WOGOGOB-2004 ESTONIA May 13–18

WOGOGOB-2004 Conference Materials

edited by O. Hints and L. Ainsaar

WOGOGOB-2004
8th Meeting on the Working Group on the Ordovician Geology of Baltoscandia
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Preface

Since 1985, when Prof. Maurits Lindström first presented the idea of WOGOGOB, seven conferences on the Ordovician geology of Baltoscandia, together with associated excursions have been held in Sweden (1986, 1990), Estonia (1988), Norway (1992), Denmark (1994, 2001) and Russia (1997). The number of participants in these meetings has increased from a few tens to about 60 in the Copenhagen meeting in 2001. Also the scope of the meeting has widened notably during these years, both in terms of geographical coverage and research fields.

The Eighth WOGOGOB Meeting is the second one held in Estonia. Geological development of the region, palaeogeography, stratigraphy, palaeoceanography, different aspects of life and other problems related to the Ordovician System will be considered in the proposed schedule of the meeting. The working groups from all around the Baltic are represented in the list of participants, but seemingly the meeting is gaining wider popularity as well.

More than 50 delegates have registered for the meeting and excursion in northern and central Estonia. On the two-day technical session at the University of Tartu 40 oral and 15 poster presentations by delegates from Denmark, Sweden, Norway, Germany, Russia, Great Britain, USA, Poland, Lithuania and Estonia will be held. The conference talks and the volume can be considered as contributions to the new IGCP project No. 503 "Ordovician Palaeogeography and Palaeoclimate".

During the one-day pre-conference and two-day post-conference geological excursions, the participants will be shown a number of Ordovician key sections. The site list also includes the section in the Pakri Peninsula, which was not visited during the previous meeting. This locality was not accessible for geologists over several decades, as it was located in the area controlled by the Russian Army. Also a new and fresh section of kukersite oil shale deposits in NE Estonia will be visited.

We are looking forward to hosting you in Estonia, EU (since 01.05.2004), and hope that you will have a great time here.

The organisers are grateful to the Environmental Investment Centre of Estonia (KIK) for the financial support.

The Organisers

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A new IGCP project No 503: Ordovician Palaeogeography and Palaeoclimate (2004–2008)

In October 2003, we proposed a new project, entitled "Ordovician Palaeogeography and Palaeoclimate: The impact of the changing palaeogeography and palaeoclimate on the major biotic changes through the Ordovician (Ordovician biodiversification, end-Ordovician extinction, Silurian radiation)" to the board of the International Geological Correlation Programme (IGCP). This new project was accepted in February 2004 and designated IGCP No 503. The project will run for five years from 2004 to 2008.

Arguably the most sustained rise in marine biodiversity took place during the Ordovician and the second largest mass extinction event occurred close to the end of that Period, coincident with an episode of major climate change. The results of the very successful IGCP project No 410 "The Great Ordovician Biodiversification Event" not only included the development of an improved globally-integrated biozonation for graptolites, conodonts and chitinozoans, but also generated biodiversity curves that have been constructed for all Ordovician fossil groups. Numerous questions arise from these results, for example how has changing palaeogeography affected the biodiversifications observed during the Ordovician, the extinctions at the end of the period, and the subsequent radiation in the Silurian. What was the influence of the climate on the major biotic changes through the Ordovician and Silurian? Following the work of the numerous regional teams and of the clade teams, that were established for each fossil group in IGCP project No 410, the new successor project should attempt to answer some of these questions (and generate others), and develop a better understanding of the environmental changes, e.g., global sea-level, temperature, oceanic circulation, climatic thresholds, etc. during the Early Palaeozoic. The new project will be developed in collaboration with the Subcommission on Ordovician Stratigraphy (SOS) together with the Subcommission on Silurian Stratigraphy (SSS).

The following work plan is proposed: (1) the first year of the project will culminate in an assessment of all available information on ocean and climate modelling, as well as data from stable C- and O-isotopes in the Lower Palaeozoic; (2) the second year will focus on the evolutionary palaeoecology of the Early Palaeozoic, in order to understand the architecture of the ecosystems that developed and changed during the Ordovician biodiversification and extinction, and during the subsequent Silurian radiation and extinctions; (3) in the third year, relationships will be sought between the changing palaeogeographical patterns and the changing biodiversities observed during the investigated intervals; (4) in the fourth year, we plan to compile all information on Early Palaeozoic events and stratigraphy, in order to improve the international standard time scale; (5) in the last year of our project, all information collected from the different regional teams should allow us to reconstruct Early Palaeozoic sea-level changes. At the end of our project, we also plan to produce a final synthesis bringing together the varied facets of the project and how they have addressed the objectives.

Five major meetings are scheduled between 2004 and 2008. Each of these meetings comprises indoor sessions, including the presentation of new results, and field excursions: In 2004, the official opening meeting of the new IGCP project will be organised at the Universität Erlangen, Germany, September, 1-3, 2004, just following the International Geological Congress at Florence, Italy. This meeting will be focused on Early Palaeozoic Climate Modelling and isotope geochemistry. The associated geological excursion will cover the Ordovician of Öland and the Silurian of Gotland (Sweden), September 4-12, 2004. The second major meeting will be held in 2005 at the Milwaukee Public Museum (Wisconsin, USA) with a focus on the ecological evolution in the Early Palaeozoic. The geological excursions will be organized to the Ordovician of the American mid-Continent (Kentucky, Ohio, etc.). We plan to organize this conference just before or after the North American Paleontological Convention meeting. In 2006, a major meeting of the new IGCP project will be held at Glasgow University, Scotland, UK, and concentrate on palaeo(bio)geography, and the relation between Early Palaeozoic biodiversity and climatic belts. A geological excursion to the Lower Palaeozoic of Scotland and northern England will include a visit to the GSSP of the base of the Silurian at Dob's Linn and to Scottish Silurian inliers in the Southern Uplands and Pentland Hills. The Nanjing Institute of Geology and Palaeontology, Academia Sinica, China, will organize the 10th ISOS in 2007. Jointly with the Ordovician Meeting, the Silurian Subcommission will also hold its International Symposium. An IGCP related symposium on Early Palaeozoic events is scheduled at Nanjing, as well as several geological excursions to Lower Palaeozoic sections of southern China, to Ordovician to Early Silurian sections of the Upper Yangtze Platform. The closing meeting of our project should take place in 2008 at the Université des Sciences et Technologies de Lille, France. This meeting is dedicated to the reconstruction of Early Palaeozoic sea-level changes. Geological excursions to Lower Palaeozoic sections of France and Belgium will be organised.

A series of additional meetings are already scheduled (Argentina, Spain, Czech Republic, Estonia, etc.).

We warmly invite all Ordovician and Silurian workers to join IGCP 503.

T. Servais,
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J. Li,
A. Munnecke,
A.W. Owen and
P.M. Sheehan

More information on the project and the Opening Meeting in Erlangen is available at http://www.pal.uni-erlangen.de/IGCP503/

Abstracts

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Middle and Upper Ordovician stable isotope stratigraphy across the facies belts in the East Baltic

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Trends in the oceanic carbon stable isotopic composition ($\delta^{13}C$) are considered to be an indicator of different oceanographic and environmental changes. Stratigraphic variations in the carbon isotopic value of marine carbonates have become an important tool in regional and global correlation of sedimentary successions. Here we present new carbon isotope data from the Billingen to Porkuni stages (Hirnantian) from the Kadriorg outcrop, and Männamaa (F-367), Ruhnu (500) and Jurmala (R-1) drill cores. Our stable isotope data set covers almost completely the Middle and Late Ordovician, altogether 30 million years. Using these data in combination with published materials, we can distinguish at least six positive carbon isotope excursions, which can be followed in more than one section in Estonia or Latvia. All of them, except for the end-Ordovician $\delta^{13}C$ peak, have a relative value around 1–2‰.

1. The Middle Darriwil excursion (see Meidla *et al.* this volume) is documented from the Segerstad Formation (Fm) in the Ruhnu and Jurmala core sections. The Segerstad Fm is correlated with the Aseri Stage (Nõlvak 1997). The Middle Darriwil excursion has not been found in northern Estonian sections (Männamaa, Kadriorg), probably because the boundary beds of the Kunda and Aseri stages are missing in the middle–upper shelf facies. Thus, the exact stratigraphic age of the excursion (late Kunda or Aseri) remains unknown.

2. The Middle Caradoc excursion is documented from the upper part of the Kahula Fm in the Ristiküla and Tartu drill cores in southern Estonia (Ainsaar *et al.* 1999). It has late Keila age and correlates with the sedimentary gap on the boundary of the Keila and Oandu stages in the upper shelf sections (e.g. Männamaa). The excursion is weakly expressed in the basinal sections (Blidene Fm in the Valga, Jurmala and Ruhnu core sections), probably due to condensed sedimentation and low carbonate content of sediments.

3. The Early Late Caradoc excursion is documented from the lower part of the Rägavere Fm, Rakvere Stage, in the Rapla (Kaljo *et al*. 1999) and Männamaa sections. The excursion is well developed in micritic limestones of the middle–upper shelf, but not in basinal sections, probably due to condensed sedimentation and prevalation of clay-rich sediments.

4. The Late Caradoc excursion is documented from the Saunja Fm (upper Nabala Stage) in the Ristiküla, Valga, Jurmala, Ruhnu and Männamaa sections. Like the previous one, it is developed in micritic limestones, but can be traced across the whole facies profile.

5. The Early Ashgill excursion is documented from the lower part of the Jonstorp Fm in the Kaugatuma (Kaljo *et al.* 1999), Ruhnu and Jurmala sections, and from the lower part of the Moe Fm in the Rapla (Kaljo *et al.* 1999) and Männamaa sections. It has early Pirgu age and can be also traced across the whole facies profile.

6. The end-Ordovician (Hirnantian) excursion, a prominent glaciation-driven isotope event, is found in many sections in the area (Kaljo *et al.* 2001; Brenchley *et al.* 2003). The comparison of the Jurmala, Ruhnu and Männamaa curves shows an increase in the end-Ordovician stratigraphic gap towards the upper shelf. The Jurmala section is probably the most complete section of the Ordovician–Silurian boundary beds in the area, being closest to the Hirnantian $\delta^{13}C$ composite model of Brenchley *et al.* (2003).

The Ordovician carbon isotope excursions have probably different origins. Some of the excursions (Middle Darriwil, Middle Caradoc, end-Ordovician) can be correlated with sedimentary gaps in the upper shelf and may be related to (glacioeustatic) sea level falls. Others seem to be related with sea level highstands or transgressions and need different environmental interpretations.

The study was supported by the Estonian Science Foundation Grant 4574 and Paleostudies Programme at the University of Bremen.

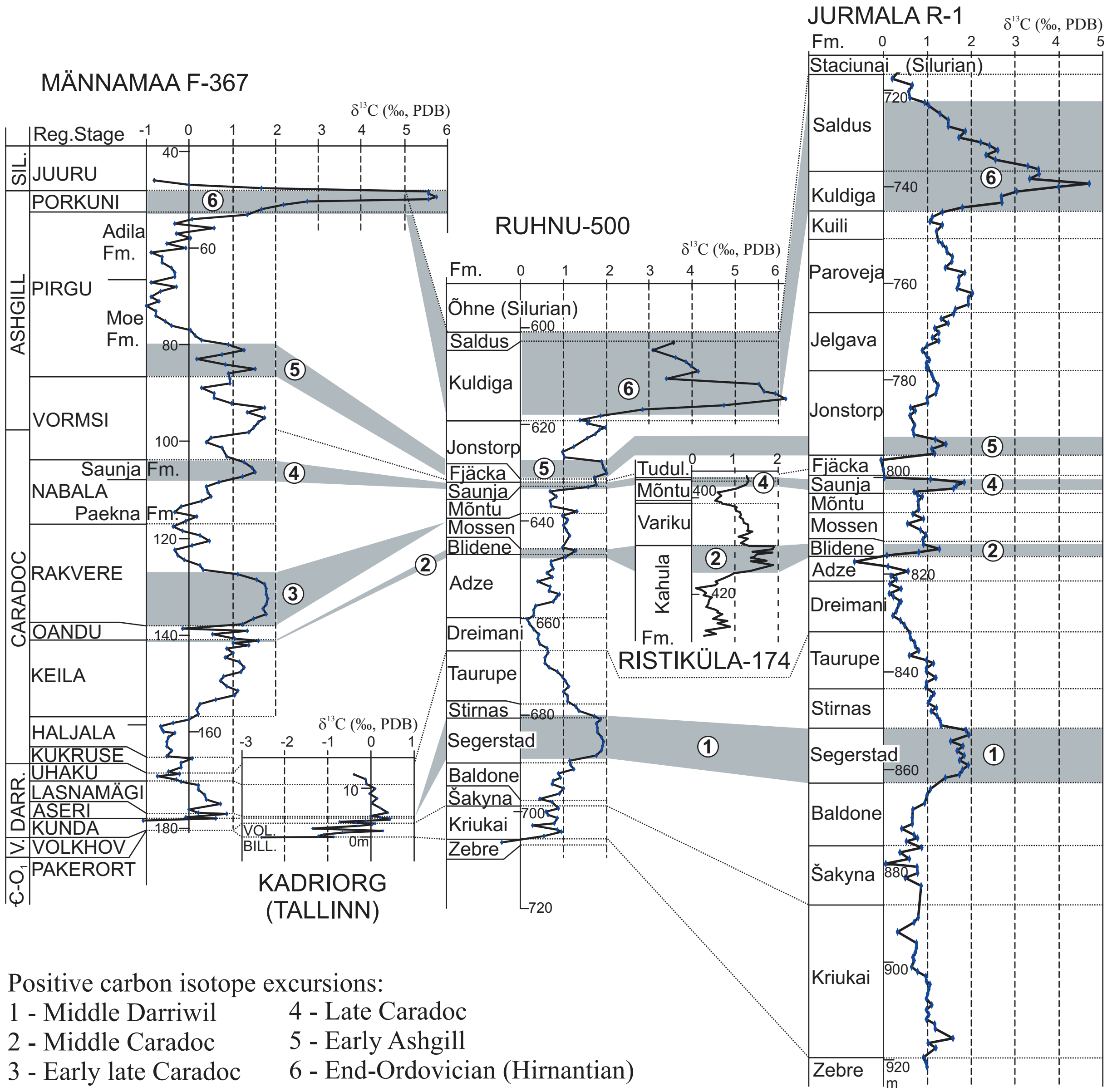

Fig. 1. Stable carbon isotope stratigraphy in the East Baltic area.

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Upper Ordovician in the Chernyshev Swell, Timan – northern Ural region

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The Chernyshev Swell is a large linear-folded structure delimiting the northern Pre-Urals Foredeep in the west. It extends from the western slope of the Subpolar Urals in the south-west to the Chernov Swell in the north-east and represents an important area between the margin shelf and foreshore environments in the Timan–northern Ural sedimentary palaeobasin.

The Upper Ordovician (=Ashgill) of the West Urals and Pre-Urals Foredeep includes the Rassokha, Polydov, Syr'ya, and Kyr'ya regional stages which correspond to the Zyb (or Ust'Zyb), Malaya Tavrota, and Yaptikshor formations. Beginning from the middle Caradoc, carbonate sedimentation occurred on the passive margin where shelf and continental slope zones are distinguished. Source areas were located to the west of the considered region where the Timan Ridge is presently situated. From west to east, a transition from inner to outer carbonate ramp environments is recognized. At that time terrigenous-carbonate successions formed over terrigenous shelf facies and the platform-to-basin transition was initiated.

In the late Ashgill (Syr'ya time=Upper Malaya Tavrota Formation) the former ramp transformed into the epicontinental platform. At the edge of an incipient late Ashgill evaporitic shelf margin grew the first Palaeozoic reefs (up to 400 m thick). A sea level drop in the middle of the late Ashgill is reflected in a various degree of erosion of the upper part of the reefs. It is possible that at the end of the early Ashgill differentiation between stable and tectonically active structural zones along large regional faults began and the Kos'yu-Rogovskaya pericratonic depression was formed. The composition of Ashgill deposits shows evidence of a crisis in the sedimentary situation on the level of the boundary between the early and middle Ashgill and gradual degradation of carbonate sedimentation in the Timan–northern Ural region. This was caused by a global eustatic fall of sea level and migration of the shoal belt to the shelf margin where it became a part of a system of high amplitude reefs. This resulted in the formation of a wide shelf basin. The isolation of extensive intrashelf lagoons in the conditions of arid climate resulted in a change of carbonate sedimentation to siliciclastic-sulphate-carbonate one in a great part of the basin, and to halite-sulphate sedimentation in the pericratonic area (Rasskazova 1988). At the maximum shallowing on the Malaya Tavrota–Yaptikshor formations boundary the whole territory was drained and intensively denudated.

The lowermost Palaeozoic units in the Chernyshev Swell sections are considered to be represented by carbonate breccias of the uppermost Ordovician strata. The presence of carbonate-sulphate strata (Muker and Khorejver formations) in the Pechora Syneclise and Pre-Urals Foredeep depressions (Kaljo *et al.* 1987) is a basic argument for the formation of the Chernyshev Swell as a result of dislocation of rocks by shifting along the Upper Ordovician salt-bearing strata (Timonin 1998, etc.). However, in the Chernyshev Swell there are some localities where beds with the Yaptikshor faunas occur below the carbonate breccias, e.g. in the Iz'yayu River section in the southern Chernyshev Swell (Antoshkina & Beznosova in press).

In some localities of the Chernyshev Swell the Yaptikshor Formation (thickness up to 70 m) consists mostly of thick-bedded dark-grey, partly argillaceous, containing various fossils, intensively dolomitized limestones with a transverse and fracture cleavage. The brachiopods *Holorhynchus giganteus*, *Proconchidium muensteri* (Antoshkina *et al.* 1989) and conodonts *Belodina confluens* (Mel'nikov 1988) and *Pseudobelodina kirki* (Zhemchugova & Mel'nikov 2000) are common. As it is known, the base of the uppermost Ordovician (=Hirnantian) Stage is defined by changes in lithology, which generally mark a transition to shallower marine facies. The deposits of the uppermost Ordovician sequence of the Subpolar Urals formed in clearly shallower environments than the deposits of the underlying dark-grey argillaceous and richly fossiliferous Yaptikshor Formation. Probably, the succession between Yaptikshor carbonate breccias and Llandovery beds with *Pentamerus* (?) sp. in the Iz'yayu River section corresponds to the uppermost Ordovician and lowermost Silurian. The first Silurian brachiopod species identified as *Pentamerus* sp. (aff. *oblongus*) occurs in the Aeronian (Beznosova *et al.* 2002). Near the end of the Ordovician various areas of the carbonate platforms were exposed during the glacio-eustatic sea level lowstand. According to Mel'nikov & Zhemchugova (2000), data on the sedimentation suggest a hiatus at the base of the Silurian, and the Rhuddanian sequence is separated from the overlying Aeronian one by an unconformity in the Pechora Syneclise. Most probably the uppermost Ordovician (?Hirnantian) stage and the whole Rhuddanian Stage have strongly reduced thicknesses in the Chernyshev Swell sections.

Correlation of the sea level curve for the Upper Ordovician successions of the Subpolar Urals, Chernyshev Swell and Pechora Syneclise (Khorejver Depression) in Fig. 1 relied upon the facies succession in the sequence, evidently reflecting changes in relative water depth.

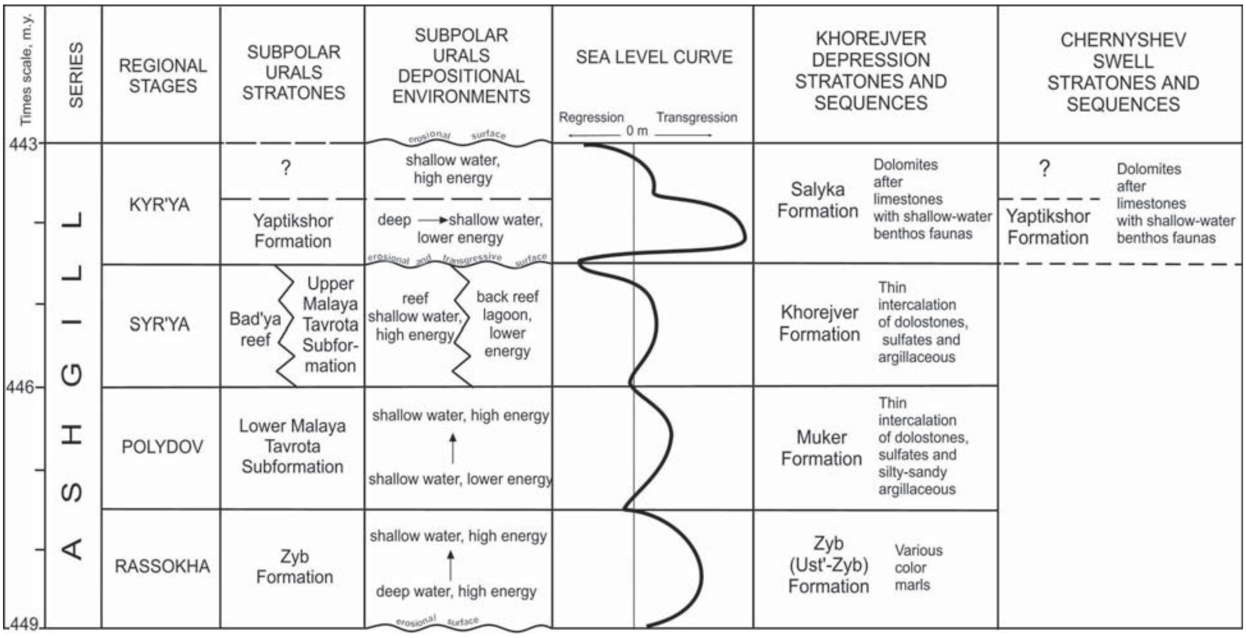

Fig. 1. Correlation of the Upper Ordovician stratigraphic units of the Subpolar Urals, their depositional environments and sea level changes with the stratigraphic units of the Khorejver Depression and Chernyshev Swell.

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Late Ordovician faunas of Kerman Province, east-central Iran

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In Kerman Province Upper Ordovician deposits form the upper part of the Katkoyeh Formation, exposed in a number of sections to the east and southeast of Zarand. Two barren units of possibly fluviatile red and green sandstones and siltstones form the base and top of the Upper Ordovician part of the formation, with a fossiliferous middle unit comprising shallow marine silty argillites and limestones, including several channelled storm beds up to 0.4 m thick containing dense coquinas of brachiopods. This middle unit represents a distinct transgressive trend and contains three successive low diversity faunal assemblages. Up to 10 m of argillites in the lower part contain mostly bivalve molluscs (*Modiolopsis*) and gastropods indicative of inshore Benthic Assemblage (BA) 1-2. It is succeeded by a characteristic interval up to 17 m thick in which the main fossil beds are several storm-generated coquinas with concentrations of disarticulated valves of the brachiopods *Drabovia* aff. *crassior* (Barrande) and *Rhynchotrema* sp., with several species of bivalves and gastropods and fragments of bryozoan colonies. A notable feature of these storm beds is the occurrence of clusters of the small rhynchonellide *Rostricellula* sp. mostly preserved as articulated valves. It is possible that these brachiopods were adapted to a cryptic mode of life in cavities between large bioclasts. Accompanying ostracodes are possibly the most numerous faunal element in the assemblage. Lingulate brachiopods are represented by *Schizocrania* sp. and numerous fragments of an unidentified lingulide, possibly belonging to the subfamily Glossellininae. Other groups include a single species of cephalopod, scolecodonts and chitinozoans. This assemblage can be attributed to BA2-3. A monospecific assemblage of the athyridide brachiopod *Cryptothyrella*? sp. occupies the succeeding 7 m, suggestive of BA3-4.

The occurrence of *Drabovia* aff. *crassior* suggests a close correlation with the late Caradoc Zahorany Formation of Bohemia, supported by the identification of the conodont *Icriodella* cf. *superba* reported by Hamedi *et al*. (1997), but an Ashgill age for the upper beds cannot be excluded because of the presence of *Cryptothyrella*?, a brachiopod genus that is otherwise widespread close to the Ordovician–Silurian bounary.

Drabovia is a characteristic taxon of the Mediterranean Province faunas of Morocco and Bohemia. Together with the remarkably low diversity of the fauna, this suggests a location of the region in high southern latitudes. However, the abundance of the rhynchonellides *Rhynchotrema* and *Rostricellula* is somewhat unusual and could represent a link to the low latitude faunas of East Gondwana.

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Early Ordovician faunas from the Tabas and Damghan regions, Iran and their biogeographical significance

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Substantial new fossil collections have been made from Lower Ordovician sequences in the Derenjal Mountains north of Tabas and in the Alborz Mountains north-west of Damghan. The Cambrian-Ordovician transition in both areas coincides approximately with the replacement of predominantly carbonate deposits of the Derenjal and Mila formations, respectively, by mainly shallow water clastic deposits of the Shirgesht Formation in the Derenjal Mountains and the somewhat deeper water Lashkarak Formation of Alborz, comprised mainly of fine clastics with some carbonate units.

