The scales of *Helenolepis* morphological type are fine and not abundant (no more than 100 exemplars found).

The complex of vertebrates of North-West Mongolia share close resemblance with complexes of vertebrates of the Lower Silurian (Middle to Upper Llandoverian) of Siberian Platform (Irkutsk Amphitheatre) and Tuva [3] as well.



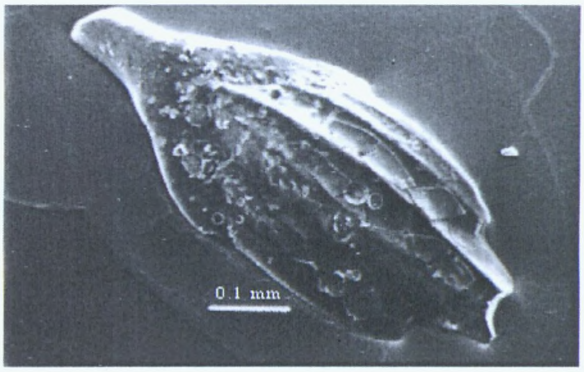


Fig. 1 A, B. Loganellia sp. nov. Trunck scales, dorsal view. NW Mongolia, Khutsynbulak beds, Lower Silurian, Upper Llandoverian.

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