The Shirgesht Formation is remarkable in the development of extensive shell beds formed by the brachiopod *Protambonites*, which are closely analogous with the *Billingsella* beds of the underlying Derenjal Formation. This shallow water *Protambonites* Association (possibly BA2) is widespread in West Gondwana and associated terranes (e.g. Morocco, Bohemia, Spain) and is also present on the Uralian margin of Baltica, which was facing West Gondwana during the Late Cambrian–Early Ordovician. Carbonate units in the uppermost Shirgesht Formation contain a medium diversity rhynchonelliformean brachiopod assemblage, which in addition to *Protambonites* includes impunctate orthides (e.g. *Archaeorthis*) and syntrophinides, whilst linguliformean brachiopods are represented by *Thysanotos* in association with micromorphic acrotretides and siphonotretids dominated by *Eurytreta chabakovi*. This latter species is known otherwise only from the South Urals. Preliminary data on the occurrence of diverse trilobite and conodont faunas in the Shirgesht Formation have been reported previously (for reference, see Hamedi *et al*. 1997) and new collections are currently under study.

The section of the Lashkarak Formation north-west of Damghan is unique for the Lower Ordovician of Iran, because of the co-occurrence of conodonts and graptolites at several levels. Conodont assemblages of the *Paltodus deltifer* and *Prioniodus elegans* biozones are present in the middle part of the sequence, which allows direct correlation with the Baltoscandian sections. The third conodont assemblage with *Baltoniodus* sp., *Periodon flabellum, Microzarcodina flabellum, Drepanoistodus basiovalis, Scolopodus* cf. *rex, Oistodus lancealatus,* etc. can be correlated broadly with the Volkhovian. An argillaceous unit below limestone beds with conodonts of the *Prioniodus elegans* Biozone contains *Tetragraptus* sp., early *Didymograptus* and a diverse trilobite assemblage including *Taihungshania mequeli, Asaphellus* sp. nov., *Ampyx* cf. *priscus, Nileus, Basilicus (Basiliella)* sp., etc. and can be correlated with the Hunnebergian Regional Stage of Baltoscandia. The Lashkarak Formation also contains a diverse echinoderm fauna which includes glyptoshaeritid and shaeronitid diploporans together with caryocystid and hemicosmitid rombiferans. This echinoderm fauna bears mixed Baltic and West Gondwanan signatures, but affinity with Middle Ordovician echinoderm faunas of West peri-Gondwana is even more evident (Nardin *et al*. 2004). Other components of the fauna include agglutinated foraminifers, rhynchonelliformean brachiopods, bryozoans and ostracods, which suggests an early development of benthic assemblages with a fully developed community and trophic structure characteristic of the Palaeozoic Evolutionary Fauna. The presence of the *Thysanotos* lingulate brachiopod assemblage in the Lashkarak Formation is especially noteworthy because it occurs in beds which cannot be older than Volkhovian (in Baltoscandian terms), and therefore it is the latest occurrence of the genus yet reported. The *Thysanotos* Assemblage is widespread in Baltica. It is reported from the late Tremadoc to early Arenig of the South Urals, Poland and Estonia. The only other peri-Gondwanan occurrence of *Thysanotos* is in Perunica (Bohemia) where it is regarded as mainly Tremadoc in age. Our new data suggest that the *Thysanotos* fauna may have a wider range in Iran than in Baltoscandia.

In general, a West Gondwanan affinity of the early Ordovician benthic faunas of Iran is most evident from the rhynchonelliformean brachiopods, trilobites and echinoderms, but distinct links with Baltica can be also traced from the biogeographical distribution of the *Thysanotos* Assemblage, some echinoderm taxa and the appearance of *Protambonites* along the Uralian margin of Baltica. The Iranian faunas may represent a possible source for the Baltic Ordovician Fauna dominated by rhynchonelliformean brachiopods, bryozoans, echinoderms and ostracodes, which emerged in the Billingenian.

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The Dicellograptus Shale (Upper Ordovician) of Bornholm, Denmark

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The Dicellograptus Shale is about 10 m thick on central southern Bornholm, where it is accessible in outcrops along the Læså and Risebæk rivulets. The unit is roughly equivalent to the Skagen, Mossen and Fjäcka formations of Sweden, recently also recognised in western Scania (Bergström *et al.* 1999), but a distinction of these units are not possible on Bornholm and for the time being the old lithostratigraphic nomenclature is upheld.

The lower part of the shales, which is described in great detail by Funkquist (1919), is non-graptolitic, but correlation of bentonite beds shows that this interval represents the *D. foliaceus* Zone (Caradoc) (Bergström & Nilsson 1974). The upper main part of the Dicellograptus Shale is generally quite graptolitic and spans the *D. clingani* and *P. linearis* Zones (Caradoc-Ashgill). However, the graptolite fauna has not been investigated since Hadding (1915) and is in need of revision.

In order to undertake such reinvestigation the classical Læså section at Vasegård was systematically sampled in 10 cm intervals. The accessible strata represent the upper 1.2 m of the *D. foliaceus* Zone, followed by a 4 m thick *D. clingani* Zone, in turn overlain by a 3.5 m thick *P. linearis* Zone. Thin bentonite beds are common in the investigated section and 7 beds (up to 1 cm thick) were encountered in the upper *D. foliaceus* Zone, 11 beds (1-5 cm thick) are present in the *D. clingani* Zone and 4 beds (1-2 cm) are present in the *P. linearis* Zone. The *D. foliaceus* Zone consists of medium grey mudstones, whereas the upper part of the Dicellogratus Shale (*D. clingani* and *P. linearis* Zones) comprises dark to black mudstones rich in graptolites. In the uppermost 10 cm of the *P. linearis* Zone bioturbation is common and the shale is medium grey.

The Dicellograptus Shale is overlain by the grey-brownish Lindegård Mudstone (previously Tretaspis Shale/Jerrestad Mudstone) of which the lower 3.5 m is exposed at Vasegård. There is a 15 cm thick pyritic-phosphoritic conglomerate at the base of the Lindegård Mudstone. The accessible part of this mudstone has not been sampled, since it is generally non-graptolitic (Poulsen 1936); it probably represents the *D. complanatus* Zone.

Based on the newly sampled material the graptolite fauna of the Dicellogratus Shale is currently under study. The reinvestigation has so far demonstrated that the *P. linearis* Zone extends slightly further down the section than indicated by Hadding (1915) and that the lower boundary is associated with a 5 cm thick pyritic conglomerate containing phosphorite nodules. *P. linearis* itself has also been encountered on Bornholm for the first time. The *D. clingani* Zone includes another thin (1 cm thick) phosphoritic level about 30 cm below the *P.linearis/D.clingani* zonal boundary. The graptolite material from the interval around the *D. foliaceus/D. clingani* zonal boundary has not yet been investigated; it is a general impression, however, that the *D. foliaceus* Zone yields very few graptolites, most likely reflecting that the interval is burrowed.

Several levels in the shale are rich in inarticulate brachiopods. These have also been sampled but apart from documenting their distribution they will not be studied any further. The museum collections contain gastropods and bivalves from the Dicellograptus Shale; the level of origin is uncertain. No molluscs have been recorded during the new sampling, suggesting that they are rare.

The study is co-supervised by Dr. Jörg Maletz, Department of Geology, State University of New York at Buffalo.

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The Ordovician margins of Baltica

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Until the very end of the Ordovician, when it collided with Avalonia at 443 Ma, Baltica had been an independent terrane since the breakup of the Rodinia superterrane in the Precambrian. Although Rodinia breakup started soon after 800 Ma, it seems probable that Laurentia and Baltica did not rift apart, with the consequent formation of the Iapetus Ocean, until about 560 Ma, and thus Baltica was a separate terrane for over a hundred million years, with accretions occurring along many of its margins at intervals. Particularly in the early Ordovician, the oceans surrounding Baltica were wide enough to make many of the abundant terrane-diagnostic brachiopod and trilobite faunas to be very endemic.

As with all old terranes, the margins of Baltica today consist entirely of sutures which represent tectonic activity after the Ordovician, although, because of the great thickness of Precambrian rocks making up the terrane centre, the craton of Baltica has remained with very little deformation since the Caledonian Orogeny.

During the late Cambrian and earliest Ordovician, palaeomagnetic measurements indicate that Baltica had rotated by more than 120°, making today's eastern margin, the Ural Mountains, face northwestwards in the Cambrian across the Iapetus Ocean towards Laurentia, the continent which occupied most of today's North America as well as Greenland, northwest Ireland, Scotland and Svalbard. The Iapetus was at its widest in the earliest Ordovician, after which it progressively dwindled in size as the Ordovician continued. Within the Iapetus there were two island arcs, and thus as the Iapetus closed there were a series of continent-arc and arc-arc collisions, and thus many of the arc fragments as well as Laurentian margin fragments ended up within the Scandinavian Caledonides as well as in the tectonic collages within central Newfoundland and the northern British Isles (Cocks & Torsvik 2002). Thus the northwestern margin of Baltica partly lies within the Scandinavian Caledonides, but some of the old terrane also underlies them and probably extends as far westwards as the edge of the continental shelf today.

To the north there are great problems in identifying where the margin of Baltica lay. Because of similarities between the Ashgill brachiopod faunas of the Taimyr Peninsula in northern Siberia and those of central Sweden it had been postulated that that peninsula formed part of Baltica; however, further analysis of the more terrane-diagnostic early Ordovician trilobite faunas (Fortey & Cocks 2003) indicate the Taimyr formed an integral part of the Siberian Terrane rather than of Baltica. The islands of Novaya Zemlya and Vaigatsh were, however, certainly within Baltica. The northeastern margin there has a substantial embayment which parallels the east coast of Novaya Zemlya, but that arcuate deformation was part of the tectonics associated with the eruption of the Siberian Traps at the very end of the Palaeozoic. The position of the margin of Baltica beneath the Barents Sea is uncertain: it has been postulated that Franz Josef Land and the island of Kvitøya to the east of Svalbard also formed part of Baltica, but the evidence for that is not yet supported either by palaeomagnetism or by terrane-diagnostic faunas. However, if it were true, then those areas probably became accreted to Baltica at about the same time as others in the late Vendian as part of the Timanide Orogeny.

Today's eastern margin of Baltica, the Urals, is only a relatively straight tectonic belt because of the severe Upper Palaeozoic strike-slip faulting between Baltica and the independent Kazakh and Siberian fragments formerly to the east of it. Some of these fragments were microcontinents with cores as old as Precambrian, whilst others were short-lived island arcs of Ordovician to Devonian age. However, characteristic Baltic Lower Ordovician trilobite faunas are known from the Urals as far south as Kazakhstan.

Although the Trans-European Suture Zone (TESZ) represents much of today's southwest Baltica margin, the area around the Holy Cross Mountains of Poland, which are south of the TESZ, also formed part of Baltica in the Lower Palaeozoic (Cocks 2002). The Baltic island of Rügen, in northern Germany, although lacking the more terrane-diagnostic benthic macrofossils, had been thought to contain Lower and Middle Ordovician carbonate sedimentary particles more indicative of Baltica than Avalonia, but following more detailed work on the micropalaeontology, it is now thought that these rocks represent sediments deposited in a deeper-water sedimentary ocean basin within the Tornquist Ocean rather than on the margin of Baltica itself. It is not clear where the real margin of the old terrane actually lay, but it was probably not very far south of the TESZ in the Silurian, since thick turbidite basins are present in the Holy Cross Mountains. To the southeast of the latter area the margin of Baltica is contentious; some authors have placed the area of Moesia (within Bulgaria and Romania) within Baltica, but after analysis of the somewhat sparse Lower Palaeozoic faunal data of Yanev (2000), Moesia seems better placed within the Hellenic Terrane, which formed part of the extensive peri-Gondwanan collage.

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Magnitude of sea-level changes in the Ordovician of Baltoscandia

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The two main factors controlling the general shape of any reconstructed sea-level curve are: (1) time scale, e.g., relative duration of the time intervals corresponding to stratigraphic units and gaps in the succession under consideration; (2) magnitude of relative sea-level changes. The latter factor is most difficult to calculate. The sea-level drops caused by forced regressions are usually manifested by erosional unconformities and gaps in shallow-water environments. To estimate the magnitude of a sea-level drop, one should take into consideration the distribution area of the erosional surface and significance of the erosion of the underlying rocks. Precise estimation of the sea-level rises is even more difficult as they are almost not expressed in deep-water settings. In shallow-water environments one can take into account the area covered by sediments of the corresponding depositional sequence and facial expression of the transgressive systems tract deposits.

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 1.5–3.0 to 8–9 m.y. To ease reference and identification, an individual name has been given to each depositional sequence (from base to top): (1) Pakerort, (2) Latorp, (3) Volkhov, (4) Kunda, (5) Tallinn, (6) Kegel, (7) Wesenberg, (8) Fjäcka, (9) Jonstorp and (10) Tommarp (Dronov & Holmer 1999). The most prominent unconformities with extensive erosion of the underlying beds coincide with the base and top of the Ordovician succession as well as with the base of the Latorp and Wesenberg sequences. The strong erosion and development of regional unconformities can be regarded as evidence for forced regressions and sea-level drops of great magnitude comparable to modern glacial regressions (about 100 m). Unconformities at the same stratigraphic levels can be recognised in the Siberia (Kanygin 2001), Gondwana (Carr 2002) and Laurentia (Ross & Ross 1992, 1995) palaeocontinents.

The Latorp, Volkhov and Kunda sequences demonstrate 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). A rapid change in the depositional environment, from tide-dominated to storm-dominated at the base of the Latorp sequence, as well as an invasion of new groups of fauna, are also attributed to a rise of the sea level. At that time epicontinental sea covered almost all of the Russian platform, providing connections between the Urals, Moscow basin and Baltoscandia.

The lower boundary of the Volkhov sequence is interpreted as a 2nd -type sequence boundary (Dronov & Holmer 2002) with a long period of stillstand 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 sea-level drop at the Volkhov–Kunda boundary was larger than that at the Latorp–Volkhov boundary (30–40 m). The Tallinn sequence is represented by shallower-water deposits than 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 grey deposits. The organic-rich kukersite-bearing strata demonstrate progradational stacking patterns and form the 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 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 notable conformities in the entire Ordovician of Baltoscandia, with an estimated extent of as much as 40–50 m. The base of the next Fjäcka sequence is interpreted as a 2nd-type sequence boundary. The black shale sedimentation reached its maximum in the transgressive systems tract of the Fjäcka 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 Fjäcka sequence. 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 differs from that of Vail *et al.* (1977), Ross & Ross (1992, 1995) and Nielsen (2003). These models assume a prominent sea-level drop at the base of the Middle Ordovician and a long-term lowstand during the entire "Volkhovian" and Darriwilian (80–100 m lower than in the Lower and Upper Ordovician). In contrast, our data point to a moderate sea-level drop at the base of the Volkhov sequence 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 seem to be the most prominent transgressions in the whole Ordovician of Baltoscandia, which means that the Middle Ordovician was not a lowstand but rather a highstand interval. The marked differences in the sea-level curve patterns established for North America and three other palaeocontinents could be explained by tectonic uplift of the eastern margin of North America during the Middle Ordovician.

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Mishina Gora section: results of recent investigations

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The section is situated in an old limestone quarry in the former Mishina Gora village (Pskov Region) within an area of exclusively Devonian outcrops. The quarry displays a succession of nearly vertically inclined Ordovician rocks from the Pakerort to Lasnamägi regional stages (Assatkin 1938). After five years of clearing (Kushlina *et al.* 2001) the following 36.98 m thick section is now available for study (Fig. 1).

1. Tosna Formation (Fm.): Grey medium-grained poorly sorted quartz sands with scattered fragments of *Obolus* shells (0.24 m).
2. Leetse Fm.: Greenish-grey poorly sorted quartz-glauconitic sands (0.27 m).
3. Zebre Fm.: Red dolomites with numerous hardgrounds depicted by yellow goethitic impregnation (0.24 m).
4. Kriukai Fm.: Red dolomites with iron impregnated hardgrounds. At the base of the formation "amphora-like borings" of *Gastrohaenolites oelandicus* have been found (1.83 m).
5. Sillaoru Fm.: Greenish-grey clayey limestone with abundant ferriferous ooids developed around bioclasts ("Lower oolite bed") (0.40 m).
6. Baldone Fm.: Intercalation of grey clayey limestone, sometimes with scattered glauconite grains, and red limestone with iron impregnated hardgrounds and abundant *Cephalopod* shells (11.66 m).
7. Segerstad Fm.: Relatively hard thick-bedded limestone of greenish-grey and red colour. Contains *Echinosphaerites* (8.80 m).
8. Stirnas Fm.: Intercalation of grey clayey limestone and marl beds with abundant *Echinosphaerites* (13.54 m).

The Mishina Gora section could be regarded as a type section for the transitional zone (Põlma 1967) between the North Estonian and Central Scandinavian confacies belts (Männil 1966; Jaanusson 1976). The lithological and faunal changes across the confacies belts are much more dramatic than along their boundaries. In that context, the Mishina Gora section provides an excellent opportunity to establish a precise litho- and biostratigraphical correlation between typical Scandinavian and East Baltic facies for the Billingen–Lasnamägi time interval. The core material from the Mishina Gora district stored in Pskov (Russia) and Särghaua (Estonia) can be used as a reference section.

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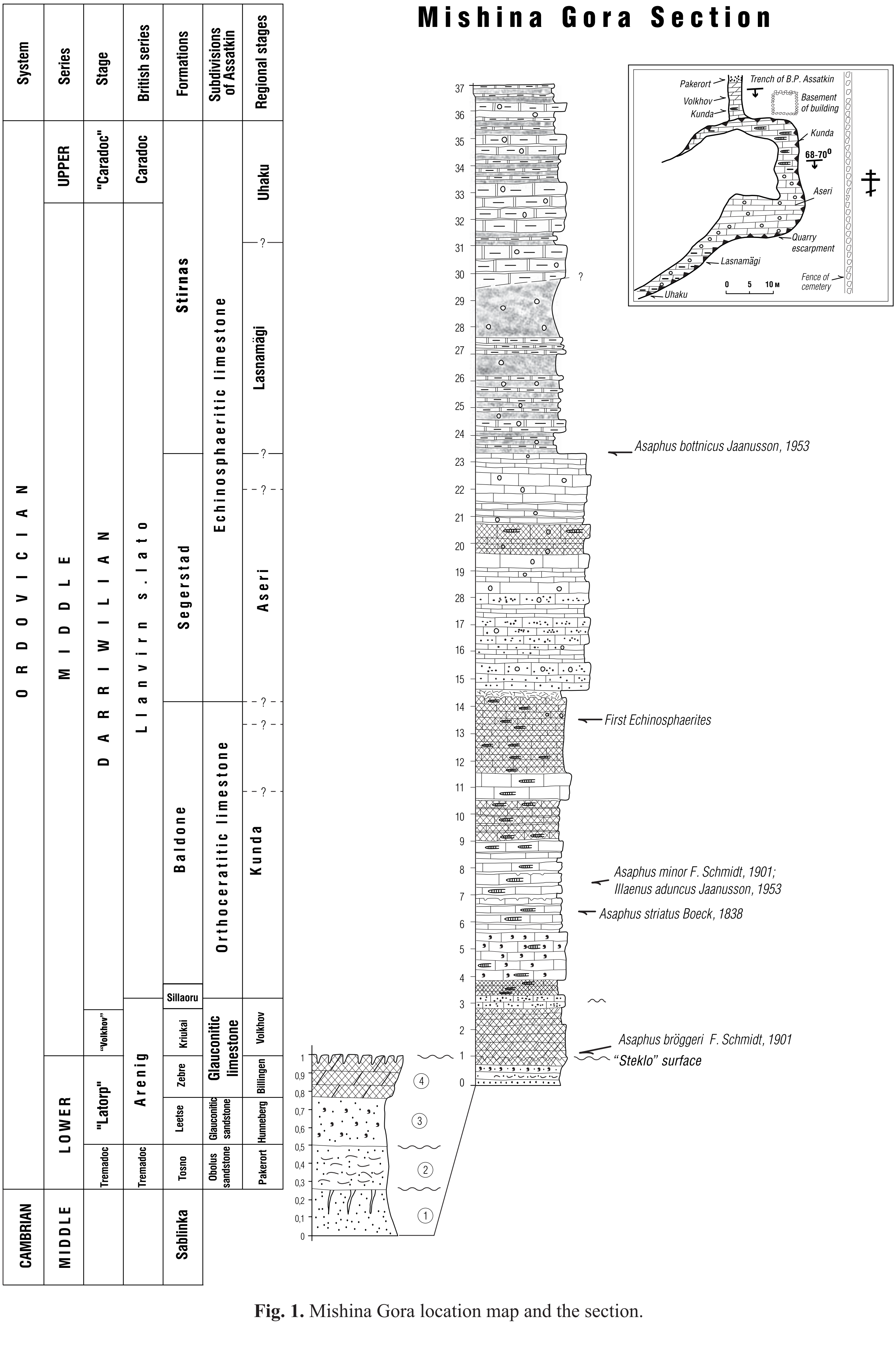

Fig. 1. Mishina Gora location map and the section.

Conodont stratigraphy of the Ordovician volcanogenic and volcanogenic-sedimentary rock assemblages of the Southern Urals

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The finds of Ordovician conodonts in volcanogenic and volcanogenic-sedimentary rock assemblages in tectonically different zones of the Southern Urals have resulted in a new understanding of regional stratigraphy. In many cases the Ordovician age has been proved for the volcanogenic complexes considered earlier as Devonian or Silurian ones. In the Sakmara Zone the black smoker-related massive sulphide ore deposits are connected with volcanogenic rocks, which allows their correlation with similar in the composition and age sulphide-bearing complexes of the Tagil Zone. Five types of Ordovician rock assemblages (or complexes) have been established: (1) chert/basalt type, (2) mixed composition volcanogenic type, (3) volcanogenic/tuffaceous type, (4) tuff/chert type, and (5) terrigenous/arkosic type. They differ in terms of completeness of stratigraphic sections and are related to various elements of the active margin of the Palaeo-Urals Ocean.

In the northern part of the Magnitogorsk Zone, to the east of the Uraltau Uplift separated by the Main Urals Fault, the Western Magnitogorsk Subzone is distinguished. The latter probably represents the relict of an accretionary prism with the Devonian island arc complex located on it. A dislocated package of nappes has been established in the accretionary prism. Here, the elements of the section of the chert/basalt complex (i.e., Polyakovskaya Formation) are tectonically juxtaposed so that units differ in terms of the composition, thickness and stratigraphic range. The chert/basalt complex represents the upper part of the ophiolite association formed in a back-arc basin. The stratigraphic range, as established by conodonts, is mid-Arenig–Ashgill (Ryazantsev *et al.* 1999; Dubinina & Ryazantsev 2000). It can be extended further down because redeposited late Tremadoc *Loxodus bransoni* has been found here.

Other types of the Ordovician rock assemblages represent a facies trend and characterize various elements of an island arc and back-arc basin. They are attributed to the structure of the Sakmara Zone (Ryazantsev *et al.* 2001).

In a section of mixed composition volcanogenic type we see the rock alternation in which effusive basic or acid rocks prevail, but andesites and their tuffs are present as well. Massive sulphide ore deposits are connected with these rocks. The ore-bearing deposits are considered to be of Ordovician, Silurian or Devonian age. The Yamankasy ore deposit, genetically connected with black smokers, is recognized as a typical one. The ore body is interpreted as a submarine dome at the foot of which the ore-clastic colluvium is accumulated. The ore deposit contains vestimentiferan tube-worms, brachiopods and pelecypods of a wide age range. Ore-bearing beds occur in transition series from the rhyolite-dacite to the basalt part of the section. Above ore bodies thin horizons of quartz rhyolites and their tuffs as well as rare thin horizons and lenses of red cherts among basalts are observed (Ryazantsev *et al.* 2003). In Blyava quarry these horizons yield the Caradoc–Ashgill conodont assemblage represented by *Dapsilodus mutatus* (Branson et Mehl), a transitional form from *Periodon* cf. *aculeatus* Hadding to *P*. cf. *grandis* (Ethington), *Drepanodus robustus* Hadding, *Protopanderodus* cf. *liripipus* Kennedy, Barnes et Uyeno, *Panderodus* sp. Upsection, the beds of carbonaceous shales among basalts contain Llandovery graptolites. In the Komsomolskoye ore deposit the cherts above the ore-bearing bed include the late Caradoc–Ashgill conodont assemblage represented by *Hamarodus brevirameus* (Walliser), *Protopanderodus liripipus* Kennedy, Barnes et Uyeno, *Periodon grandis* (Ethington), *Scabbardella altipes* (Henningsmoen), *Belodina confluens* Sweet, *Drepanodus* cf. *robustus* Hadding, *Istorinus* sp. Similar conodont associations are present in a type section of the Bauluskaya Formation at Baulus Mountain, in the Utyagulovskaya synform, and in the vicinity of Khmelevka village.

The volcanogenic/tuffaceous rock assemblage is referred to as the Guberlinskaya Formation. Tuffites, acid tuffs and basalts predominate in a section. On the right side of the Guberlya River, in a thin bed of red tuffites among basalts we have recovered Llandeilo conodonts *Ansella nevadensis* (Ethington et Schumacher), *Periodon aculeatus* Hadding, *Dapsilodus similaris* (Rhodes), *Walliserodus* sp. To the north, the monotonous red- and pistachio-coloured tuffites with rare members of basalts and basalt tuffs with carbonate cement prevail in a section. On the left side of the Malaya Kayala River, in a horizon of red tuffites the Early Llandeilo conodonts *Pygodus serra* (Hadding), *Periodon aculeatus* Hadding, *Eoplacognathus* cf. *robustus* Bergstrom, *Dapsilodus mutatus* (Branson et Mehl), *Protopanderodus varicostatus* (Sweet et Bergstrom), *Drepanoistodus suberectus* (Branson et Mehl), *Plectodina* sp., *Ansella* sp. have been found for the first time (Ryazantsev *et al.* 2003). In the north-eastern limb of the Utyagulovskaya synform, at the junction of Almash and Kyzylyar revines, the early Llanvirn conodont association represented by *Baltoniodus medius* (Dzik), *Strachanognathus parvus* Rhodes, *Periodon aculeatus* Hadding, *Drepanodus arcuatus* Pander, *Protopanderodus* sp. is found in a member of red aleurolites located between acid tuffs and basalts. A rich complex of the late Llanvirn–early Caradoc conodonts including *Spinodus spinatus* (Hadding), *Walliserodus ethingtoni* (Fahraeus), *Protopanderodus liripipus* Kennedy, Barnes et Uyeno, *Drepanodus robustus* Hadding has been collected from the tuffaceous section in the vicinity of Churaevo village.

The tuff/chert rock assemblage is referred to as the Kuraganskaya Formation overlying the Tremadoc Kidryasovskaya Formation with terrigenous/arkosic type of section. In the vicinity of Blyava station in the lower part of the Kuraganskaya Formation section aphyric basalts alternate with red tuff-aleurolites, in which we have recognized Mid-Arenig *Acodus delicatus* Branson et Mehl, *Drepanodus conulatus* Lind, *Scandodus furnishi* Lind. Higher stratigraphic levels are known in the Shaitantau Mountains and to the east of Churaevo village. Here, above the tuffites analogous to those in the vicinity of Blyava station, cherty tuffites and tephroites containing Llanvirn–Ashgill conodonts appear (Ryazantsev *et al*. 2000; Dubinina *et al.* 2001). Upsection, these rocks are conformably overlain by Silurian carbonaceous shales of the Sakmara Formation.

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Microgastropods in the Ordovician of Baltoscandia: potentials and prospects

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A recent evaluation of gastropod diversity dynamics during the Great Ordovician Biodiversification event in five palaeocontinents has presented an interesting pattern for Baltica (Novack-Gottshall & Miller 2003). In contrast to for instance Laurentia where gastropods diversify during the Tremadoc, the curve in Baltica rises slightly at the middle of the Volkhov and more prominently at the end of the Darriwil Stage (Kunda Regional Stage). This was thought to be an ecological signal, though uncertain age determinations and a lack of a revised taxonomy was also pointed out (Novack-Gottshall & Miller 2003).

A similar picture appear when comparing their diversity curve of to the most recent monographic study of Baltoscandian gastropods (Koken & Perner 1925), including the works of Yochelson (1962, 1963) on Lower and Middle Ordovician gastropods of the Oslo Region, Norway. The distribution is based on number of species, and gives a very crude approximation owing to the problems of taxonomy and correlation mentioned above. However, fieldwork by both of us in Baltoscandian areas further confirm the virtual absence of macrogastropod remains in the earliest part of the succession. Thus, these factors seem to confirm that an underlying ecological signal is expressed in the diversity curve.

This may not be the entire story. Some macro remains of gastropods are still to be described, but more importantly, we hope to show that there is a large undescribed material of microgastropods available for the lower part of the succession in particular. How this may effect a diversity curve remains to be studied.

Microgastropods refer to specimens smaller than 1 mm. The material we are aware of have usually occurred in samples analyzed for conodonts or ostracods. Most frequently phosphatic or glauconitic infillings occur, with more rare phosphatic replacement of entire shells. Bockelie & Yochelson (1979) analyzed various coiled to straight steinkerns from the Ordovician of Svalbard, and pointed out the problems of identifying mollusc among such tube-like infillings. This is a problem we are very much aware of, but the material we have seen so far is consistent with an interpretation favouring gastropod affinity.

Material

We suspect that there is an abundance of material of microgastropods across the Baltoscandian platform. This notion stems from a few pilot investigations, and from samples made available with the help of several colleagues. Conodont workers especially have known for years about this material, but it has yet to be studied and analyzed.

Some examples may serve to illustrate the potential. In the Oslo Region, infillings of microgastropod shells are found in the Huk and Steinsodden formations (J.A. Rasmussen, Copenhagen, pers. comm. 2004). No further material is known from this area so far. In central Sweden, several collections are known from the Hällekis quarry in Västergötland, where a rather diverse assemblage is seen. Most of the specimens seem to be steinkerns, but the abundance of material is high promising interesting finds.

The earliest material we are aware come from Estonia, and are glauconitic infillings from the Päite Mbr of the Billingen Stage, but similar glauconitic material occurs more frequently in the Volkhov Stage. Here, the genus *Mimospira* can be identified. Isakar & Peel (1997) described a minute *Miospira* from the younger Rakvere Stage, which illustrates well the fine preservation that may occur. Further material with preserved protoconchs come from the Porkuni quarry. Various samples of microgastropods come also from the St. Petersburg Region, with some superb preservation through phosphatic replacement in material from Mishina Gora (made available by T. Tolmacheva, St. Petersburg). Occurrences of material in the upper part of the Ordovician succession are less well known, and in depth studies are required. One known example comes from the Ashgill Boda Limestone of Sweden, where one or perhaps two species of microgastropods dominate the gastropod fauna (Ebbestad pers. obs.; Gubanov *et al*. 1999).

Potential and prospects

There seems to be a large, untapped potential for material of microgastropods in the Ordovician of Baltoscandia. Study of such a material may lead to a better understanding of diversity and distribution patters of gastropods in this region, especially for the lower part of the succession. An important aspect is that there is a potential for understanding of phylogeny through study of protoconch morphology, a potential that if it is realized may be huge and significant (Dzik 1994; Bandel & Fryda 1999; Wagner 2001). There may also be a potential for correlation across the platform, especially if the material occur in abundance and identification can be reliable.

Intuitively one would expect that microgastropods most likely represent protoconchs or juvenile specimens. It is therefore curious that macro remains of adult specimens are missing, especially in the lower part of the succession. This may be a taphonomic effect, or it may be that a true meiofauna was present. This is an aspect to investigate further. Miller & Connolly (2001) examined substrate affinity for various groups in the Ordovician, on the general hypothesis that radiation of major groups in the Ordovician was linked to this property, especially increased development of siliciclastic facies. This is also something to address for the Baltoscandian setting, where only recently a better and updated understanding of facies development is emerging (Dronov & Holmer 1999; Egenhoff 2004). Finally, the biodiversity curves for the Ordovician of Baltoscandia presented by Hammer (2003) are also interesting food for thought, in that low diversities are apparent for several groups in the early part of the succession, though the gastropod distribution as now known does not follow the diversity peaks shown by other groups in Baltoscandia.

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Diversity trends of Ordovician jawed polychaetes

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Scolecodonts, the jaws of polychaete annelid worms, are common microfossils in Ordovician sedimentary rocks and are currently known from many different regions of the world, e.g., Australia, Kazakhstan, China, South America, North America, and Europe (Hints *et al.* 2004). However, except for the two latter regions, the available data are too meager to allow analyses of diversity trends and biogeographic distribution patterns at the genus and species level.

Preliminary data on the inter-regional distribution (between Baltica and Laurentia) of jaw-bearing polychaetes indicate that the faunas of different continents were similar in composition during the Ordovician (Hints *et al.* 2000; Eriksson & Bergman 2003; Hints *et al.* 2004). Most genera were common to Laurentia and Baltica, but also the majority of the few species recovered from other regions apparently belong to these widespread genera. Approximately 50 Ordovician genera, belonging to some 15–20 families, are currently known. The total number species is difficult to assess particularly because most of the historically described taxa were recognized using the outdated single-element taxonomy and are of little use without careful revisions of old literature and the corresponding type collections. Nonetheless our estimations, primarily based on multi-element taxa, indicate that up to 150 species occur in the Baltic region and some 100 species have been recorded from the North American mid-continent region.

During the Early Ordovician, the frequency and diversity of jawed polychaetes appear to have been very low. However, at least three different genera with a primitive placognath/ctenognath type jaw apparatus (see Fig. 1A, B) are recorded from different parts of the world (including recent discovery from Estonia). Although these collections require further taxonomic study it seems that some of the early forms may be related to xanioprionids and conjungaspids and some are very similar to *Lunoprionella*, a common genus in younger Ordovician strata. More advanced taxa, such as those with labidognath and prionognath type apparatuses (Fig. 1C, E), are unknown in the Lower Ordovician thus far. *Archaeoprion quadricristatus* (Fig. 1D), a primitive and enigmatic taxon with xenognath type jaw apparatus, appears in the Darriwilian (Lasnamägi Stage).

Members of *Oenonites* and *Mochtyella* (Fig. 1B, E) are first recorded at the beginning of the Middle Ordovician (Underhay & Williams 1995). These genera, particularly the former, commonly dominate the assemblages in younger Ordovician as well as Silurian strata (Eriksson 1997; Hints 2000). Although the diversity remains relatively low until the early Darriwilian, it seems evident that already by the earliest Middle Ordovician the main polychaete lineages had become differentiated.

A major increase in species diversity and abundance, as well as a rapid increase in the number of genera are recorded in the Darriwilian and the earliest Late Ordovician. Moreover, this interval marks the first appearance of most of the genera that become common in younger strata. Therefore the recorded genus-level diversity remains rather stable through the Late Ordovician. Since most genera occurring in the uppermost Ordovician range into the Silurian, there is no distinct drop in genus-level diversity at the Ordovician–Silurian boundary which is marked by a significant diversity drop in many other fossil groups. The jawed polychaete faunas and their response to the end Ordovician extinction event remain however to be studied in detail.

Estimates of the diversity trends, including the rates of appearance and disappearance, indicate that the Ordovician was the main period for the diversification of Paleozoic scolecodont-bearing polychaetes.

Because of the relatively scarce and patchy data set (particularly for the Lower and lower Middle Ordovician) the present diversity estimates should be viewed as tentative and pending ongoing studies the number of genera will most certainly increase and the ranges of many taxa will be extended. However, the diversity was at least not lower than here indicated and this abundant and diverse metazoan group clearly merits further attention.

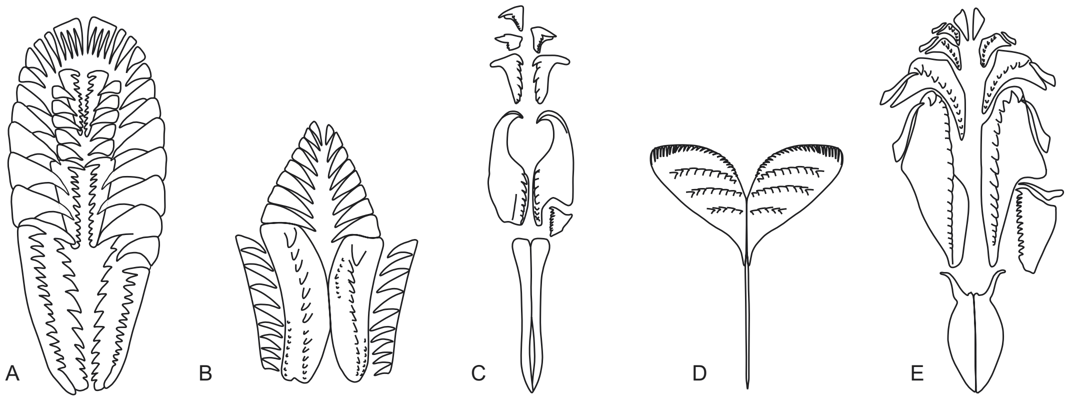

Fig. 1. Maxillary apparatuses of fossil eunicidan polychaetes (modified from Kielan-Jaworowska 1966; Mierzejewski & Mierzejewska 1975; Bergman 1998); A — Ctenognath type (*Tetraprion pozaryskae*), B — Placognath type (*Mochtyella polonica*), C — Prionognath type (*Atraktoprion sp.*), D — Xenognath type (*Archaeoprion quadricristatus*), E — Labidognath type (*Oenonites wyszogrodensis*).

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Ordovician Bryozoa of Estonia: diversity dynamics and biogeography

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Bryozoa are a very common element of communities that prospered in shallow seas during the Ordovician. Eighty-six bryozoan genera have been described from the Estonian Ordovician, of which 18 are endemic. The earliest Bryozoa in Estonia are known from the Latorp Stage (Billingen) of the Lower Ordovician. These are simply-built, dome-shaped trepostome and cystoporate species. During the Volkhov age a few additional cryptostomes and trepostomes appeared here belonging to taxa mostly restricted in their distribution to the Baltoscandian region. By the Llanvirn all stenolaemate orders were represented in the faunal assemblages of the Estonian Ordovician, and morphological diversity was high. Most of the species are apparently endemics. However, the presence of some North American species points to increasing faunal exchange between these ancient provinces. Taxonomic diversity increased rapidly in early Caradoc, reaching a peak during the Oandu and Rakvere stages (115 species). Many North American species are present in Estonian faunas during the Caradoc, providing evidence of intensive faunal exchange. At the end of Caradoc taxonomic diversity dropped greatly. Only 38 species have been described from the Nabala and Vormsi stages of the Estonian Ordovician. The bryozoan fauna recovered slightly at the beginning of Ashgill, but was reduced to 23 species in the Porkuni Stage, apparently due to the Hirnantian extinction event. The Ashgill bryozoan faunas of Estonia contain a small number of species known also from North America and Siberia.

Lithofacies zonation of the “Glauconite Sandstone” and the lower part of the “Glauconite Limestone” through the Russian part of the Baltic–Ladoga Klint: preliminary results

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Detailed lithostratigraphic restudy of many natural outcrops of the “Glauconite Sandstone” and the lower part of the “Glauconite Limestone” along the Russian portion of the Baltic–Ladoga Klint was carried out in the last year.

The “Glauconite Sandstone” overlies laminated black shale (“*Dictyonema* Shale”) with a major stratigraphical gap. The base of the “Glauconite Sandstone” is placed at different stratigraphical levels from the *Paltodus deltifer* Zone to the *Prioniodus elegans* Zone. The “Glauconite Sandstone” is gradually passing upwards into the “Glauconite Limestone”. The lower part of the “Glauconite Limestone” is limited by the omission surface bored by *Gastrochaenolites*. This surface separates the Billingen and Volkhov regional stages.

The correlation of discrete sections along the Klint allowed us to determine a specific facies zonation in the lithology of the “Glauconite Sandstone”. We distinguished four different facies zones within a distance of approximately 230 km from the Suma River in the west to the Sjas’ River in the east. Differences between the zones lie in the completeness of a section, composition, grain size (for sandstone), etc. The boundaries of lithofacies zones coincide with narrow fracture-flexure zones. These zones separate the large blocks of the Klint with different altitudes. Zander & Salomon (1971) dated these zones as pre-Devonian. Our data testify that disturbance zones probably caused lithofacies differences between large blocks of the Ordovician sea bottom.

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Phosphatized burrows from the basal layer of the “Glauconite Sandstone” (Billingen Regional Stage) at the Tosna River

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Unique phosphatized burrows were discovered in the “Glauconite Sandstone” at the Tosna River near St. Petersburg.

The “Glauconite Sandstone” (total thickness 45–60 cm) is accessible in 10–12 outcrops along the canyons of the Tosna and Sablinka rivers. It is represented by alternating green quartz-glauconite sandstone, grey and claret-coloured clay and olive-grey sandy marl, underlain by the basal layer of quartzose sandstone. The homogeneous, non-laminated sandstone is medium- to fine-grained, weakly cemented by clay. The basal sandstone (thickness 8–18 cm), with the admixture of dark fragments of phosphatic brachiopods, is dirty-yellow- or pink-coloured. It overlies laminated black shale (also termed as “*Dictyonema* Shale”) with a major stratigraphical hiatus. According to T. Ju. Tolmacheva (pers. comm.), the conodonts from the basal sandstone are represented by the assemblage of the *Prioniodus elegans* Zone (Billingen Regional Stage, Arenig).

The phosphatized burrows are accumulated within very narrow stratigraphic intervals in the upper part of the basal layer. The burrows are usually vertical vase-like, heart-like, bullet-like, pumpkin-like, rarely amphora-like structures 1–4 cm in diameter and 3–10 cm long. Most of the burrows are individual structures but compound structures composed of 2–3 burrows also exist. The apertures of the burrows coincide with the roof of the basal layer. The burrows number 25–50 per square metre.

The burrows are hard and easily extracted from the weakly cemented host sandstone. They are brown inside, but the thin crust on the surface is light beige or light pink. The inside material of the burrows is composed of two components: cryptocrystalline phosphorite and sandy quartz particles. The amount of the components is changeable. The main part of the burrows is represented by phosphorite with scattered sandy particles. There occurs also some pure phosphorite and sandstone with phosphorite cement. Central parts of the burrows are often broken by syneresis cracks. No body fossils or *in situ* skeletons have been discovered inside the burrows.

Up to now, such phosphatized burrows have not been reported from other outcrops along the Baltic–Ladoga Klint line. These burrows are unique examples of *in situ* phosphorization of biogenic structures on the palaeobottom during the Early Ordovician in Baltoscandia. We suggest that the burrows were formed in coastal shallow water. The probable source of phosphorous in ground water was organic matter formed in the local zone of high productivity.

Trilobite biostratigraphy of the Tremadoc Bjørkåsholmen Formation on Öland, Sweden

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The Lower Ordovician of Baltoscandia is characterized by the initiation of extensive carbonate deposits of the Ceratopyge Limestone. The limestone succession has a broad regional distribution and its associated sediments were deposited in a shallow water epicontinental sea across the Baltic platform during the late Tremadoc (Jaanusson 1976, 1982; Dronov & Holmer 1999). Those particular depositional factors have no present day equivalents. The unit is definitely remarkable in its apparent homogeneous facies, lithological and faunal composition. The significance of the Ceratopyge Limestone succession was early referred by numerous authors and recognized sedimentologically and stratigraphically. Tjernvik (1956) reinvestigated the Lower Ordovician beds in Sweden in detail and completed an account of the rich Ceratopyge fauna.

In older studies a combined bio-litho stratigraphical concept of the unit has mostly been used, however a modern lithostratigraphical definition was given by Owen *et al*. (1990). The definition of the formation is based on the hypostratotype at Bjørkåsholmen in Slemmestad, Norway, and should be referred to instead of the historical synonym. The transition from the dark underlying Alum Shale Formation to the grey limestone beds marks the base of the Bjørkåsholmen Formation (BHF), this is evident in the change in not only lithology but also the typical associated fauna e.g. the Ceratopyge fauna (Tjernvik 1956; Ebbestad 1999). In Sweden the BHF is followed by the Hunneberg–Billingen Latorp Limestone and Volkhov Lanna Limestone. Shales of the Tøyen Formation follow the partially glauconitic limestone in Norway (Owen *et al*. 1990). The most recent revision of the trilobite fauna recognized 36 species assigned to 28 genera (Ebbestad 1999).

The present study investigates trilobite distribution of the BHF in southern Öland, the easternmost outcrop of the formation. The underlying crystalline bedrock on Öland dips weakly to the east resulting in the exposition of the oldest overlying sedimentary rocks in the west and the youngest beds in the east (Jaanusson & Mutvei 1982). The Upper Cambrian Alum Shale Formation is continued in the Lower Ordovician successions and is subsequently overlain by the upper Tremadoc limestone deposits represented by the BHF. Outcrops of this unit are fairly rare, confined to a few localities in the southern and south-central parts of the island.

The material presented in this study was collected at the coastal section at Ottenby and at the Cementa quarry in Degerhamn. Sequences at Ottenby and Degerhamn were logged and material collected for a biostratigraphical study. To obtain trilobite abundances the sample frequency method (Jaanusson 1979; Nielsen 1995; Ebbestad 1999) was applied. The lower boundary of the BHF in Öland is marked by the occurrence of glauconiferous limestone nodules, represented in both localities. The main limestone beds in Öland, e.g. constituting continuous beds having the main trilobite abundance, are grey and micritic, and with some intercalations of glauconitic shale. In addition scattered grains of glauconite and small accumulations of pyrite are evident in the main limestone. The uppermost bed in the two successions, devoid in trilobites, marks the upper boundary of the formation, it is also exceptionally glauconitic suggesting slow deposition and starvation of sedimentation. The BHF at both Degerhamn and the Ottenby section has an approximately thickness of 0.6 meter and several discontinuity surfaces are evident.

The trilobite abundance logs give a proposal for the distribution of the trilobite fauna during the upper Tremadoc in Öland. The material collected from the section in Degerhamn and Ottenby belongs to the Ceratopyge fauna and biostratigraphically the unit is connected to the *Apatokephalus serratus* Zone. The trilobite abundance distribution from each of the localities is very consistent. Additionally, the fossil assemblages in Öland and the Oslo Region (Ebbestad 1999) are built up by the same typical Ceratopyge fauna, indicating a correlation. Faunal signals show several similarities in the trilobite abundance data and both areas have an upwards declination of trilobite quantities in the sequences. Faunal distribution of trilobites of the Ceratopyge fauna is thus very coherent throughout the platform. This suggests widespread stable conditions of the fauna throughout the Baltic Platform during sedimentation of the BHF. The Central Baltoscandian Confacies Belt (Jaanusson 1976), to which Öland belongs, and the Oslo Region belonging to the Oslo Confacies Belt, demonstrate no facies differentiation of the platform until post-Tremadoc (Jaanusson & Mutvei 1982).

The basal limestone nodules at the two investigated sections are devoid of trilobite remains. Periods of oxygenated lime mud sedimentation, reflected by the basal limestone nodules incorporated in the shale, may indicate times of fluctuations in the sea level. The main limestone units have little or no intercalations of shale, thus representing episodes of more stable sedimentary facies. The unstable settings may suggest non-favourable environments for the establishment of nileid communities (Fortey 1975), here represented by the Ceratopyge fauna, described by the absence of trilobites in the lowermost limestone nodules. Fluctuations of the sea level were probably confined to local settings and are suggested by extended periods with large inputs of oxygen during lime mud sedimentation. Constrained settings are most likely a result of diverse bottom topography and differences in current and wave distribution throughout the area (Ebbestad 1999). The faunal logs clearly demonstrate this idea. Ebbestad (1999) stated similar conditions for the depositions of the basal limestone beds in the BHF of the Oslo Region. The Ceratopyge fauna display coherent distribution in Öland and certainly across the platform, moreover, the distribution does not correlate with specific beds, presumably demonstrating dissimilar sedimentological developments of the two localities, suggesting that the deposition was diachronous.

The upper Tremadoc and early Arenig represents a time of large global sea-level changes by the lowering of the seas (Fortey 1984), this event the Ceratopyge Regressive Event (CRE) (Erdtmann & Paalits 1995) could have resulted in the depletion of sediments. In Baltica this is clearly shown by the end of the *Apatokephalus serratus* zone, the top of the Bjørkåsholmen Formation, and in addition the disappearance of its associated Ceratopyge fauna.

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Early Silurian brachiopod palaeocommunities from the Oslo Region: the Ordovician-Silurian transition

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The Ordovician–Silurian transition marked a major taxonomic turnover in marine faunas, particularly within the dominant sessile benthos, the Brachiopoda. Ecological changes across the boundary were less severe but nevertheless significant. Within the Oslo Region, southern Norway, the Ordovician–Silurian transition is represented by a number of contrasting faunal successions in different environmental settings (Baarli *et al.* 2003), each revealing different aspects of this event.

In the central Oslo Region relatively deep-water facies may have formed a refugium for a variety of more typically Ordovician taxa within this part of the basin during the early Rhuddanian (Baarli & Harper 1986). In the northern part of the region two variably incomplete but informative sections through the Ordovician–Silurian transition provide contrasting faunal developments. In Ringerike the boundary section is incomplete. Late Rhuddanian assemblages, associated with carbonate and mudstone facies are dominated by mixed Late Ordovician and Early Silurian elements such as *Clorinda*, *Isorthis*, *Mendacella*, *Zygospiraella* (Thomsen & Baarli 1982), *Rostricellula*, and *Platytrochalos*. The nature of the boundary, farther north in Hadeland is less certain but the basal Sælabonn Formation is probably early Rhuddanian in age. It contains *Dalmanella*, *Leptaena* and *Coolinia* together with *Zygospiraella*, the latter two commonly occurring as monospecific shell concentrations within siliciclastic facies (Heath & Owen 1991). Data from complete successions across the boundary interval in South China (Rong & Harper 1999) suggest that following the two main extinction events the brachiopod fauna regrouped during two survival intervals (late Hirnantian and early to mid Rhuddanian, respectively), a recovery interval (late Rhuddanian-early Aeronian), and finally a sustained radiation during the mid to late Aeronian. All three, well-documented sections in the Oslo Region provide different windows on this succession of events. The near coeval Asker and Hadeland developments contrast the deep-water refugium of the basin with eurytopic opportunistic survivors together with some new Silurian taxa occupying relatively shallow-water clastic environments. Whereas in the younger Ringerike part of the succession the recovery was well underway, spiked by associations of more typically Silurian taxa.

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The Upper Ordovician of Estonia: facies, sequences, and basin evolution

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Upper Ordovician strata of Estonia accumulated on the Estonian Shelf, and span the transition from the Baltic Shield and Livonian Basin. The section is predominately carbonate with variable siliciclastic content, and numerous boreholes provide good-to-excellent coverage of facies patterns. The correlation of the Baltic regional stages to international stages is relatively well established.

The upper Nabala-Porkuni Stages (equivalent to the uppermost Caradoc to Ashgill Series) consist of seven complete stratigraphic sequences, and the lowstand deposits of an eight sequence (Fig. 1 and 2). The sequences are represented in shelf sections by grain-supported (grainstone and packstone), mud-supported (wackestone, mudstone, and marls), and mixed facies. These occur in shallowing-upward, prograding carbonate shelf successions. Slope sections are predominately mudstones that are locally truncated by submarine erosion surfaces. Basinal deposits are mudstones and marls with greater siliciclastic content in more distal areas.

The lowest sequence (upper Nabala Stage) consists of low-energy mud-supported facies that extend across most of the study area except for some moderate-energy mixed facies associated with updip isopach thicks. In sequence 2 (Vormsi Stage), facies belts are more differentiated and siliciclastic muds are most abundant, probably reflecting tectonic influences. Sequences 3-6 (Pirgu Stage) record the step-wise progradation of shelf carbonates that produced a relatively wide Estonian Shelf. The Pirgu/Porkuni stage boundary is also a sequence boundary marked by localized updip erosion that marks the widely recognized, early Hirnantian sea-level drop. Within sequence 7 (lower Porkuni Stage), shelf sections contain high energy (including cross-bedded grainstone) facies that indicate the relatively low sea level associated with the Hirnantian. The lowstand unit that caps the succession (uppermost Porkuni Stage) is restricted to slope and basin areas, and includes redeposited ooids (probably derived from lower Porkuni strata) and quartz (presumably from the Baltic Shield) sands. This lowstand unit corresponds to the widely documented, latest Hirnantian glacial sea-level drop. The remainder of the sequence occurs at the base of the Silurian section.

The Estonian sequences appear correlative to those defined in Laurentia, suggesting an eustatic sea-level control on sequence timing. However the facies change between sequence 1 and the overlying sequences probably reflect tectonic changes within the Baltic region, perhaps related to the initial convergence of Baltic and Avalonia. This suggests that the controls on facies patterns are a mixture of eustatic and tectonic influences.

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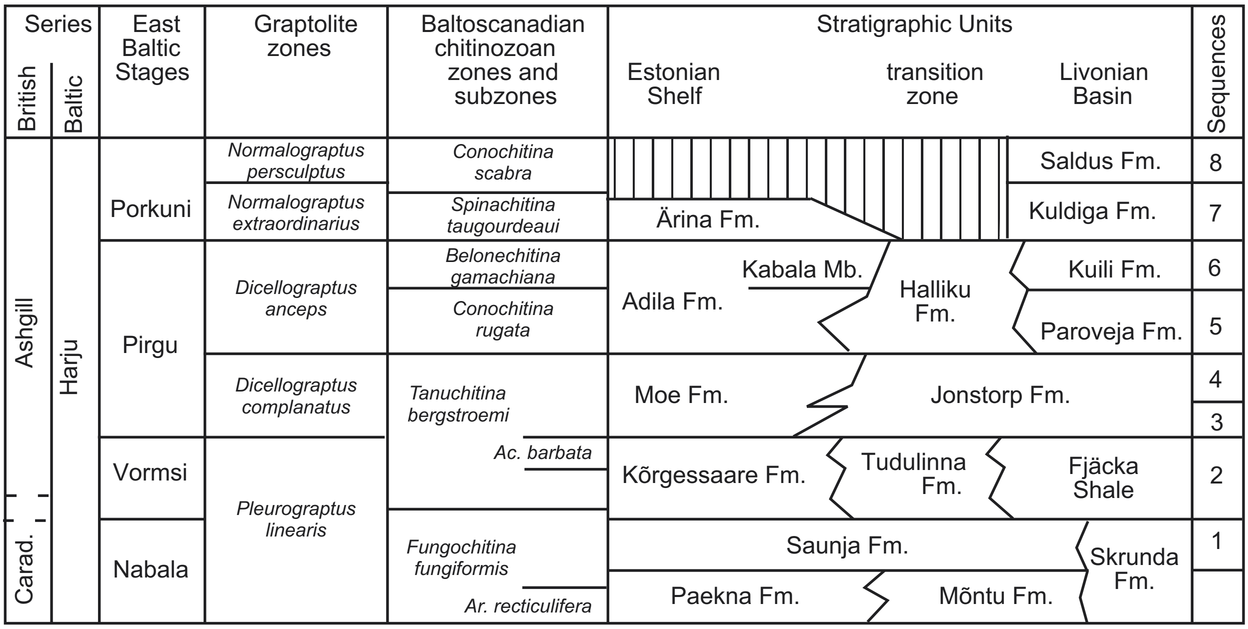

Fig. 1. Stratigraphy of the study interval (modified from Raukas & Teedumäe 1997). Abbreviated chitinozoans are: *Ac. barbata = Acanthochitina barbata,* and *Ar. reticulifera = Armoricochitina reticulifera.*

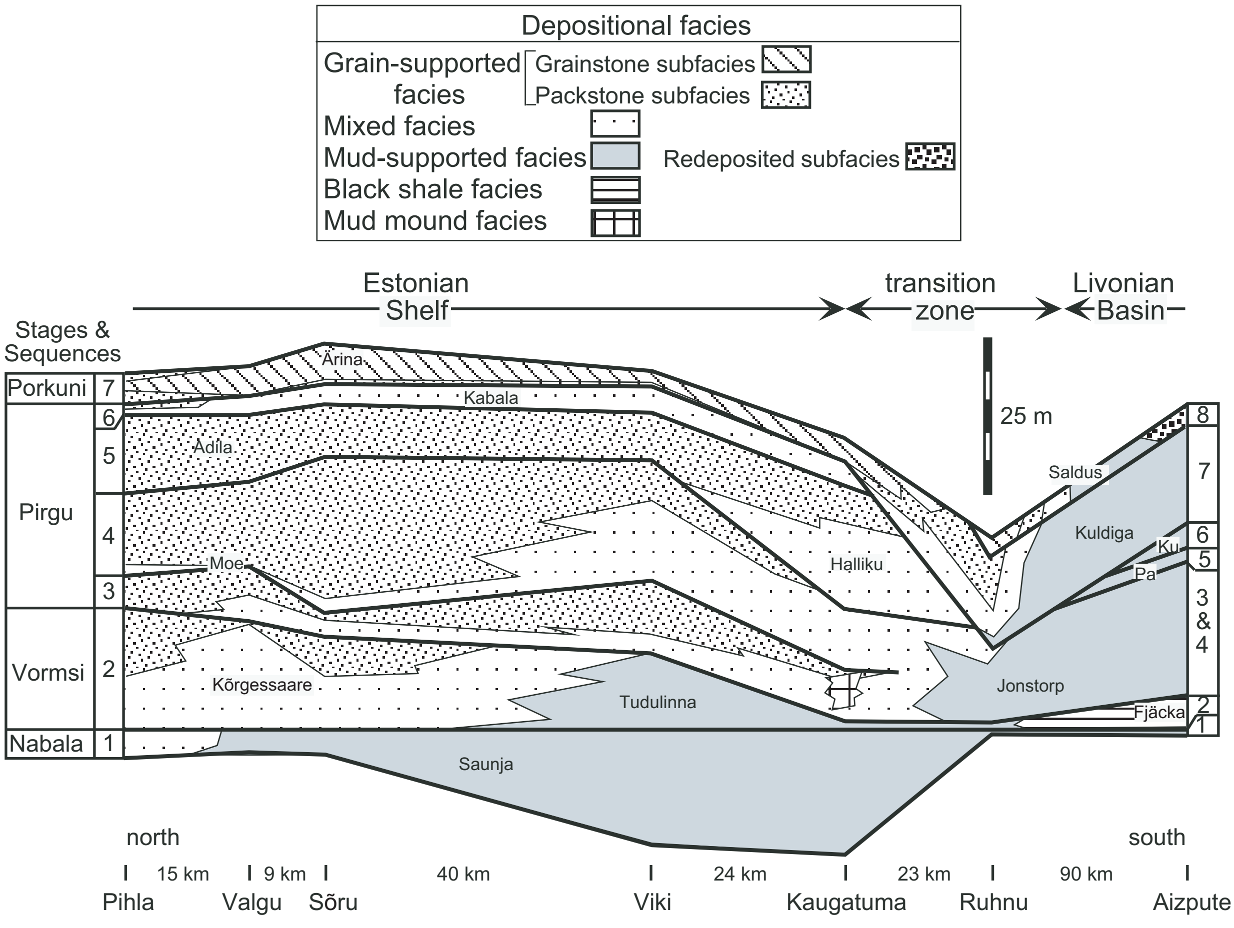

Fig. 2. Facies cross-section. Distances between the cores are scaled except for the Ruhnu–Aizpute distance. Abbreviated formation names are: Ku = Kuili, Pa = Paroveja.

Late Ordovician ripple marks in Vasalemma quarry, NW Estonia

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The occurrence of ripple marks in the Ordovician strata of Estonia is only described for a few localities. For example, they have identified in the Oandu-age sandy limestones of the Saku Member of the Vasalemma Formation (Põlma 1982). Somewhat older ripple marks within the same formation, are well represented in the Vasalemma quarry (and will be visited during the WOGOGOB-2004 excursion). The surface with the ripple marks forms the floor in the northern part of the quarry where it coincides with the upper surface of the Pääsküla Member (Keila Stage). In the drill core sections in the NW Estonia, including the Vasalemma quarry area, the upper boundary of the Pääsküla Member is a pyrite-impregnated discontinuity surface and *Trypanites*-type borings. Several discontinuity surfaces are also common within the upper part of the Pääsküla Member. The real nature of these surfaces, at least of the upper boundary surface, becomes evident due to the extensive excavating works in the largest quarry in the Vasalemma settlement.

In the Vasalemma area, the Pääsküla Member, is a comparatively homogeneous unit of micritic limestone that is 4–5 m thick and that underlies the Upper-Kahula and Vasalemma Formations. The latter formation consists of three lithotypes: grainstones, reefs/mud mounds and interbedded argillaceous limestones, as described in earlier guides (Põlma & Hints 1984; Hints 1990, 1996). The first appearance of pelletal limestones in the Pääsküla Member (Kahula Formation) and the development of mud-mound and reef-like structures in the Vasalemma Formation indicate the essential changes in the sedimentation regime in the northern Estonia. The ripple marks and their morphology on the Vasalemma quarry floor serve as the indicators of the pre-Vasalemma environment.

In the Vasalemma quarry, the area of about 120 m^2 was cleaned using a fire hose to facilitate a detail study of the ripple marks. The ripple marks were measured along two lines, one (W–E) across the ripples and another (N–S) along the ripples. The following measurements were made: 1) the wavelength (L), measured between the highest points on ripple crests, 2) the ripple height (H), measured perpendicular to L in the deepest part between the ripple crests, 3) and 4) the horizontal distances from the top of ripple crests to the deepest point in the trough on both (western and eastern) sides of ripple (WL, EL). The measurements and corresponding statistical analysis reveal only the main characteristics of the ripples due to the strong erosion along the upper surface of ripples.

Description of the ripple marks. The ripple marks on the upper boundary of the Pääsküla Member are oriented roughly in a N–S direction (NE 10°, SW 190°). They have flat-topped, rarely rounded, slightly sinuous and bifurcating crests. The crests are rather broad, the troughs broad and flat. The wave-length (L) is dominantly 20–40 cm (varies from 14 to 75 cm), the height (H) is 0.2–3.9 cm. The wave index (L/H) is 21.14 by 268 measurements. The symmetry index is 1.28 in cases (165 measurements) when the WL is longer than the EL and 1.30 (133 measurements) in cases when WL is shorter than EL. The ripple marks are asymmetrical, but there is no clear tendency of the prevailing direction of asymmetry (see Fig. 1).

The upper surface of the ripple marks is pyritizised. The pyrite is more extensive in the troughs, probably due to more extensive erosion of the surfaces of the crests versus the troughs. The surface is covered by the two types of burrows. No evidence of internal structure or lamination was observed, probably because they were destroyed by intense boring activity. There is no clear trend to the orientation or concentration of fossils debris in relation to the ripples. In some places, eroded colonies of bryozoans occur. The size of these colonies and the possible depth of erosion indicate that the ripples were originally at least 10–15 mm higher. Thus, the actual value of the wave index may turn much lower than indicated above.

Near the buildups in the southern part of the study area, the ripple marks become indistinct, the quarry floor become undulate, and the bedding plane locally disappears under the younger strata.

The ripple marks in the Ordovician rocks exposed in so large area are exciting and important for the interpretation of the nature of the discontinuity surfaces in drill core sections. The ripple marks are similar to wave-current ripples by size and form. Following the interpretation of the ripple and symmetry indexes by Tanner (1967) it is assumed that these wave-current ripples were formed in very shallow water, maybe in breaker or even swash zone. The abundance of the *Trypanites*-type borings suggests the same interpretation.

The ripple marks with a longer wave-length (100–110 cm) occur at a horizon 0.60–0.70 m above the surface described above in the Saue Member (?) of the Kahula Formation that laterally replaces the Vasalemma Formation. The occurrence of the ripple marks at several other stratigrapical levels is suggested by the wavy structure of the fine-grained limestones and grainstones observed in the quarry wall sections.

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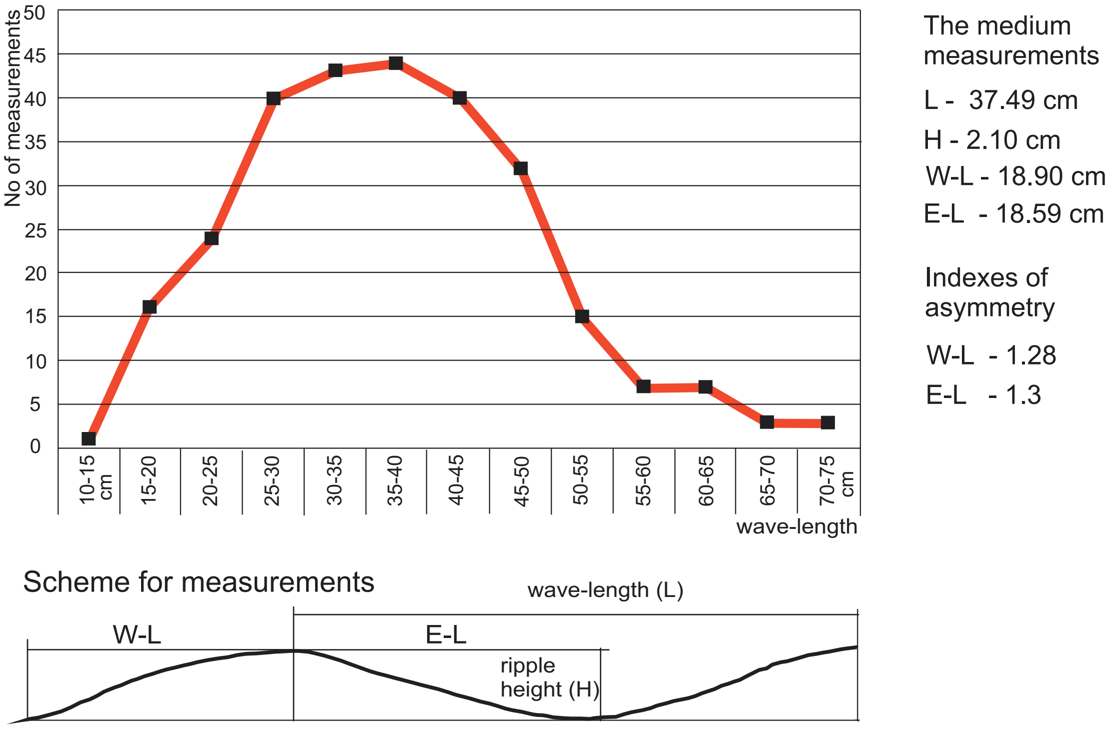

Fig. 1. Frequency curve of wave-length by 5 cm classes (269 measurements).

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Pirgu Stage in the East Baltic: lithotypes, biozonation and problems of correlation

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A number of studies have focused on the correlation of different lithounits of the Upper Ordovician Pirgu Regional Stage, the thickest stage in the Ordovician sequence of the East Baltic, yet the correlation and age of some units are still debated. The stratigraphical arrangement of more than 20 lithounits of the Pirgu Stage proposed here (Fig. 1) is based on the analysis of the rock composition of the units and facies successions, distribution of chitinozoans and ostracods and K-bentonite data. In some cases macrofossils are used. For the first time some formations (Adila, Oostriku, Kuili) are subdivided into two parts.

By the composition, structure and texture, and some other lithological features the rocks of Pirgu age represent four main lithotypes replacing each other in time and space according to the changes in environmental conditions (sea-level, climate).

The first lithotype comprises onshore grey to brownish-grey, micro- to cryptocrystalline seminodular to nodular limestones of early and late Pirgu age, with a variable content of skeletal debris (Moe, Oostriku, Svedasai, Baltinava, Ludza and Taučionys formations). **The second** type of rocks is represented by grey micro- to cryptocrystalline seminodular limestones with sparse skeletal debris and darker spots of dispersed pyrite in the middle of the Pirgu Stage in the central part of the basin (lower and upper parts of the Paroveja Formation). **The third** lithotype of late Pirgu age is represented by grey to greenish-grey, micro- to very fine-crystalline limestones and marls with variable contents of argillaceous material and skeletal debris (Adila, Ukmerge and Kabala formations). The **fourth type** comprises grey or red-coloured, micro- to fine-crystalline marls and argillaceous limestones of early and late Pirgu age, containing sparse skeletal debris or lacking it (Jonstorp Formation including the glauconite-containing Tootsi Member, Halliku, Jelgava, Kuili formations and the middle part of the Paroveja Formation). The lithological similarity cannot be used without additional data for the correlation of sections from different facies belts. So for example, in some cases the micritic limestones included to the Moe Formation presumably belong to the younger Oostriku Formation represented by similar limestones but containing mainly fragments of *Vermiporella*.

The Pirgu Stage comprises the *bergstroemi* (upper part), *rugata* and *gamachiana* chitinozoan biozones, which have enabled the most probable correlation up to now.

The *Acanthochitina barbata* Subzone of the *bergstroemi* Zone is a good marker for the identification of the uppermost Vormsi Stage in different facies belts and different lithounits. *A. barbata* disappears just below the Moe Formation, traditionally considered as the lowermost Pirgu unit. The lower boundary of the Moe Formation is usually defined by frequent occurrence of the alga *Palaeoporella* (in earlier publications *Dasyporella*). Numerous dasycladacean algae (especially *Vermiporella*) have been identified in several other stratigraphical levels in the Pirgu Stage (Männil & Rõõmusoks 1984). A gap in the range of the zonal species *bergstroemi* and *rugata* occurs in several sections in different parts of Estonia. It may have some relationship with the Oostriku Formation, but its exact meaning for stratigraphy of the Pirgu Stage is not yet clear.

The greatest complications arise from the red-coloured rocks (fourth type) forming in some sections the entire Pirgu Stage or its lower half (Jonstorp Formation) and occurring in higher levels within the Jelgava, Paroveja and Kuili Formations. These red rocks are barren of chitinozoans and their correlation in different sections is based on other criteria, such as lithological successions, ostracods, macrofossils and isotopic data.

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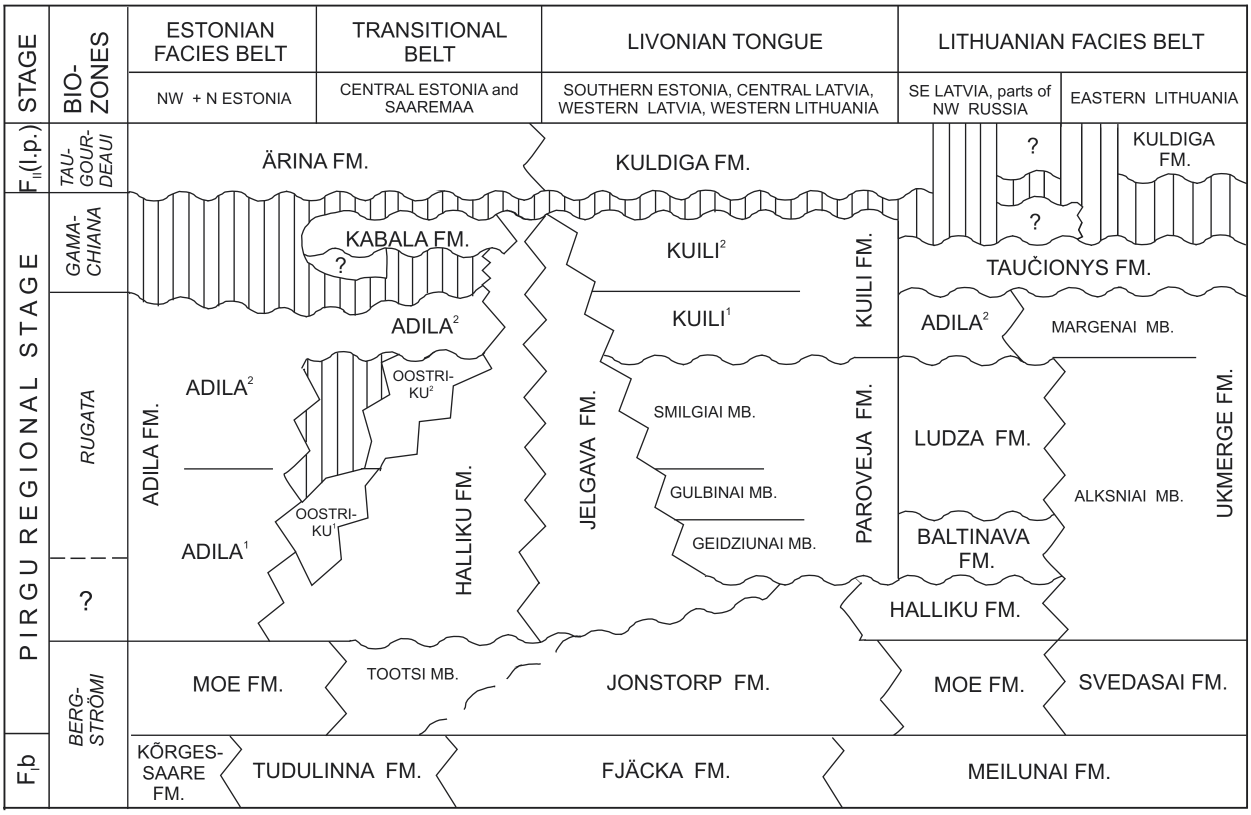

Fig. 1. Correlation of the Pirgu Regional Stage in the East Baltic.

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New palaeontological finds from the Tremadocian of Kadriorg, Tallinn, northern Estonia

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In early spring 2003, a huge excavation was blasted into the bedrock just on the edge of the Baltic Klint in the southern bounds of Kadriorg, Tallinn, to prepare the place for the new building of the Art Museum of Estonia. In this temporary outcrop, the succession from the Cambrian sandstones up to the Middle Ordovician limestones was exposed. Fresh Tremadocian rocks, which can rarely be seen in comparable extent, were sampled for mineralogical and micropalaeontological study. The global Tremadocian Stage can be recognised in northern Estonia based on graptolite and conodont biostratigraphy. In the Kadriorg locality, the Tremadocian succession consists of quartz sandstone (upper part of the Kallavere Fm.), argillite and clay (Türisalu Fm. and Varangu Fm.) and glauconitic sandstone (lower part of the Leetse Fm.; see Fig. 1).

Conodonts. The conodont biozonation provides a precise tool for the subdivision and correlation of Lower Ordovician rocks in Baltoscandia (Viira *et al.* 2001). From the Kadriorg section six samples from the Türisalu, Varangu and Leetse formations were analysed and two conodont zones, the *Paltodus deltifer* and *Paroistodus proteus* zones, were distinguished (see Fig. 1). The three lower samples representing the *P. deltifer* Zone contain light-coloured indigenous conodonts. Samples from the lower part of the Leetse Fm., representing *P. proteus* Zone, yield on the other hand numerous reddish-brown redeposited conodonts and only rare light-coloured indigenous ones. The redeposited conodonts show different generations of the "redeposition index" (see Viira *et al.* in this volume). The conodont fauna of the Leetse Fm. is dominated by *Paroistodus*, a mix between *P. numarcuatus* and *P. proteus*. Indigenous *Drepanodus arcuatus* is also rather common. The occurrence of *Juanognathus*? sp. and *Tripodus* sp. allows precise correlation of the studied section with the *P. proteus* level of the Mäekalda section (Viira *et al.* 2001).

Chitinozoans. Early Ordovician chitinozoans are rare and rather poorly known in Baltoscandia, the oldest specimens coming from the Varangu and Leetse formations (Nõlvak 1999 and unpublished data). Most of the samples from the Kadriorg section are barren of chitinozoans. The only productive sample comes from an irregular lens of soft clay within the glauconitic sandstone of the Leetse Fm. (maximum thickness ca 1 cm) and contains an exceptional chitinozoan assemblage. The samples previously studied from other northern Estonian sites contain as a rule few specimens of up to three species. From the Kadriorg sample, however, several thousands of specimens were recovered representing eight different species: *Cyathochitina primitiva* Szaniawski, *Cyathochitina*? *clepsydra* Grahn, *Lagenochitina longiformis* (Obut), *L. esthonica* Eisenack, *Eremochitina* sp. sensu Nõlvak & Grahn 1993, (pl. V, figs C, E), *Rhabdochitina* cf. *gracilis* Eisenack, *Desmochitina* cf. *ornensis* Paris and rare *Velachitina* sp. The occurrence of *Cyathochitia primitiva* marks the *primitiva* Zone. *Desmochitina* cf. *ornensis*, *Velachitina* sp. and *Eremochitina* are well known zonal species from approximately coeval sections of North Gondwana (Paris 1990), indicating close relationships between chitinozoan faunas of Baltica and Gondwana. Interestingly, the newly recovered Estonian forms are somewhat larger that the corresponding specimens from Gondwana. It remains to be tested whether they in fact represent intraspecific variability and whether the size difference may be attributed to, e.g., different palaeoenvironmental settings.

Scolecodonts. Scolecodonts (polychaete jaws) are common microfossils in the Middle and Upper Ordovician rocks of different parts of the world, the Baltic region inclusive (e.g., Hints *et al.* 2004). Although the group is present already in the topmost Cambrian, the Lower Ordovician record of scolecodonts is remarkably poor. In the Baltic area, the oldest scolecodonts were recorded from the upper Volkhov Stage (lower Middle Ordovician) and the attempts to find material from Lower Ordovician strata have so far been unsuccessful.

The study of samples from the new section at Kadriorg revealed scolecodonts at different stratigraphical levels beginning from the basal portion of the Türisalu argillite (Tremadocian, Pakerort Stage).

Hence, the known range of scolecodonts in Baltica can be extended some 20 million years. This lowest occurrence is represented by rare simple teeth and jaw fragments, which cannot be confidently identified even at the family level. It is still clear that they derive from rather simple (primitive) jaw apparatuses of placognath/ctenognath type. Similar unidentified scolecodonts come from the upper part of the Türisalu Fm. as well as from the Varangu Fm. (Varangu Stage). In the lower part of the glauconitic sandstone of the Leetse Fm. (Hunneberg Stage), no material was found, but this may well be a methodological bias. The clay sample that was rich in chitinozoans contained an assemblage of well-preserved isolated scolecodonts and some fragmentary fused jaw apparatuses allowing description of the material based on multi-element taxonomy and discussion about the affinities of these early polychaetes. One of the two species distinguished represents probably a new genus of conjungaspids, a group that is suggested to share features of both placognath and labidognath apparatuses (Hints 1999). The other is a xanioprionid, possibly attributable to the genus *Xanioprion*, which is common in the Middle Ordovician and onwards. These new data indicate that xanioprionids and conjungaspids might have played an important role in the evolution and diversification of early jaw-bearing polychaetes. It also encourages making further attempts to collect additional material from the Baltic Lower Ordovician.

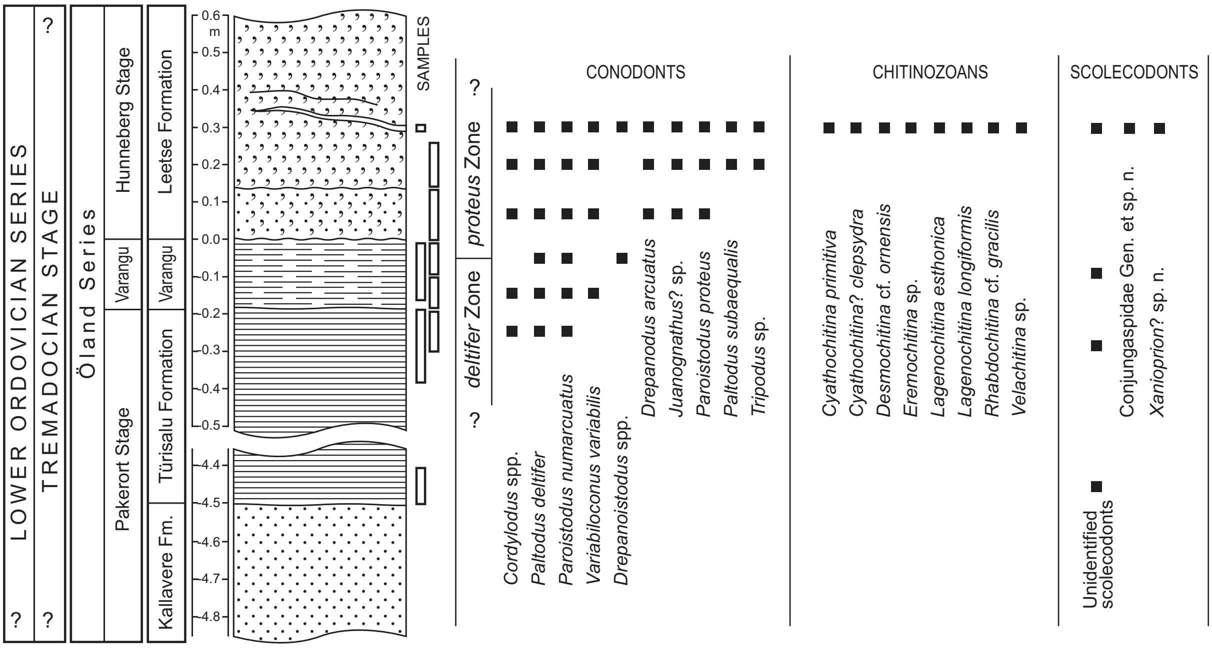

Fig. 1. Stratigraphy of the Tremadocian succession at Kadriorg, Tallinn, northern Estonia, and the distribution of conodonts, chitinozoans and scolecodonts therein.

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Preservational aspects of the Upper Ordovician Fjäcka Shale, central Sweden – exemplified by a medusoid problematicum

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The Fjäcka Shale in Dalarna, Sweden, is a distinct transgressive mudstone facies in an otherwise carbonate dominated Upper Ordovician Succession in Baltoscandia. In addition to the Siljan District the unit is also recognised in Jämtland, Östergötland, Västergötland and possibly in Scania, with a thickness variation ranging between 5.9 m to 0.12 m (Jaanusson 1963; Skoglund 1963; Pålsson 2001). A south-eastern extension of the facies is found in borings in the Livonian tongue of southern Estonia and Latvia, while corresponding beds in the Oslo Region, Norway, are known as the Venstøp Formation, formerly the Lower Tretaspis Shale (4cα). Biostratigraphically the Fjäcka Shale belongs to the upper part of the *linearis* Graptolite Zone of the early late Ordovician Ashgill series, Pusgillian stage (Vasagaardian Stage, Harju Series) (Jaanusson 1963, 1982).

The Fjäcka Shale consists of dark brown to black bituminous shale or black to dark grey mudstone, usually made up of originally laminated beds. Less fissile beds are also present and small nodules (3–4 cm across) appear at certain horizons. The sediments are organic rich and a source rock for the hydrocarbons typically trapped in the carbonate bodies of the Kullsberg and Boda mounds (Spjeldnæs 2000). Despite being formed from a hemipelagic facies of black, organic rich clay and mud, the bottom conditions were not completely anoxic, though probably poorly ventilated. However, traces of fossils at certain levels and complete trilobite moult stages (Ebbestad & Högström 2002) testify to more oxygenated conditions at given intervals. The original lamination of the sediment, along with complete trilobite moult stages, conodont assemblages of 2–3 elements preserved on bedding planes, and an example of a crinoid holdfast on a cephalopod shell, also point to quiet bottom conditions. Spjeldnæs (2000) speculated that some of the bryozoans found could have been washed out in the shale environment from the flanks of the contemporaneous carbonate mounds. These elements may, however, constitute a smaller part of the fossil community.

The fauna of the Fjäcka Shale is one of high abundance but low diversity. Trilobites are the most diverse group, while minute ostracodes (2–3 species, T. Meidla, Tartu, pers. comm.) and phosphatic inarticulate brachiopods (*Hisingerella*) are by far the most abundant. Articulate brachiopods are dominated by '*Strophomena' arachnoidea*, though mostly represented by juvenile forms, but *Chonetoidea* sp. is also common. The problematic scleritome-bearing machaeridians also show a relative high abundance, with at least three species represented, where *Lepidocoleus suecicus* is the most common. Semi-articulated specimens are not unusual at intervals. Gastropod steinkerns occur frequently, mostly minute specimens (1–3 mm). Only some large specimens are known (*Tropidodiscus* sp.). Other molluscs seem to be represented by orthoceratid cephalopods alone. Sporadically bryozoans, crinoid ossicles, and scolecodonts may also occur. A number of "Problematica" have also been collected during the course of this project.

The fossils are most commonly preserved as internal and external moulds, with the original shell material gone, but this varies with the degree of weathering which may reach deep into the ground in places. Although preserved in shale, many of the fossils studied appear to retain most of their original convexity. This preservation is not restricted to the brachiopods alone but also characterizes most groups of organisms with mineralised hard parts.

The medusoid specimen presented here was collected from the Fjäcka Formation at the Fjäcka Rivulet. It is associated with specimens of *Chonetoidea*, *Calymene*, *Hisingerella*, and a sclerite of *L. suecicus* as well as some minute trails that appear impressed through the specimen from underneath. The specimen itself consists of part and counterpart and comprises approximately half of the medusoid "disc", unfortunately without the centre preserved. An estimated diameter of a complete disc is around 60–70 mm. There is only a slight "topography" visible on the disc essentially consisting of a few concentric ridges raised above the surface and increasing in width as they approach the edge of the disc. Scattered over the surface are a number of brachiopods, especially *Hisingerella*, but none of them seem to have been epibionts rather their occurrence on the surface of the disc is considered a coincidence. These brachiopods are common throughout the Fjäcka Shale and can often be seen randomly scattered over the surface of other fossils.

In addition to the large concentric ridges there are minute concentric "ridges" as well as radial "lines" giving the surface a slightly wrinkled appearance. The disc in itself gives the impression of having been a relatively tough structure but without hard parts. There are no indications of a ventral surface with preserved tentacles and such. However, based on the extremely rare occurrences of structures such as those even in confirmed scyphozoan fossils this comes as no surprise.

Unusually the present medusoid is preserved on what appears to be at least a semi-oxygenated bottom when otherwise most medusoids seem to be stranded specimens that at least partially dried out prior to burial (Bruton 1991). Nevertheless this appears to be the case here. The preservation is also in contrast to the impressions of medusoid fossils found in mature sandstones representing shallow water beaches or shore banks (Hagadorn *et al*. 2002). Neither is the preservation of comparable type to the exceptional soft body preservation by clay mineralization seen in for example the Late Ordovician Soom Shale of South Africa (Gabbot 1998). The present case have no preserved muscle tissue and/or other details of the internal anatomy, instead we largely seem to have had a formation of external moulds possibly in combination with some replacement by clay minerals.

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Comparative biotite compositions in the Late Ordovician Deicke, Millbrig and Kinnekulle K-bentonites: a test of consanguinity

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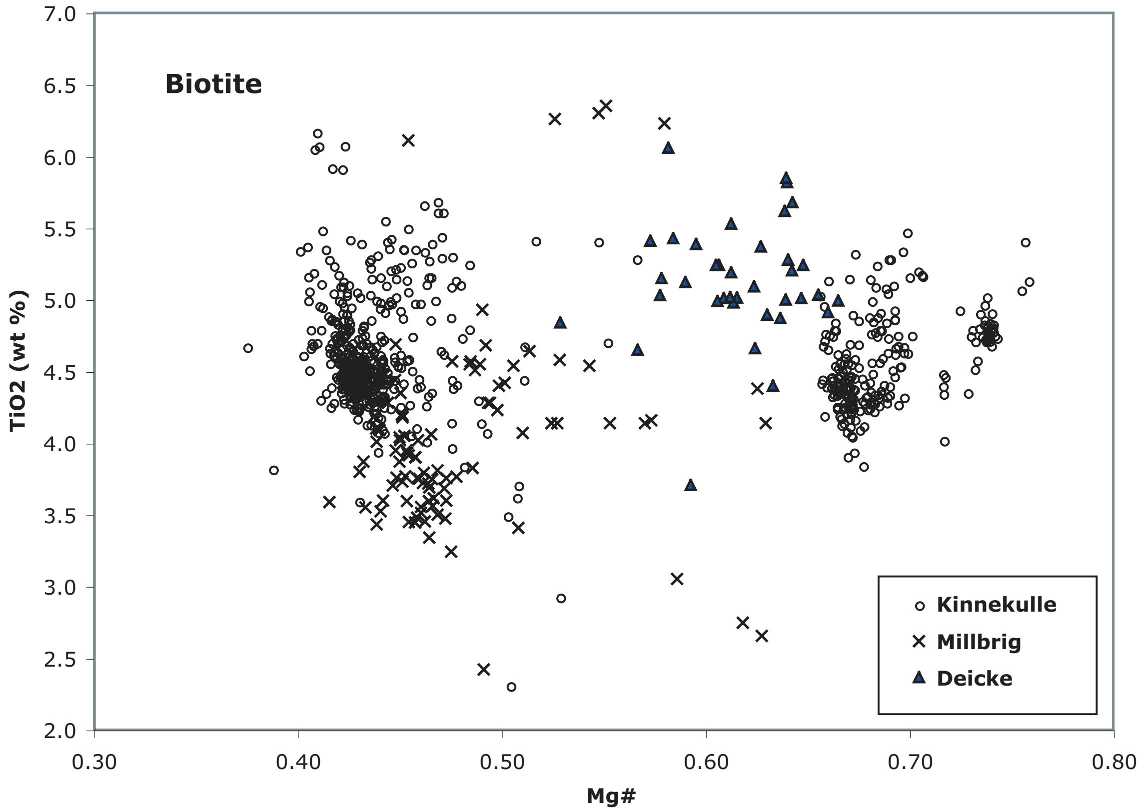
Fine-grained volcanic ash may be deposited many hundreds of kilometers from the source in highly explosive eruptions, but its preservation as pyroclastic units is dependent upon the accumulation in a basin where background sedimentation rates are slow, and where reworking and erosion are minimal. Generally these ash layers were deposited over short time intervals into shallow marine environments of platform limestones and shales. Thus these beds can still provide evidence of their volcanic origin in spite of alteration to bentonite during early diagenesis and the progressive illitization to form K-bentonites during the late diagenesis. Since they were deposited over large areas in a short time period, they have potential for event-stratigraphic correlation and for biogeographical, paleoecological and sedimentological studies on both local and regional scale. They also have tectonomagmatic significance because they preserve mineralogical and geochemical evidence of their origin.

Late Ordovician K-bentonites in eastern North America and northwestern Europe are characteristically clay-rich, but also contain primary minerals in the form of isolated and euhedral phenocrysts such as quartz, biotite, plagioclase, K-feldspar, ilmenite, apatite, zircon and magnetite. Major and trace element analysis of whole-rock Ordovician K-bentonites indicates that the parental magmas consisted of a calc-alkaline suite ranging through andesite, rhyodacite, trachyandesite and rhyolite. The chemical compositions indicate a tectonomagmatic setting characterized by destructive plate margin volcanics. At least 50 K-bentonite beds occur in Ordovician strata of eastern North America and around 150 beds are recorded in Baltoscandia (Bergström *et al.* 1995). The Deicke and Millbrig K-bentonites are two of the most widespread and persistent of the many beds of altered volcanic ash that occur in middle to upper Mohawkian/Champlainian (= lower to middle Caradoc) strata of eastern North America. The Kinnekulle K-bentonite is the thickest and most widespread among the many K-bentonites in the Ordovician of Baltoscandia and it is generally present in sections having an apparently uninterrupted Jõhvian-Keilan succession. The possibility of a common source for these ash beds from North America and northwestern Europe was suggested by Huff *et al.* (1992). They presented biostratigraphical, geochemical, isotopic and paleogeographic data that indicated that the Millbrig K-bentonite, a widespread isochron and marker bed in North America, and the Kinnekulle bed in Baltoscandia are coeval and probably derived from the same eruptive event. Additional information supporting this interpretation was recently presented by Saltzman *et al.* (2003).

However, Haynes *et al.* (1995) suggested that the proposed intercontinental correlation of the Millbrig and Kinnekulle beds is suspect. They suggested that the utility of the discriminant function analysis of whole rock compositions, which are obtained by neutron activation analysis for the correlation of K-bentonite beds, is not valid for large-scale regional or global correlations. They maintained that uncertainty results from the variable mobility of several major and certain trace elements during diagenesis might result in regional shifts in bulk composition. Haynes *et al.* (1995) studied the compositions of volcanically generated biotite phenocrysts that had survived diagenesis and concluded that they are more reliable as specific bed indicators than bulk rock composition. They reported a compositional difference between Kinnekulle and Millbrig biotites with respect to their FeO, MgO, Al_2O_3, MnO, and TiO_2 content. They further suggested that these variations represent separate eruptive events. However, Haynes *et al.* (1995) used data from only one Millbrig site in North America and one Kinnekulle site in Baltoscandia, so they failed to evaluate lateral variation as well as within bed variation in biotite compositions. Furthermore, they culled some of the analyses, leaving a data set representative of both the modal and average composition of biotite in each sample.

Here we present a comprehensive study of K-bentonite biotite composition covering a more extensive geographic and stratigraphic range for these Ordovician beds. After removing several hundred analyses due to alteration or inadequate preparation, a total of 666 Kinnekulle biotite analyses, 97 Millbrig biotite analyses and 39 Deicke biotite analyses representing 32 separate localities provide the most comprehensive view to date of the nature and extent of internal compositional homogeneity of these widespread Laurentian and Baltoscandian ash beds. The data show clearly that the Kinnekulle and Millbrig are multiple event ash beds, some parts of which are indistinguishable from one another. Fig. 1 shows the bivariate distribution of TiO_2 vs. the Mg number (MgO/(FeO + MgO)). The Mg number is commonly used as a fractionation index in the evaluation of igneous rocks and it here provides a very dramatic way of visualizing the compositional zonation of the Millbrig and Kinnekulle beds. The Deicke appears to be a single event deposit. On the basis of this and other similar plots we conclude that portions of the Millbrig and Kinnekulle beds are compositionally indistinguishable. Published age dates are inconclusive as to the true ages of each bed and are thus permissive of a common age and a common origin.



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New data on the Volkhovian megistaspid trilobites from Putilovo, western Russia

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A large-scale bed-by-bed sampling of macrofossils was undertaken in the Putilovo quarry 70 km east of St. Petersburg during two field seasons in 1999 and 2002. The sampling targeted the entire carbonate succession from a level near the base of the Billingen Stage to a level within the uppermost part of the Kunda Stage. The megistaspid material under study mainly derives from the Volkhov Formation, but the Päite Beds (upper part of the Billingen Stage) are also included. The material contains several hundreds of specimens and is far larger than the materials previously studied by Schmidt (1898, 1906) and Balashova (1976). Besides, the precise sampling level of all specimens within the succession is also known.

The aim of the study is to undertake a taxonomic revision of the megistaspid group, comprising the genera (or subgenera) *Megistaspis*, *Paramegistaspis*, *Rhinoferus* and *Megistaspidella*. The group has previously been treated by a number of authors including Schmidt (1898, 1906, 1907) and Balashova (1966, 1976), both focussing on East Baltic material, as well as Tjernvik (1956, 1972, 1980), Jaanusson (1956), Nielsen (1995) and Hoel *et al.* (2002) among others studying mainly Scandinavian material.

The chronostratigraphic zonation of the Hunneberg, Billingen and Volkhov stages of Scandinavia is based on the distribution of megistaspids and it is hoped that the restudy of the east Baltic faunas will provide useful ties for correlation, both across Baltoscandia as well as within the East Baltic region.

The taxonomic revision has just started and so far only the *Rhinoferus* group [= *M. hyorrhina* species-group sensu Schmidt (1906)] has been investigated in greater detail. The available material of *Rhinoferus* includes 5 articulated specimens, 1 cephalon, 62 cranidia, and 102 pygidia plus numerous free cheeks and hypostomes, which, however, have not been investigated as yet and thus the generic identification (and amount of material) is not confirmed.

Schmidt (1906) divided the *M. hyorrhina* species group into the variants *typica*, *kolenkoi*, *mickwitzi* and *stacyi*. Balashova (1976) ranked these taxa as independent species. At this preliminary stage of investigation we are inclined to agree with Balashova (1976) that the forms should be separated at species level, but it has (at least so far) proved impossible to distinguish *R. kolenkoi*, whereas the other three species exhibit cephalic differences, notably concerning the outline of the glabellar “horn”. So far the pygidia contains no diagnostic features of proven use to allocate them to species. We suspect that the material previously attributed to *kolenkoi* belongs to *R. hyorrhina*, but this needs final verification by studying the type material. The preliminary data suggests that *R. hyorrhina* ranges though the upper Saka to lower Zheltyaki members, and then reappear sparsely in the lower part of the Frizy Member. *R. stacyi* is common in the Zheltyaki and lower Frizy members, whereas *R. mickwitzi* is present only in the lower Frizy Member. Balashova (1976) also reported *R. mickwitzi* from the $B_{II}\alpha$ zone; we have not found any specimens at such low level. It is peculiar that Schmidt (1906) only reported *R. stacyi* from Estonia; this species is very common in the main upper part of the Zheltyaki Member at Putilovo and overall constitutes a greater share of the *Rhinoferus* material than *R. hyorrhina*.

Megistaspis (*Rhinoferus*)	$B_{II}\alpha$	$B_{II}\beta$	$B_{II}\gamma$
hyorrhina	**Z ?**	**X / Y / Z**	**X / Y / Z**
kolenkoi		**X / Y**	**Y ***
mickwitzi	**Y ***		**X / Y / Z**
stacyi		**X # / Z**	**X # / Y / Z**

X = Schmidt (1906) # = only in Estonia
Y = Balashova (1976) * = rare
Z = This study, based on cephalic material

Table 1. Distribution of *Rhinoferus* in the East Baltic area according to various authors. Zonation following Lamansky 1905.

The distribution of megistaspids at Lynna River (near Volkhov), where only the upper part of the Zheltyaki and the Frizy members were sampled, matches the occurrence pattern seen in the equivalent interval at Putilovo, but suggests that the Lynna section is less complete.

The taxonomic revision of *Megistaspis* itself has just been initiated when this abstract was submitted; a status of the study will be presented at the WOGOGOB meeting in Estonia, May 2004.

The study is part of the joint project between the State University of St. Petersburg (Andrei Dronov) and the University of Copenhagen, focussing on the stratigraphy and sea level changes in the Volkhov-Kunda interval. Andrei Dronov's help organising the fieldwork is gratefully acknowledged.

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Carbon isotope dating of several uppermost Ordovician and lower Silurian sections in the Oslo Region, Norway

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Carbon isotopes are increasingly applied in stratigraphy. A more or less complete carbon isotope trend for the late Ordovician and Silurian has been ascertained based on the studies carried out mainly in Baltica and Laurentia, but also elsewhere (for a summary see Kaljo *et al.* 1998; Brenchley *et al.* 2003). There remain some small gaps (e.g. at the Ordovician–Silurian boundary; Kaljo *et al.* 2001) or debates about dating the isotope shifts (Melchin *et al.* 2003), but nevertheless we do believe that the general pattern can serve as a stratigraphic tool for correlation. It should be stressed that it is the shape of the curve that is most important, not so much actual values of the $\delta^{13}C$, because the values may depend to some extent on the facies characteristics of the rocks measured.

The main positive excursions of carbon isotope ($\delta^{13}C$) values used in this report as time markers are as follows (in brackets East Baltic values according to the works cited above, and Kaljo & Martma 2000): early Ashgill (2.5‰), Hirnantian (4–7‰), early Aeronian (3.7‰), early Telychian (2.7‰) and early Wenlock (4–6‰). Very characteristic are also negative shifts in the late Rhuddanian (-1.2‰) and late Aeronian (-1.3‰).

Using this general pattern of the carbon isotope trend as a standard, we will discuss the applicability of isotopes in the correlation and dating of outcrops in the Norwegian study area, where conditions of regional metamorphism are locally less favourable for the preservation of the primary isotopic signal than in Estonia. At least the occurrence of tiny amphibole needles in the limestones of the Skien–Langesund area calls for attention. In all, 150 bulk rock carbon isotope analyses were made from 21 localities, among these 10 from the Oslo-Asker, 6 from the Ringerike and 5 from the Skien–Langesund areas. Samples were tied to the local stratigraphical columns and schemes available (Worsley *et al.* 1983; Owen *et al.* 1990, and some unpublished data).

Analysis of the above isotope data and consideration of the Baltic and Norwegian stratigraphical background revealed the following relationships.

The Hirnantian peak is best represented in the Etage 5b at Konglungen (Asker), where the $\delta^{13}C$ values reach 5.9‰. The peak is rather typical, rising from 1.1‰ in the Etage 5a and falling to 1.4‰ in the lowermost Silurian. The positive shifts established in the sections at Kalvøya and Semsvannet from rocks assigned to the Etage 5b show respectively only 2.9 and 3.4‰ as peak values, indicating an incomplete excursion and a possible gap in the upper part of the Hirnantian or some other disturbing reason.

The $\delta^{13}C$ values identified at Gunnekleiv (Herøyavegen) and Skrapekleiv of the Skien–Langesund area are even smaller, reaching 1.3‰ at the top of the Upper Limestone Member and remaining below 0 (-0.3…-0.9‰) in the underlying Middle Siliciclastic Member (terminology from Rønning 1979, PhD thesis). The latter values were measured in the *Holorhynchus* coquina interbeds occurring in the member. At Kulten, the beds with *Holorhynchus*, the overlying thin limestone bed and the Silurian rocks exhibit negative values. Brenchley *et al.* (1997) demonstrated that the beds containing *Holorhynchus* belong to the pre-Hirnantian part of the Ashgill. The isotope data support this conclusion — only a small part of the beds, traditionally belonging to the Etage 5b in the Skien–Langesund area, can be considered as the Hirnantian and a gap might be supposed before the Silurian.

In Ringerike, the Etage 5a and the junction of the 5a and 5b of the Stamnestangen outcrop revealed normal values close to 1‰, but the Ullerntangen outcrop, usually assigned to the Etage 5b, showed the $\delta^{13}C$ values of 1.2‰ in the mounds and 0.1–0.7‰ between them. The Ullerntangen values are too low for the Hirnantian.

The early Wenlock excursion is clearly expressed in all studied sections embracing the corresponding rocks, e.g. Malmøya in Oslo (max. $\delta^{13}C$ values reach 5.2‰), Garntangen (6.3‰), Loretangen (5.7‰) and Storøya (6.6‰) in Ringerike, and Frednes (4.3‰) in

Porsgrunn. We can conclude that this peak is a very stable marker also in the Oslo Region allowing some adjustments of correlation.

The Llandovery of the area showed no early Aeronian excursion, but late Rhuddanian and late Aeronian negative shifts were recorded at Skytterveien (-2.4‰), Øyekast (-1.2‰), etc. The interpretation of these curves, as well as of that at Solhaugveien, is complicated and needs the support of additional biostratigraphic data.

In summary, both major carbon isotope peaks are well expressed in the Oslo Region and allow the use of the isotope information for dating and interpretation of sections. Minor shifts need additional studies before these can be fully understood, as do some other questions.

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Metabentonites of the Pirgu Stage (Ashgill, Upper Ordovician) of the East Baltic

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Twenty-two metabentonite (MB) samples from the *complanatus* graptolite zone, Pirgu Stage, Ashgill, of 15 Estonian cores and one Latvian core (Aizpute-41) were studied (Fig. 1). For correlation purposes the coarse-grained fraction containing pyroclastic sanidine was separated from MB and measured by X-ray diffractometry, and mol% of the Na-compound was calculated using the method of Kiipli & Kallaste (2002).

The investigation of MB yielded three correlative layers according to the content of the Na-compound of pyroclastic sanidine, and two single MBs, which do not correlate with the others.

1. The group with 37–39 mol% Na-sanidine forms the lower correlated eruptive layer. This layer correlates the Moe Member with the Tootsi Member and Jonstorp Formation (Fm.) (compare the Rabivere and Soonlepa cores with the Põltsamaa and Are cores, and with the western Latvian Aizpute core in Fig. 1B).
2. The group with 43–44 mol% Na-sanidine forms another eruptive layer. Metabentonites of this group are found in the Põltsamaa, Pärnu, Valga, Viljandi, Varbla, Eikla, Kaugatuma, Laeva-4, Laeva-18, Tartu and Aizpute cores. This MB layer correlates the lower Kabala Member with the Halliku Member, Jelgava Fm. and upper Jonstorp Fm. (compare the Põltsamaa core with the Pärnu, Valga and Aizpute cores in Fig. 1B).
3. The group with 34–35 mol% Na-sanidine forms the uppermost eruptive layer, found in the Põltsamaa, Pärnu, Valga and Taagepera cores (Fig. 1A). This layer correlates the uppermost Kabala Member (Põltsamaa and Pärnu cores) with the Jelgava Fm. (Valga core) and supposedly with the uppermost part of the Jonstorp Fm. (Taagepera core) (Fig. 1B). In the Taagepera core, the identification of MB host rock as belonging to the Jonstorp Fm. is unclear, because of the alternation of red and greyish beds.
4. In the Ruhnu core the Pirgu Stage (depth interval 619.1–631.1 m) is entirely represented by the red limestone of the Jonstorp Fm. The recorded MB layer (depth 628.05 m) reveals a low broad peak in X-ray diffractograms and is badly determined in relation to the Na-sanidine content. This layer is not correlated to any other layer.
5. Another single MB comes from the Moe Member of the Käbi core. This layer is badly characterised by Na-sanidine due to a small amount of grain fraction. Metabentonites from the Ruhnu and Käbi cores may be coeval, but the correlation is not yet proved.

In the Ashgill Jerrestad Fm. of Sweden five metabentonite layers have been recorded in the Lindegård core (Jaanusson 1963).

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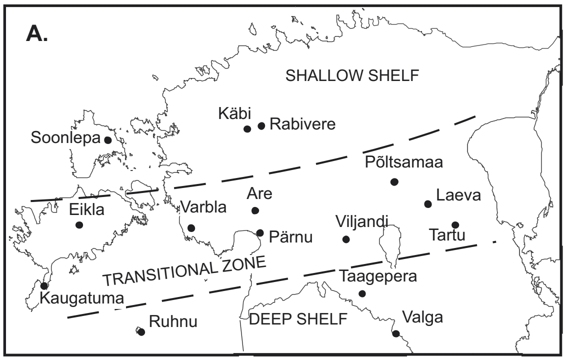
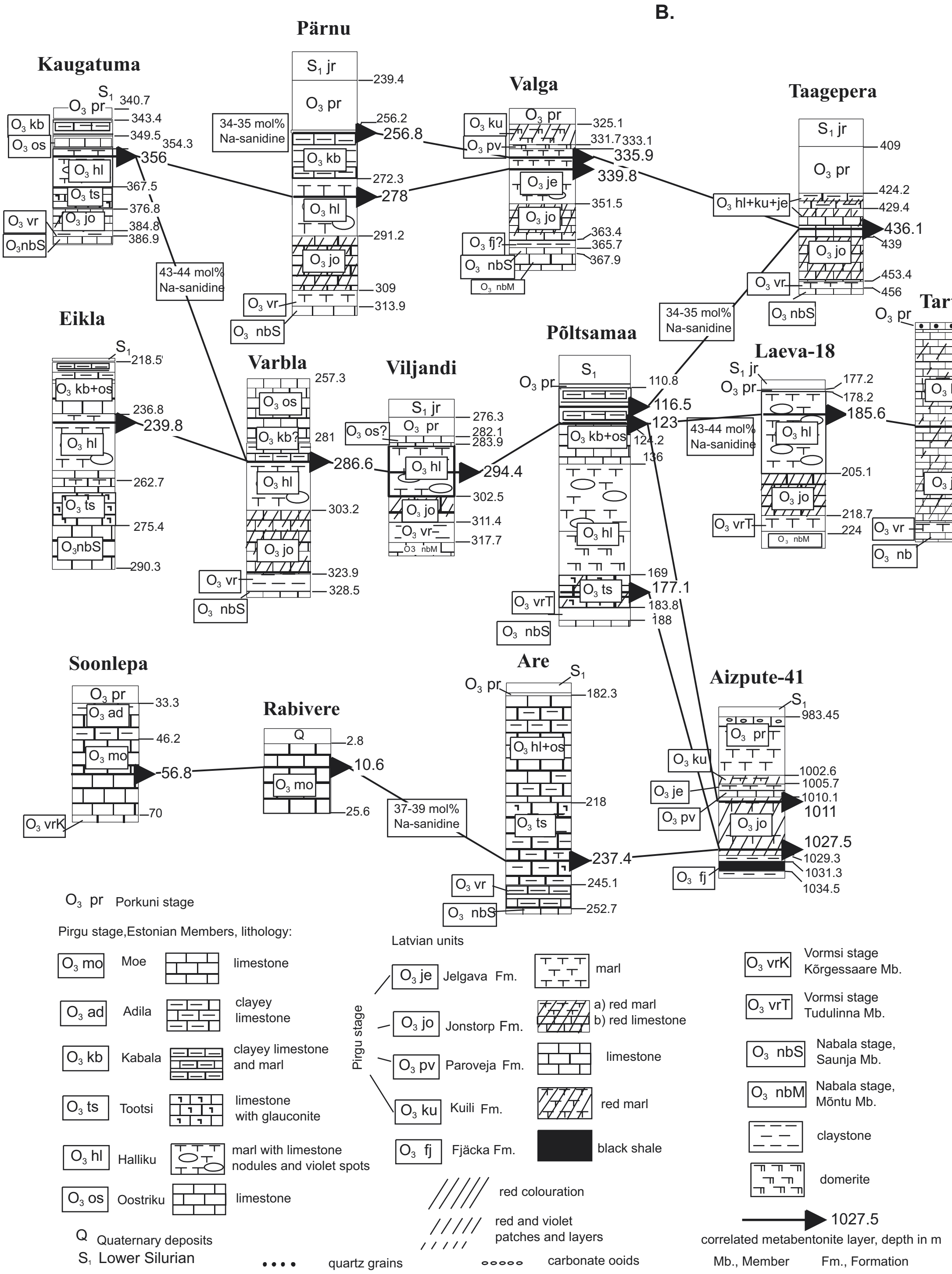


Fig. 1. A — Location of boreholes (the Aizpute-41 core is situated southwards, out of the borders of the figure); B — Metabentonites of three correlated layers of the Põltsamaa core. Na-sanidine mol% 34-35 of the upper, 43-44 of the medium, and 37-39 of the lower layer.



Areal variations in the composition of the Kinnekulle altered volcanic ash bed (Caradoc) – relations to sedimentary facies

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XRF (main chemical components) and XRD (main minerals) study of altered volcanic ash of the Upper Ordovician Kinnekulle bed in well-preserved Estonian and Latvian sections was carried out. An association of K-feldspar and illite-smectite was recorded at Pääsküla subsurface tunnels (Hints *et al.* 1997), where the bed has asymmetric vertical zonation as shown by field observations and laboratory analyses. The composition of altered ash varies regularly from northern Estonia to Latvia and Lithuania, depending on facies conditions in Ordovician normal marine shelf sea. Ash beds in shallow shelf limestones contain the association of illite-smectite with K-feldspar (K_2O content varies from 8.0 to 13.9%). Limestones in the transition zone contain altered ash dominated by illite-smectite (K_2O 6.0–7.7%). In deep shelf marlstones and shales of south-eastern Estonia, Latvia and Lithuania the volcanic ash bed consists of illite-smectite with kaolinite (K_2O 4.1–5.8%). Illite-smectite in Estonian sections contains 18–30% expandable layers (Kirsimäe *et al.* 2002). The K_2O content varies from 6.0 to 6.2% in Gotland sections, which corresponds to the K_2O content in the Estonian transition zone. Much lower values (K_2O 2.4–4.0%) have been recorded at Kinnekulle in mainland Sweden. Altered volcanic ash in this region contains more expandable illite-smectite than in Estonia. In the Oslo region, in Scania and Bornholm island, which are located on the ocean side of the palaeoshelf, K_2O values are 6.2–7.3%.

The following arguments speak in favour of early sedimentary diagenetic control on the formation of authigenic silicate minerals:

1. Regional variations in the composition of altered volcanic ashes concordantly with confacies belts.
2. Asymmetric zonation of altered ash beds suggests that dissolution-recrystallisation occurs mainly in an asymmetric environment between older sediment and open seawater.
3. Chertification of limestone often takes place below, but rarely above thick (more than 50 cm) ash beds, proving that main dissolution-recrystallisation processes occurred commonly before the deposition of the next sediment layers.

The following arguments contradict main late diagenetic control on the formation of authigenic silicates from volcanic ash:

1. Light colour of conodonts proves that palaeotemperatures (high temperature around 120°C is necessary for the illitisation of smectite) have never reached more than 50–80°C.
2. Excellent preservation of biotite and magmatic sanidine proves that these minerals (considered as a common source of potassium in the process of the illitisation of smectite) have not been decomposed. Therefore no potassium source exists inside the volcanogenic sediment.
3. Altered volcanic ashes richer in potassium occur in limestones with a low potassium concentration. Therefore the surrounding sediment did not serve as a source of potassium.
4. Frequent vertical alternation of volcanic beds with a different composition (feldspar, illite-smectite, kaolinite) in sections within a few metres proves that pervasive late diagenetic brines are not important in determining chemical transformation of volcanic ashes.

The following argument supports late diagenetic control on mineralogical changes:

1. Age determinations by the K–Ar method have revealed Early Devonian ages of 320–380 my for the fine clay fraction from the Kinnekulle altered volcanic ash bed (Kirsimäe *et al.* 2002). The same age has been determined from the fine fraction of Lower Cambrian clays in Estonia.

The described regularities allow us to conclude that the formation of authigenic silicates from volcanic ash was controlled by sea- and pore-water chemistry, which depended on water depth and other facies parameters. Volcanic glass and seawater were main potassium sources. Early Devonian ages determined from fine clay fractions probably point to a moderate thermal event (below 80°C), which released argon from fine particles. In Norway and Sweden, thermal influence due to tectonic and magmatic processes may have been of importance, causing transformation of primary authigenic silicates.

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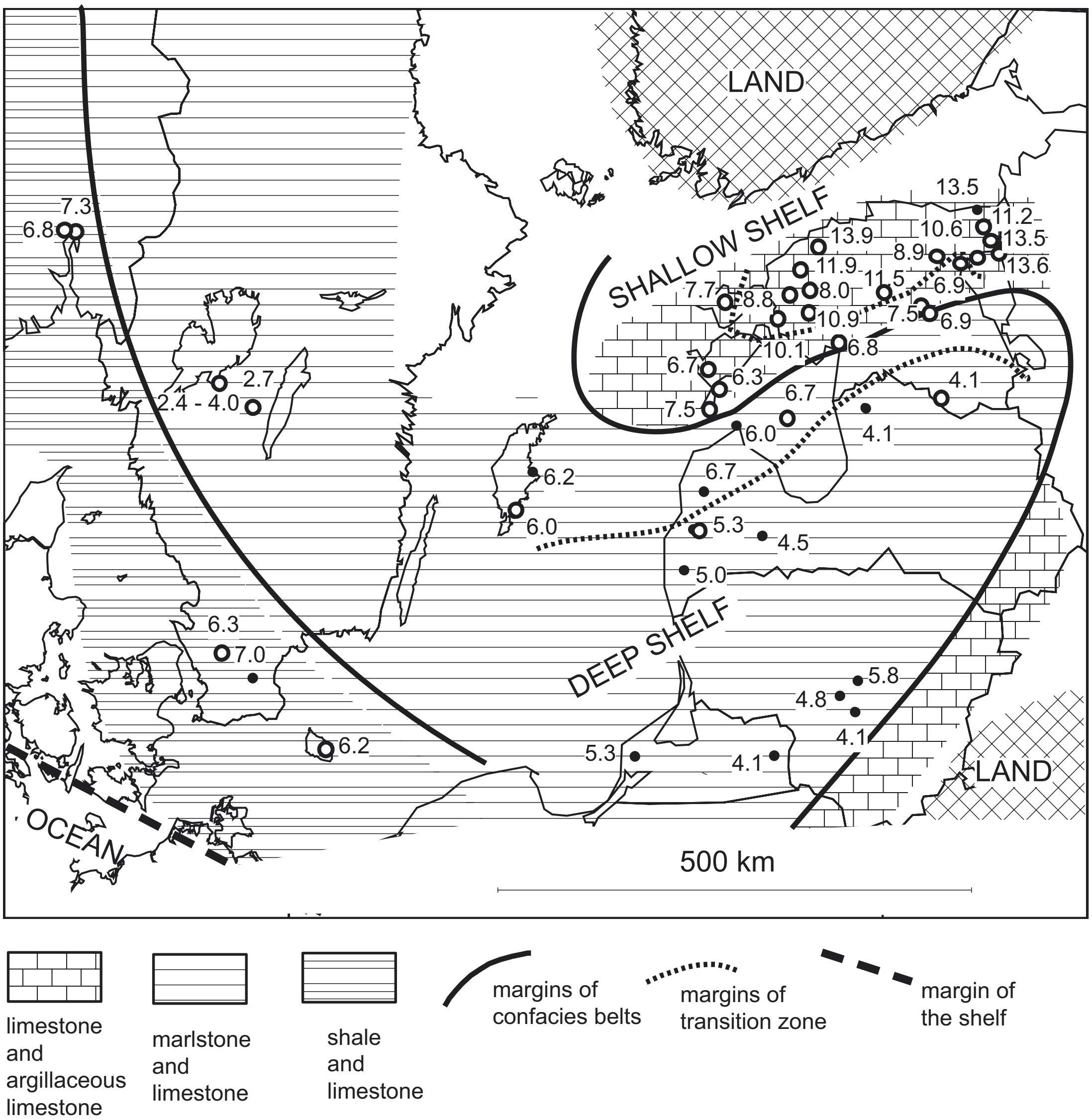

Fig. 1. The K_2O content of the Kinnekulle altered volcanic ash bed. Numbers show the K_2O content in %. Open circles mark the Kinnekulle bed. Filled circles mark the Caradoc beds whose correlation with Kinnekulle is not proved. Confacies borders after Jaanusson (1995).

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Chloritization of Upper Ordovician Pirgu bentonites – source material or diagenetic environment?

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The sedimentary sequence of the Palaeozoic Baltic Basin contains numerous bentonite layers whose composition is dominated by the mineral assemblage of illite-smectite–K-feldspar–kaolinite. The contents of these minerals may vary between individual bentonite layers as well as laterally from almost pure illite-smectite to K-feldspar and/or kaolinite end-member compositions, but the assemblage remains principally the same. In this respect the bentonites of the Upper Ordovician Ashgill Pirgu (Regional) Stage are exceptional and unique in the Baltic Basin. The clay mineral composition of these bentonites is characterized by the chlorite-smectite (corrensite) and illite-smectite assemblage. The micritic-bioclastic to argillaceous limestones of Pirgu age in the northern Baltic Basin include up to three (four?) individual bentonite beds, all of which contain chlorite-smectite and/or corrensite minerals. In this contribution we present preliminary data on the clay mineral composition of these beds.

The clay fraction (<2 μm) of 21 samples studied contains random (R0) mixed-layered chlorite-smectite and/or R1 ordered corrensite and corrensite-chlorite type phases together with illite-smectite. The chloritic phases are the most abundant clay minerals in the majority of samples, but also illite-smectite may dominate in the mixture with minor chlorite-smectite/corrensite. The occurrence of a R1 ordered corrensite (0.5/0.5 interlayered chlorite and smectite mineral) phase is confirmed by the expansion of the superstructure *d*(001) spacing from 29 Å in air-dried state to 31 Å in EG saturated state. The heating of the corrensite-rich sample at 500 °C for 1 h caused the collapse of the spacing to 24 Å. The proportion of smectite layers in mixed layer chlorite-smectite is according to NEWMOD modelling 0.2–0.4 and probably 0.6–0.7 in R0 and R1 (corrensite-chlorite) ordered minerals, respectively.

Chlorite-smectite and corrensite are trioctahedral clay minerals characterizing evaporitic- and volcanoclastic sedimentary-diagenetic, and hydrothermal alteration environments (e.g., Reynolds 1988; Inoue 1995). Smectite-to-chlorite transformation in Mg–Fe-rich rocks/sediments begins usually with the formation of saponite-type Fe-rich smectite, which in progressive diagenesis or hydrothermal alteration transforms into corrensite (Reynolds 1988). The next stage in corrensite-to-chlorite conversion is the growth of chlorite layers in corrensite to form discrete chlorite domains in a corrensite matrix (Beaufort *et al.* 1997). Corrensite, however, can form directly under hydrothermal conditions at temperatures between about 100 and 200 °C (Inoue & Utada 1991).

In alternating pyroclastic sedimentary sequences the formation of dioctahedral smectite and subsequent smectite-to-illite conversion is related to acidic volcanic materials, whereas the trioctahedral smectite (saponite) and chlorite-to-smectite transformation takes place at the expense of the basic Mg–Fe-rich volcanics (e.g. Son *et al.* 2001). Chlorite and chlorite-smectite are also found in Ordovician and Silurian K-bentonites in North America, British Isles and rarely in Baltoscandia (e.g., Bergström *et al.* 1992, 1998). Chloritic minerals in these bentonite beds are suggested to form during alteration under low-grade metamorphic conditions (Krekeler & Huff 1993).

However, the origin of chlorite (chlorite-smectite, corrensite) in Pirgu K-bentonites remains unclear. The Palaeozoic sedimentary sequence of the northern Baltic Basin shows no signs of low-grade metamorphic alteration or even deep diagenesis. Therefore, we suggest that the clay mineral composition of Pirgu bentonites is controlled either by the original pyroclastic material composition or by a specific early diagenetic environment during the volcanic glass devitrification, both of which have provided high Mg needed for saponite-type smectite formation and consequent saponite-to-chlorite transformation.

The trace element composition of whole rock and analysis of glass melt inclusions of Ordovician–Silurian bentonites suggest the rhyolite–rhyolite-dacite and (rhyo-)dacite–trachyandesite composition of the source magma (Huff *et al.* 1996; Kiipli & Kallaste 1996). However, the saponite-to-chlorite series have been found to alter diagenetically or hydrothermally from more mafic andesite(-basalt) and andesitic pyroclastics (e.g., Chang *et al.* 1986; Inoue & Utada 1991; Son *et al.* 2001). Alternatively, the chlorite-smectite (corrensite) and chlorite formation in alkaline lacustrine or marine evaporite environments is controlled by high pH and Mg^{2+} activity, which may shift the activity ratio of Mg^{2+}/H^{+} sufficiently to precipitate brucite-like interlayers in precursor smectite (Reynolds 1988; Hillier 1993). High Mg influx can be also attained by the fluids common to secondary dolomitization.

Since there is no direct sedimentological evidence of lacustrine-evaporative facies sediments or regional dolomitization in Pirgu rocks, it seems that chlorite-smectite (corrensite) is the result of the alteration of unusually mafic Mg-rich pyroclastics deposited on the seabed during this period. On the other hand, the end of Pirgu time is characterized by rapid seal-level changes, denudation and/or tectonic movements, accompanied by the deposition of probably microbially enhanced carbonate mounds found at that time all over Baltoscandia (the Boda mounds; Hints & Meidla 1997). This would suggest a periodic formation of sabkha-type environments that would have provided the reflux of saline Mg-rich aqueous fluids through subaerally exposed carbonate rocks and bentonite beds.

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Palaeomagnetic investigations of the Early–Middle Ordovician limestones of the St. Petersburg area

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The position of the Baltica continent during the Early Palaeozoic is still poorly known. Determinations of the positions of the Early–Middle Ordovician magnetic poles of Baltica are mostly based on the palaeomagnetic investigations of Swedish Ordovician rocks (e.g. Claesson 1978; Trench & Torsvik 1991; Perroud *et al*. 1992). Recently new palaeomagnetic data on the Lower–Middle Ordovician rocks of the St. Petersburg area were obtained. According to these data, Baltica was located at 50°S latitudes during the Arenig–Llanvirn (Gurevich *et al*. 2003; Smethurst *et al*. 1998) or at 20°S latitudes (Lubnina *et al*. 2003) and at 40°S latitudes during the Llandeilo (Rodionov *et al*. 2002). The position of Baltica during the Late Ordovician has been reconstructed only based on the Scandinavian data (Perroud *et al*. 1992; Torsvik & Trench 1991; Torsvik & Rehnstrom 2003).

New palaeomagnetic investigations of the Ordovician rocks from the Baltic Klint could resolve these contradictions. We sampled the Lower–Middle Ordovician deposits at the Lamashka River and near Shirokovo (western part of the Baltic Klint) and at the Lava, Volkhov and Lynna rivers (central and eastern parts of the Ladoga Klint). This interval corresponds to the Arenig–Llanvirn boundary and includes the Volkhov, Kunda and Aseri regional stages. All measurements were carried out in the palaeomagnetics laboratories of the Institute of Physics of the Earth (Moscow, Russia) and Paris Institute of Physics of the Earth (France).

Two sections were studied in the western part of the St. Petersburg area: the upper part of the Volkhov Regional Stage (grey and green-grey glauconitic wackestones and packstones), and the Kunda and Aseri regional stages (grey packstones with abundant ferrigenous ooids of the Sillaoru Formation, grey wackestones of the Loobu and Napa formations, and white and light grey wackestones). About 150 oriented samples were collected for palaeomagnetic studies. Two components were separated by detailed laboratory measurements at 200° to 580°C. Component A (removed mostly at 400–450°C) exhibited both normal and reversed polarities and is typical of low-magnetic samples (magnetisation <1–2.10–4 A/m). Component B has stable reversed polarity and maximum unblocking temperatures of 500–560°C. It is typical of high-magnetic samples (magnetisation >2–3.10–4 A/m). The mean direction of component B (Dec = 230° Inc = -58° K = 31.8 α95 = 4.5) is close to the direction of Mesozoic magnetisation reversal (Smethurst *et al*. 1998). Sometimes components A and B occur together and component B is less stable.

The mean direction of component A (Dec = 164° Inc = 40° K = 31.8 α95 = 11.3) is close to the Ordovician direction (Smethurst *et al*. 1998) but is characterised by lower inclinations. It means that the Baltica continent was located closer to the equator (at 20°S latitudes) than supposed earlier.

The deposits of the Volkhov and Lower Kunda regional stages were investigated in the Lava River section. The lower part of the section consists of massive, intensively bioturbated parti-coloured (red) packstones, less often wackestones and grainstones ("Red Dikari") with numerous hardground surfaces. Upsection, the content of red-coloured rocks decreases gradually, and the content of clay components and amount and thickness of clay partings ("Grey Dikari" and "Zheltyaki") increase.

High-temperature component A, separated in this part of the section, has the mean direction Dec = 146.2° Inc = 59.5° α95 = 10.3°, which is close to the directions determined in the St. Petersburg area (Smethurst *et al*. 1998). It means that Baltica was located at 40° latitudes of the Southern Hemisphere. It is notable thet all determined directions have only reversed polarity.

The upper part of the Volkhov Regional Stage consists of alternating light-grey, green-grey wackestones, packstones and clays ("Frizy").

High-temperature component B, observed in the most part of this section, has the direction close to the Jurassic remagnetised direction (Smethurst *et al*. 1998). Moreover, in many samples from different parts of the section the unblocking temperature spectra of components A and B overlap and it is not possible to separate them fully.

In the eastern part of the klint (Lynna River) the investigated section is represented by the parti-coloured and green-grey wackestones and packstones ("Frizy") of the lower part of the Volkhov Formation and pink and green-grey wackestones of the Lynna and Obukhovo formations. Detailed laboratory treatments between 200° and 580°C revealed two monopolar components. The first component is related to the lower part of the section. Its mean direction (Dec = 172.6° Inc = 43.9° K = 15.8 α95 = 19.8) is close to the Ordovician direction. The second component, separated mainly in the lower part of the section, has the direction (Dec = 32.6° Inc = 49.2° K = 85.3 α95 = 3.9), which is close to that of the Upper Palaeozoic–Mesozoic remagnetisation (Smethurst *et al*. 1998).

Thus, the palaeomagnetic investigations revealed that Baltica was closer to the equator (20°S latitudes) during the Llanvirn than it was supposed earlier. It is difficult to explain this result by natural understating of an inclination, as this is mainly characteristic of terrigenous rocks.

We cannot state with confidence that the determined components adequately reflect the direction of the Llanvirn geomagnetic field. However, the obtained results can be considered as a new indication of the low-latitude position of Baltica at the end of the Early–Middle Ordovician. The new palaeomagnetic data indicating the location of Baltica at lower latitudes allow us to assume warmer water conditions than supposed earlier. This is confirmed by the following facts: 1) change of terrigenous sedimentation to carbonate one in the Arenig; 2) availability of a high quantity of micritic calcite (Zaitsev *et al*. 2000), which would not have formed in a cold and moderate climate; 3) presence of kaolinite, associated with ferruterous hydromica in insoluble residue of carbonate rocks, which, though in small amounts, are fixed in all sections. Since kaolinite is a mineral which forms only the tropical climate, its occurrence in the terrigenous component suggests at least subtropical climate.

Thus, palaeomagnetic investigations of Lower–Middle Ordovician rocks in different parts of the St. Petersburg area have yielded a pole position at latitude 0.5°S and longitude 45.8°E (dp/dm = 9.6/14.8). Limestone carries another remanent magnetisation component associated with the Late Palaeozoic–Mesozoic remagnetisation event.

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Morphological affinities of orthid larval shells on early development stages

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Three subtypes are distinguished in the recent brachiopod classification: Linguliformea, Craniiformea, and Rhynchonelliformea (*Treatise on Invertebrate Paleontology* 2000). Modern representatives of all three subtypes have larvae substantially differing in their morphology and mode of life. The larva of Linguliformea that hatches from the egg envelope bears all features characteristic of adult animal and has an embryonic shell or protegulum. During the growth of a shell and soft body at the floating stage (approximately one month), the newly formed larval shell grows somewhat obliquely relative to the protegulum. The embryonic shell is separated from the larval one by the discontinuity in shell growth, the formation of which is related to the change in the mode of life. Both shells are organic in composition. After the settlement larva forms an adult mineral shell that is usually separated from the larval one by conspicuous growth rings, or halos. Thus, Linguliformea representatives demonstrate three types of shells corresponding to three development stages: embryonic or protegulum, larval, and adult. The adult shell grows also under the soft larval shell and while the latter retains features of the larva morphology, the mineral shell can become its internal cast. The embryonic and larval shells of modern brachiopods are usually up to 100 and approximately 300–400 μm across, respectively (Williams *et al.* 1997).

Larvae of the Craniiformea and Rhynchonelliformea subtypes are lecithotrophic, have the planktonic existence approximately one day long, and experience metamorphosis after settlement. Although larvae of these subtypes show some substantial differences in their morphology, both of them are characterized by the presence of a gullet in the absence of a mouth and formation of a calcareous shell only after metamorphosis. Thus, recent Craniiformea and Rhynchonelliformea forms have shells of the adult type only.

However, despite the absence even of prerequisites for the formation of a larval shell in ontogenesis of recent craniids, indications of the latter have been discovered in their fossil Palaeozoic and Mesozoic forms (Freeman & Lundelius 1999). This allows us to suggest that representatives of all three brachiopod subtypes had larvae with the planktonic existence of relatively long duration and protected by a shell during at least the Palaeozoic (Freeman & Lundelius 1999), although such observations were made only for craniids and the structure of recent Rhynchonelliformea larvae prevents from the formation of an external larval shell.

The casts of orthid larval shells confirming the possibility of changes in the morphology of Rhynchonelliformea larvae since the Palaeozoic were first found in the Middle Ordovician sediments of the Leningrad and Pskov regions. Juvenile orthid and clitambonitid forms, minimal dimensions of which are 0.5 mm, were obtained from clays of the Volkhov and Kunda stratigraphic units. In some orthid (five pedicle and approximately a hundred brachial valves) and clitambonitid (three pedicle valves) shells, the apical area is located obliquely relative to the remainder of the shell. This area, approximately 400 μm across, bears no growth lines and is separated from the main shell by halo rings. The location, dimensions, and characteristic relief of the area, as well as the presence of halos, allow it to be interpreted as an internal cast of the organic larval shell. The juvenile shell itself is characterized by a fibrous microstructure. The visible length and width of fibres are approximately 50 and 10 μm, respectively. The structure of the larval shell in pedicle and brachial valves of orthid and clitambonitid shells is generally similar, although there are insignificant variations in their dimensions and shapes.

Pedicle shells of orthids bear a smooth slightly elevated near-apex area, approximately 400 μm long and up to 200 μm wide, which is separated from the main shell by one or two halo rings. The apex in some species can be shaped as a thin tube up to 100 μm long and surrounding a pedicle foramen. The insignificant quantity of pedicle valves is most likely explained by the grinding of their back parts related to the growth peculiarities of the curved beak-shaped apex.

The structure of larval shells in brachial valves of all species is similar. In the middle part of the shell, there is a wide low T-shaped ridge, the short segment of which forms an apex. In some species, the anterior edge of the ridge widens, while in others, narrows. The ridge is bordered on both sides by two (*Orthis* sp.) or three (*Nothorthis penetrabilis*) tubercles separated by pits or furrows depending on their development degree. The tubercles can join each other near the anterior edge of the ridge. The apex area can rise above the shell surface as high as 20–50 μm. Halo rings are sometimes missing. The surface of some well-preserved larval shells is entirely or partly covered by small tubercles approximately 5 μm high. Similar tubercles were previously interpreted as negative impressions of a destroyed organic shell with the reticular internal surface (Ushatinskaya 2001). Other authors who observed similar relief in the apical part of juvenile shells of recent terebratulids consider it as an artifact (Stricker & Reed 1985).

Although larval shells probably represent real casts of their soft body, reconstruction of the latter is difficult because of the absence of features in common with larvae of recent brachiopods. It is most similar to the reconstruction proposed for the larva soft body of paterinids (Williams *et al.* 1998). The anterior part of the pedicle valve in the paterinid larval shell bears a notch that is identical to that observed in clitambonitids and brachial shells in both types of larval shells are characterized by similar systems of convexities (Williams *et al.* 1998). Assuming that both groups had in the Early Palaeozoic larvae with the similar morphology, this can serve as an additional argument favouring the monophyletic origin of brachiopods with phosphate and carbonate shells. It should be kept in mind that the only known fossil lingulid embryonic shells also differ from those of their recent forms (Balinski 1997). Such a difference probably implies coenogenetic transformations in the lingulid phylogenesis.

The finds of impressions of orthid and clitambonitid larval shells allow a suggestion that Early Palaeozoic Rhynchonelliformea had feeding larvae that floated, like larvae of recent lingulids, using lophophore cilia. Light one-layer larval shells protected larvae, but were not burdensome in floating. It is probable that pits and furrows at the external surface of the shell correspond to areas of the muscle attachment and tubercles, to the digestive system and paired coeloms. Thus, the muscular system could include both diductors and adductors. Nevertheless, only the metameric structure of larvae can confidently be assumed. The structure of the orthid ventral shell implies the presence of pedicle segment in their larvae. The existence of planktotrophic larvae in Rhynchonelliformea is indirectly confirmed also by the presence of nerve plates and digestive tract in larvae of recent forms. Such structures are unnecessary to the lecithotrophic larva and their presence can most likely be explained by the inheritance of some features from larvae of different type (V.V. Malakhov 2001, personal communication). It is also essential that ontogenesis of recent Linguliformea and Craniformea is characterized by an embryo folded in two with the subsequent formation of a shell, the stage missing in development of Rhynchonelliformea. Based on the position of some structures, Malakhov (1995) assumed that the Rhynchonelliformea ontogenesis “lost” the folding stage and that this phenomenon is responsible for development of the third segment – a mantle in recent larvae, i.e., the development of recent brachiopods shows signs of coenogenetic transformations.

All three subtypes of Early Palaeozoic brachiopods are supposed to have in their ontogenesis the stage of floating larvae, which was subsequently lost in forms with carbonate shells. This is probably explained by the migration of these forms to deeper habitats. The planktotrophic larva is more advantageous in shallow-water environments, from where most brachiopods were embossed by bivalves. In any case, brachiopods with carbonate shells were forced, under unfavourable factors, to develop lecithotrophic larvae of the current type, which process represents embryonic adaptation.

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Recognition of the Mid-Caradoc event in the conodont succession of Estonia

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An episode of biotic, climatic, sea-level and facies changes, called the Mid-Caradoc Event, has been recognised in the upper Caradoc sequence (Keila and Oandu stages) of the Baltoscandian area (Ainsaar 2001, and references therein). A distinct positive $\delta^{13}C$ excursion has been described in the upper Keila Stage and a notable change within various groups of fauna documented in the Keila–Oandu transition interval. The record of a coeval isotopic event in North America suggests that the Mid-Caradoc Event recognized in Baltoscandia is a result of global oceanographic and/or climatic perturbations (Ainsaar 2001, and references therein). However, the conodont data of the Ruhnu (500) core (Männik in press) and earlier studies (Viira & Männik 1999; Männik 2001a) show that the event started already in Haljala time. The pre-event interval, the *A. tvaerensis* Zone (and strata below), contain rich conodont faunas dominated by ramiforms, such as *Baltoniodus* and *Amorphognathus*. The event caused the extinction of several taxa and a considerable decline in the abundance of conodont specimens in samples. Simple-cone taxa started to dominate (Männik 2003). Below, the levels of the most distinct changes in the conodont succession are termed as datums of the event.

Datum 1 lies in the middle of the Haljala Stage and corresponds to the level of the extinction of *A. tvaerensis*. At the same level (or very close to it) morphological changes take place in the *Scabbardella* lineage: the apparatus with high slender equally curved elements is replaced by the one having strongly curved elements with wide bases and narrow cusps. Above this datum, up to Datum 3, *Amorphognathus* is missing in samples or is represented only by few probable, almost unidentifiable fragments. Two samples from the Vasalemma Quarry (both about 15 kg in weight), from the interval between datums 2 and 3 (see below), containing several thousands of conodont specimens, did not yield a single fragment of *Amorphognathus*.

Datum 2 is defined by the extinction of *B. alobatus*, the youngest representative of the genus *Baltoniodus*. Starting from this datum, faunas are dominated by simple-cone taxa such as *Decoriconus*, *Drepanoistodus*, *Panderodus*, etc. Just above Datum 2, the first abundant occurrence of *P.* ex gr. *equicostatus*, prevalent in the event intervals in the Silurian (Jeppsson 1998), was noticed. In the samples from the Vasalemma Quarry (see above) the faunas are strongly dominated (more than 50%) by *Semiacontiodus*. Datum 2 lies also in the middle part of the Haljala Stage, about 1 m higher than Datum 1 in the Ruhnu (500) and Valga (10) cores (Männik 2001, 2003).

Datum 3 corresponds to the level in the lowermost Oandu Stage exhibiting the first signs of the recovery of the event-affected conodont faunas. At this level the *Amorphognathus* lineage, now represented by *A. ventilatus*, returns.

Datum 4 corresponds to the level of the extinction of the genus *Semiacontiodus*. In the Ruhnu (500) core, Datum 4 lies in the lowermost part of the Oandu Stage, about 1 m higher than Datum 3 (Männik 2003). In the Taga-Roostoja (25A) core, the uppermost specimens of *Semiacontiodus* were found just below the discontinuity surface at the boundary between the Keila and Oandu(?) stages (Viira & Männik 1999), indicating that some strata are missing in that section.

Samples from the interval between datums 2 and 4 showed great variations in the abundance of specimens of several simple-cone lineages (e.g. *Decoriconus*, *Panderodus*, *Semiacontiodus*). Further studies are needed to find out if there exists any pattern in these variations.

The thickness of the event interval is about 10 m in the Ruhnu (500) core and probably more than 20 m in the Taga-Roostoja (25A) core (Männik in press; Viira & Männik 1999). In the Valga (10) core only the lower part of the event has been studied and datums 1 and 2 have been recognized (Männik 2001).

It has not been checked yet if the succession of changes in conodont faunas, recognized in the Mid-Caradoc Event interval in Estonia, can be followed also in other regions of the world, but, based on isotope studies (Ainsaar 2001, and references therein), it seems most probable. Anyhow, at least in Estonia the datums of the Mid-Caradoc Event provide an excellent tool for detailed correlation of sections (Männik 2003). The sequence of environmental (sedimentological) changes described in the studied region (Nestor & Einasto 1997; Ainsaar 2001) indicates that the Mid-Caradoc Event is evidently a P–S event *sensu* Jeppsson (1990). Also, it is most probable, that further studies may reveal more details about the effect of the event not only to the conodonts but also on other faunas.

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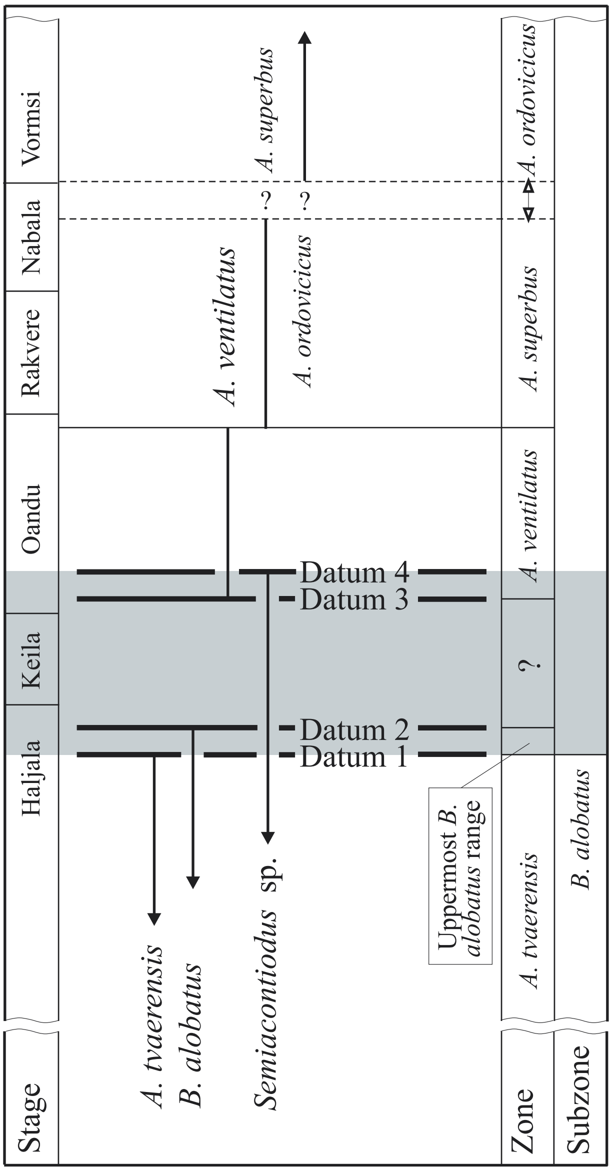

Fig. 1. Distribution of selected conodont taxa in the Mid-Caradoc Event interval. Grey area – interval showing the most distinct changes in the conodont succession. *A. – Amorphognathus, B. – Baltoniodus*.

Ordovician–Silurian boundary in the Subpolar Urals, some new developments

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The Ordovician–Silurian boundary in the Timan–northern Ural region has been under discussion for a long time (e.g. Beznosova & Männik 2002). In the latest official stratigraphical scheme for the Urals (Antsygin et al. 1993) it was drawn between the Yaptikshor and Dzhagal formations. The possible boundary level is marked by lithological changes but is not proved biostratigraphically due to the lack of well-preserved identifiable fossils. For example, in the Kozhym-108 section, Subpolar Urals, the boundary was tentatively drawn well below the upper boundary of the Yaptikshor Formation (in the lowermost part of bed 20), just above the level of the uppermost identified *Proconchidium muensteri* (St. Joseph) (Beznosova 2000). No identifiable macrofossils were found in the strata above this level.

Carbon isotope studies of the Kozhym-108 section revealed a $\delta^{13}C$ curve almost identical with that from the Dob's Linn (Scotland) sections, although the Ordovician–Silurian transition interval in the Ural region is represented by various shallow-shelf carbonates (Beznosova & Männik 2002). It became evident that the upper part of the Yaptikshor Formation corresponds to the Hirnantian, and the Ordovician–Silurian boundary should be looked for higher in the sequence, in the lower part of the Dzhagal Formation (Beznosova et al. 2002). The configuration of the $\delta^{13}C$ curve in the Kozhym-108 section also suggests that no major gap occurs at the Ordovician–Silurian boundary in the Subpolar Urals.

In 2003, another section (Kozhym-116), about 5–6 km NNW of Kozhym-108, was studied. The lithological sequence and the $\delta^{13}C$ curve of the Kozhym-116 section fit well with the earlier data from Kozhym-108, and allow detailed correlation of these sections. Also in the Kozhym-116 section all macrofossils collected are too strongly recrystallized to be properly identified. Here an interval (bed 36, about 7 m in thickness) rich in pentamerid brachiopods was recorded about 100 m above the contact between the Yaptikshor and Dzhagal formations. The study of some better preserved specimens shows that they possibly belong to the Ordovician genus *Proconchidium* and, accordingly, this level still lies below the Ordovician–Silurian boundary. However, just recently conodonts were found in one of the samples collected from the Kozhym-108 section (processing of samples from this section and from Kozhym-116 is still in progress). The sample (C 01-49) comes from the lower part of the Dzhagal Formation (about 25 m above its lower boundary; Fig. 1). So far, 25 specimens have been found, 18 of which belong to *Panderodus* sp. The occurrence of typical Silurian taxa *Oulodus*? aff. *nathani* (2), *Walliserodus* cf. *curvatus* (Branson et Branson) (2) and a probable fragment of *Ozarkodina* sp. in this sample indicates that this level is already of Silurian (Rhuddanian) age. However, the sample comes from a level below the Ordovician–Silurian boundary interval proposed on the basis of comparison of the $\delta^{13}C$ curves (Beznosova & Männik 2002; Beznosova et al. 2002; Fig. 1), and evidently far below the level with probable *Proconchidium* in the Kozhym-116 section (Fig. 1). There seem to be two explanations for this problem:

1. In a continuous section across the Ordovician–Silurian boundary the Silurian-type conodont faunas appear already in the uppermost Ordovician strata (e.g. Armstrong 1995). If this is the case, then it is another proof of continuous sedimentation in the Ordovician–Silurian transition in the studied region and the system boundary should be searched for higher in the sequence.
2. The probable Ordovician–Silurian boundary interval defined by the comparison of the $\delta^{13}C$ curves is not correct and the boundary in reality lies below sample C 01-49.

But, if the pentamerids in bed 36 (Kozhym-116) really turn out to be *Proconchidium*, the above conclusions will need a complete revision. As demonstrated above, the data available at the moment seem to be contradictory and further detailed sedimentological, geochemical and biostratigraphical studies are necessary to solve the problem.

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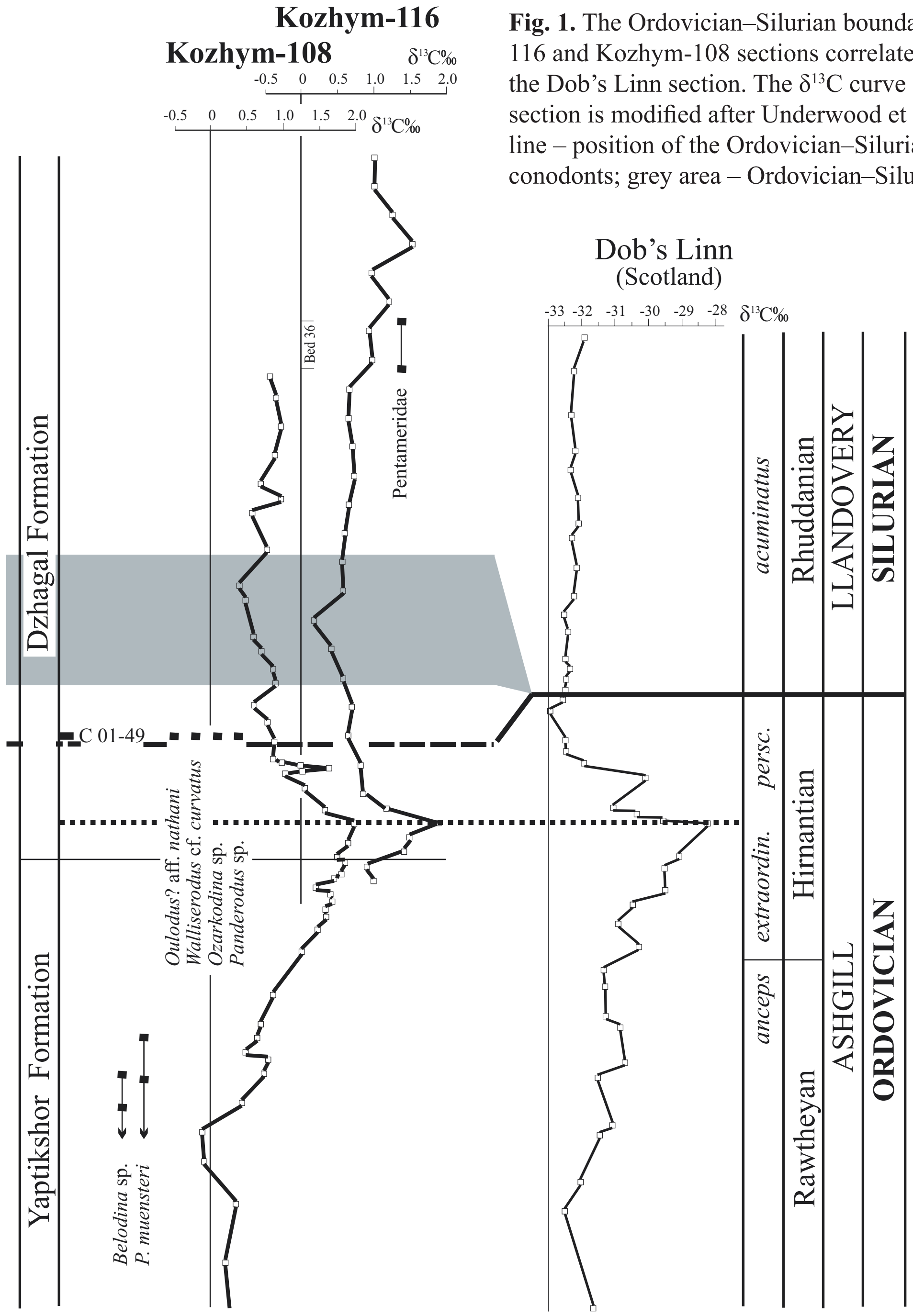

Fig. 1. The Ordovician–Silurian boundary beds in the Kozhym-116 and Kozhym-108 sections correlated by isotope data with the Dob's Linn section. The $\delta^{13}C$ curve for the Dob's Linn section is modified after Underwood et al. (1997). Thick dashed line – position of the Ordovician–Silurian boundary based on conodonts; grey area – Ordovician–Silurian boundary interval based on the comparison of the $\delta^{13}C$ curves after Beznosova & Männik (2002). *P.=Proconchidium; extraordin.= extraordinarius; persc.=persculptus.*

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Middle–Upper Ordovician carbon isotopic record from Västergötland (Sweden) and East Baltic

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A number of positive shifts in stable carbon isotopic composition have been documented from the Upper Ordovician of the Baltoscandian Palaeobasin (Kaljo *et al.* 1999; Ainsaar *et al.* 2001). Among them the glaciation-induced end-Ordovician isotopic excursion is the most prominent. The reason for other second-order isotopic changes is not fully understood yet. Here a new distinct shift recorded from the Middle Ordovician of Sweden and the East Baltic is considered. A total of 139 samples of carbonates and rocks of mixed composition were analysed for the stable carbon and oxygen isotopic composition from the Gullhögen Quarry section (Västergötland, Sweden). The contemporaneous interval in the Jurmala core section of Latvia is used as a reference section from the Central Baltoscandian Confacies Belt. The oxygen data are not discussed herein.

The stable carbon isotopic curve at Gullhögen (Fig. 1) is well resolved in the interval from the base of the Lanna Formation (Fm) up to the lower part of the Dalby Fm, i.e. from the lower Middle Ordovician up to the basal Upper Ordovician. Three samples per metre on average guarantee a good reliability of more important trends. Occasional scattered isotopic values in the interval from the middle Dalby Fm up to the Jonstorp Fm cannot be considered reliable. In particular, the Dalby Fm was macroscopically characterised as limestones, but only occasionally delivered an isotopic signal.

The Lanna–Täljsten stratigraphic interval represents the rising limb of the carbon isotopic curve at Gullhögen, with the $\delta^{13}C$ values increasing from -1 to 0.5‰. The Täljsten–lower Dalby interval could generally be interpreted as "the plateau" of the curve at $\delta^{13}C$ values of 0.5‰. Two features must be emphasised in this particular interval: (1) a well-documented positive shift in $\delta^{13}C$ values in the middle–upper parts of the Holen Fm, with a gradual rise from 0.5 to 1.5‰ and a sharp decline at the Holen–Våmb boundary, and (2) a slow rise in $\delta^{13}C$ values in the basal Ryd Fm, with the subsequent decline in the upper Ryd–lower Dalby interval.

The carbon isotopic curves from Gullhögen and Jurmala show resemblance in some important details. The positive shift in the upper Holen Fm is correlated with a similar excursion (+1.5‰) in the Segerstad Fm. This is also interpreted as evidence for similar age of the Segerstad Fm in the East Baltic area (where the formation is traditionally considered as an equivalent of the Aseri Stage – see Männil & Meidla 1994) and the upper part of the Holen Fm in Sweden (up to now considered as an equivalent of the Kunda Stage; Jaanusson 1982; Nõlvak & Grahn 1993). This particular shift, termed here as the Middle Darriwil carbon isotopic excursion, is the oldest documented variation of its kind in the Ordovician of Baltoscandia. The Middle Darriwil carbon isotopic excursion is not well reflected in northern Estonia (see Ainsaar *et al.* in this volume). The lowering trend of the carbon isotopic values in the upper part of the Taurupe Fm and in the Dreimani Fm of the Jurmala core section agrees with the changes in the upper Ryd Fm and lower Dalby Fm at Gullhögen, and with the decrease in carbon isotopic values in the coeval beds of northern Estonia (see Ainsaar *et al.* in this volume).

The above data suggest that the Middle Ordovician correlations in the Baltoscandian area may need to be reconsidered. If the second-order carbon isotopic fluctuations are interpreted exclusively as evidence for the onset of glaciation in high latitudes, we have to consider that the cooling started already in the early Middle Ordovician but this seems unlikely.

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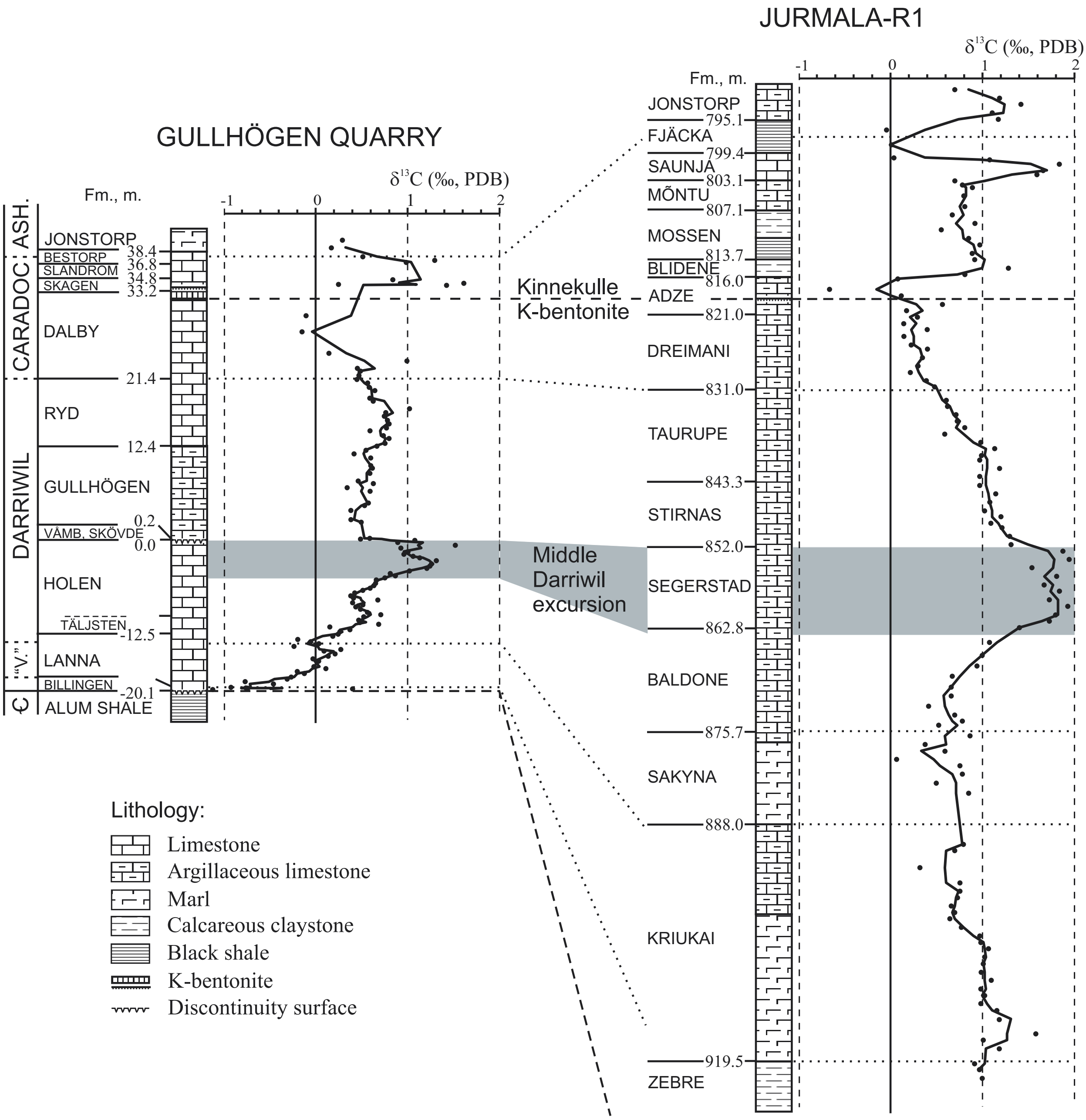

Fig. 1. Stable carbon isotopic curves from the Gullhögen Quarry section (Sweden) and Jurmala R1 core (Latvia). Raw data (solid circles) have been filtered with a three-point moving average (solid line).

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Ordovician sea-level changes in Baltoscandia

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Baltoscandia was positioned at about 60ºS latitude in the Early Ordovician, but gradually moved northwards and entered the subtropical realm in the Late Ordovician. This caused changes in the depositional environment, and four depositional phases are distinguished:

(1) Extremely sediment-starved clastic deposition and widespread dysoxia (Late Cambrian to early late Tremadoc)
(2) Very slow deposition of cool-water limestones (often glauconitic) on the midshelf with extremely condensed glauconite-rich clastics on the inner shelf (late Tremadoc-mid Arenig)
(3) Slow deposition of cool-water limestones largely without glauconite throughout the inner and midshelf (Llanvirn-early Caradoc; transitional interval with deposition of shallow-water glauconitic lim estones during the late Arenig)
(4) Slow deposition of cool-water limestones in deeper-water settings associated with more rapidly accumulating local carbonate build-ups and warm-water limestones in shallower water (mid Caradoc-Ashgill)

Lithofacies models have been established for each of these depotional phases. Overall the craton was extensively flooded and the clastic supply to the epicontinental sea was therefore very limited. The Baltoscandian Ordovician successions are therefore extremely condensed. Along the western [present day direction] periphery the clastic influx increased from the Llanvirn onwards as a result of the ongoing closure of the Iapetus Ocean with development of a foreland basin adjacent to the rising Caledonide mountain belt. As a result of the slow deposition in combination with almost quiescent tectonic conditions the sea level record essentially reflects eustasy, making the area ideal for the analysis of sea level changes.

In connection with the IGCP 410 project the present author worked out an integrated sea level curve for the Ordovician of Baltoscandia (Nielsen in press, 2003). The reconstruction is based on an analysis of the successions in the Oslo Region, Scania, Siljan, southern and northern Estonia, supplemented by data from Västergötland, Öland and Jämtland. It was aimed at to analyse a depth transect across the Baltoscandian shelf, anchored in the outer shelf successions preserved in Oslo and Scania. This approach ensures easy identification of gaps relating to lowstands in the nearshore locations and at the same time the relatively nearshore sections provide a better resolution for the highstand intervals, represented by relatively uniform mudstone successions on the outer shelf.

The preserved Ordovician successions essentially comprise vertically stacked, non-prograding thin strata, and conventional sequence stratigraphy therefore does not apply. Omission surfaces (most of which are inferred to be submarine) are numerous particularly shorewards, and identifying the sequence stratigraphical importance of individual surfaces is virtually impossible within a given area. Erosion associated with even major stratigraphic gaps is in most cases surprisingly limited and revealed mainly by comparison with more complete successions elsewhere. Channelling is reported only from the highest Caradoc-Ashgill, indicating that the area was submerged throughout the Ordovician. Hence the sea-level changes were mainly deduced from lithofacies distribution whereas the recognition of unconformities was given less weight.

The Baltoscandian sea-level curve, which probably represents 3rd order oscillations, has been generated in three steps: 1) An initial interpretation was based on lithofacies analysis of the Ordovician succession in the Oslo-Asker district of southern Norway, which is dominated by shales and mudstones interbedded with nodular limestone units (see Owen *et al*. 1990 for review). 2) The resulting curve was then compared with data from the mainly calcareous deposits described from Sweden, notably the fairly complete succession preserved in the Siljan area (see Jaanusson 1982 for review). The Jämtland, Västergötland, Scanian and Öland successions have also to some extent been included, of which Scania and Jämtland represent the deepest and Öland the shallowest palaeo-depth on the shelf profile. The correlation with Swedish sections permitted a first check on the sequence of sea-level changes, but also provided a more detailed scaling of the highstand intervals, represented by relatively uniform mudstone units in the Oslo region. 3) The adjusted curve was finally compared with the successions described from Estonia (also dominated by limestones, but often more argillaceous), where I focussed on a transect from southern to northern Estonia (for reviews of lithostratigraphy, see Männil & Meidla 1994 and Raukas & Teedumäe 1997). The successions of northwestern Estonia represent the most nearshore Middle-Upper Ordovician deposits preserved in Baltoscandia, whereas the deposits of southern Estonia and bordering Latvia represent deeper-water facies similar to those in central Sweden. This third step of pattern recognition again ensured that the succession of events was verified, and further details of the sea-level highstand intervals could be added and calibrated according to lithofacies.

The correlation of sea level changes has a vast but as yet essentially untapped potential for improving the correlation of Ordovician successions. An insight into sea level changes also provides an improved understanding of the traditional bio- and lithostratigraphic correlations and some pitfalls may be avoided. It is in this context stressed that highstand events are easier to correlate in terms of biostratigraphy, because the relatively more widely distributed deep-water faunas (graptolites/conodonts) spread across the shelves. A quick glance at the sea level chart also indicates a clear tendency for adaptive radiations among graptolites to match drowning events; this trend may also be valid for other groups. Lowstands leave conspicuous marks in the shelfal sedimentary record, and were originally the primary tool for correlation of depositional sequences (Vail *et al*. 1977). However, shallow water faunas tend to be endemic and are therefore notoriously difficult to correlate. Despite the fundamentally different correlation potential, no philosophy about sea level changes has been formulated regarding intercontinental chronostratigraphic correlation of the Ordovician. In this perspective it is a matter of concern that the base of the Middle Ordovician has been suggested at the base of an extended, composite lowstand, wherefore biostratigraphic correlation might be anticipated to prove difficult. To some extent the same may be said for the base of the Upper Ordovician, which, however, in detail seems to tie in with the terminal part of a highstand interval and thus may be anticipated to have a better potential for widespread bio-based correlation.

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Reinvestigation of the Volkhov and lower Kunda in Estonia and western Russia: preliminary biostratigraphic results

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During field campaigns in 1994, 1999, 2001 and 2003 a number of sections through the Volkhov Stage, including bounding strata below and above, have been sampled for macrofossils with the primary aim of collecting material for a revision of the trilobite fauna. The investigated Estonian sections include (from west to east) the Osmussaar Island, Pakri Peninsula, Harku near Tallinn, Nõmmeveski Canyon and Saka. In western Russia the Putilovo quarry and the Lynna River section have been sampled. The registration of the collected material is still ongoing, but it contains in the size order of 15.000 trilobite specimens of which the exact sampling level is documented. By comparison with Russia, the Estonian sections are condensed and significantly less trilobitic, and there is little doubt that the Putilovo section eventually will stand out as the reference section for the Volkhov of the eastern Baltic area.

The revision of the large material has barely begun, hence the following observations must be taken as preliminary.

The Billingenian Päite Member contains abundant *Paramegistaspis estonica*, ranging all the way to the Volkhov unconformity. Otherwise rather few trilobites have been encountered at this level. Representatives of *Megistaspis* turn up at the very base of the Volkhov and the Billingen/Volkhov transition is thus a well-defined boundary.

The hard and often dolomitic Saka Member is generally difficult to extract trilobites from, and it has been sampled only at Harku, Saka and Putilovo. The sporadic presence of *Globampyx linnarssoni* indicates a correlation with the *M. polyphemus* Zone. Judging from older literature *M. polyphemus* itself also occurs, but the newly sampled megistaspid material has not been revised as yet and such occurrence cannot be confirmed as yet. At any rate the megistaspid material appears to be dominated by a *limbata*-like form. *Rhinoferus* turns up in the upper part of the Saka Mb.

It is suspected that also the lower part of the Zheltyaki Mb. belongs to the *M. polyphemus* Zone, but this unexpected finding needs further study; the suspicion is based on the distribution of *Globampyx linnarssoni* and *M. polyphemus*-like pygidia (as yet unstudied). Lamansky (1905) suggested that *Asaphus broeggeri* is characteristic of the middle zone of what is now called the Volkhov Stage, but this species extends far down the previous zone and cannot serve as index fossil for $B_{II}\beta$. This zone contains sparsely occurring *Niobe (Niobella) lindstroemi* and *Nileus orbiculatoides*, providing ties with the *M. simon* Zone of Scandinavia.

Also the definition of the $B_{II}\gamma$ zone is problematic. In the Lynna section, situated in the Volkhov area, Russia, *Asaphus lepidurus* and "*Megistaspis" gibba* appear at the base of the Frizy Mb., i.e. like originally described by Lamansky (1905). However, in the Putilovo section there seems to be an interval low in so-called $B_{II}\gamma$, which has no equivalent in the Lynna section, and so far no undoubted *A. lepidurus* has been recorded within this interval. If *A. lepidurus* eventually proves present it is at least much more rare than at higher levels. In Estonia the *A. lepidurus* Zone is rather thin and disappears west of Harku, probably due to erosional truncation during the Kundan. It is possible to make a bed-by-bed correlation of the Telinõmme Mb. from Harku to the Pakri Peninsula, indicating that the westwards thinning of the Volkhov primarily is due to subsequent erosion. It is also evident that the strata assigned to the Lahepera Mb. in the Paldiski area are not equivalent to the Lahepera Mb. of the Tallinn district; the former should be included in the Telinõmme Mb.

The boundary between the Volkhov and Kunda stages is well-defined and within Russia the previously indicated position in the succession is verified. The *A. expansus* Zone has generally been assumed absent in Estonia, but the Pada Mb. represents this level, containing *A. expansus* itself in association with several other characteristic *expansus* Zone trilobites, e.g. *Illaenus sarsi*. A thin *A. expansus* Zone has been found in all sections east of Tallinn, whereas it is absent to the west.

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New data on the Komstad Limestone of SE Scania, Sweden: Volkhov–Kunda bio- and ecostratigraphy

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The first results of a detailed bio- and ecostratigraphic study on the Komstad Limestone of southern Scandinavia were published by Nielsen (1995). The Komstad Limestone, which in age spans the Volkhov and Kunda stages, is more than 16 m thick in SE Scania, and the original sampling density – although done bed-by-bed – was in some intervals insufficient for detailed analysis of fossil assemblages. In order to improve the resolution of ecostratigraphy new sampling of the *M. limbata* Zone (upper Volkhov) and the *A. raniceps* Zone (middle Kunda) was undertaken in the Killeröd quarry and the Killeröd Canal, respectively (the position of the mentioned localities can be seen in Nielsen 1995). A new outcrop of the lower part of the Komstad Limestone including the *M. polyphemus* and *M. simon* zones (middle Volkhov) has been investigated in the rivulet at Flagabro, located only about 1 km west of the Killeröd Quarry. A rich fossil material has been recovered from the main lower part of the *M. simon* Zone not accessible in any quarries (cf. Nielsen 1995). The new data confirms the previous biostratigraphic results and also verify the previous thickness estimates on the lower part of the Komstad Limestone (Nielsen 1995). At the same time the new material provides basis for a much more detailed palaeoecological analysis; these new results will be presented at the WOGOGOB meeting. The supplementary sampling is part of the effort to unravel the mismatch in interpretation of the middle Volkhov sea-level published by Dronov & Holmer (2002) and Nielsen (1992, 1995, 2003).

In order to get a more precise measurement of the thickness of the Komstad Limestone a cored drilling was made near the Killeröd section C sensu Nielsen (1995), where the “Upper Didymograptus Shale” and the Killeröd Formation overlies the Komstad Limestone. However, even though the drilling was spudded only 30 m from this section, Komstad Limestone was encountered directly beneath a thin Quaternary cover. It was not possible to move the sizeable drill rig closer to the outcrop. However, although the Komstad Limestone section thus was not complete, the drilling penetrated 16.25 m before entering the Tøyen Shale and the drill core therefore currently represents the greatest thickness of the Komstad Limestone documented in Scania. Total thickness of the formation is suspected to be in the size order of 18 m. It ranges from the uppermost part of the *M. polyphemus* Zone and into the *A. raniceps* Zone, but it cannot be excluded that the upper part even reaches into the *M. obtusicauda/gigas* Zone of the upper Kundan.

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Late Ordovician sea level changes in the Oslo area and Estonia: tie-points and problems

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An integrated sea-level curve for the Ordovician of Baltoscandia has recently been constructed by Nielsen (2003a, in press); named events are according to these publications. However, the suggested correlation of some units ranging from the Oandu to the Porkuni is controversial, notably regarding Pirgu units. In order to unravel these discrepansies and arrive at a better understanding of the depositional system, the authors of this abstract launched a mutual project on Late Ordovician sea levels and as a starting point we charted the disagreements.

It is uncertain exactly how to tie the base of the Upper Ordovician into the Estonian succession, but it matches a level around or just below the base of the Kukruse Stage. The sea level was low during the early Caradoc (Vollen Lowstand), matching the Kukruse Stage. The Haljala/Idavere boundary marks a drowning, suggested to match the Arnestad Drowning, and the Keila Stage broadly speaking seems to mark another drowning phase that took place just after deposition of the Kinnekulle Bentonite Bed (Keila Highstand). Then followed the conspicuous Frognarkilen Lowstand. The present authors have some disagreement about how to tie in particular the later part of this lowstand into the Estonian succession. We agree that the gradual shallowing commenced during the late Keila, and it is also obvious to suggest a match of peak lowstand with the unconformity below the Oandu. ATN suggests that the Blidene Fm of southern Estonia reflects incresed clastic influx during peak lowstand. In the interpretation of ATN the ensuing drowning started slowly (=Oandu Stage), still matching the upper part of the Frognarkilen Fm of Oslo and probably the terminal part of the *foliaceus* graptolite Zone in Scania. The sea level rise strongly accelerated at the base of the *clingani* Zone (Nakkholmen Drowning), which ATN equates with the shift to the Nakkholmen Fm of Oslo, the Mossen Fm of southern Estonia and the Rägavere Fm (Tõrremägi Mb) of northern Estonia.

TM indicates that the highest concentration of clastics (= peak lowstand) is recorded from the basal parts of the Mossen Fm (bituminous shales) and Variku Fm (calcareous siltstones), both of latest Keila age. In the East Baltic, the drowning is dated as late Keila and this seems to be in good agreement with the proposed age of the Frognerkilen Fm. (cf. Owen *et al.* 1990), which according to ostracod evidence can be dated Haljala/early Keila (Qvale 1980). TM suggests a gradual sea-level rise from the late Keila onwards, re-establishing deposition in northern Estonia since Oandu and with a gradual decrease of siliciclastic supply through the Oandu and Rakvere stages (Ainsaar & Meidla 2001).

A new shallowing, the Solvang Lowstand, took place in the upper part of the *clingani* Zone. This shallowing was less conspicuous than the previous Frognarkilen Lowstand. ATN broadly equates it with the unconformity between the Rakvere/Nabala stages, although in detail it may actually be a late intra-Rakvere event. The subsequent drowning commenced in the late *clingani* Zone, but accelerated strongly into the *linearis* Zone. This is one of the most prominent drownings (Linearis-1) of the Ordovician and it was associated with non-deposition in the central Oslo area. The same may be the case in southern Estonia, where the condensed Fjäcka Shale is suggested to reflect highstand conditions during the *linearis* Zone (upper part according to chitinozoan dating). The correlation into northern Estonia of the smaller variations in sea level during the *linearis* Zone, revealed primarily by Swedish sections, is not straightforward.

According to the interpretation by TM, the prominent Solvang Lowstand probably correlates with the late Nabala Stage as the most prominent lowstand in the area, although the correlations across the facies belts in the eastern Baltic area may need to be justified in the Rakvere-Nabala interval. This is not necessarily in disagreement with the interpretation by ATN, as the Solvang Fm according to some authors seems to be stratigraphically incomplete in Oslo-Asker and its top may at some localities reach into the *linearis* Zone (the former Høgberg Mb by Owen, see Bruton & Owen 1979 and Owen *et al.* 1990).

According to the interpretation by ATN, the composite linearis highstand was terminated by a significant shallowing, the Grimsøya Lowstand. This event was responsible for the change to deposition of the Jonstorp Fm in Sweden and southern Estonia; an unconformity is inferred below the Moe Fm of northern Estonia by ATN. The Grimsøya Lowstand was followed by a renewed as yet unnamed drowning matching the Skjerholmen Fm of Oslo (new data, not in Nielsen & Harper 2003); the sea level did, however, not reach as high as during the *linearis* Zone. For lithological reasons the Moe Fm of northern Estonia has been interpreted as a highstand deposit of ATN, who initially suggested that it thus could not match the 'dirty' Jonstorp Fm of southern Estonia. However, according to TM and other Estonian colleagues this particular interpretation is in fundamental disagreement with the biostratigraphical data, which suggest that the Moe Fm is a rough equivalent of the Jonstorp Fm (Meidla 1996) and both units overlie the chitinozoan *A. barbata* subzone (Nõlvak 1980). Hence, according to TM the possible Grimsøya Lowstand looks controversial in the eastern Baltic succession, and most probably the post-*linearis* shallowing was rather gradual and culminated only in latest Pirgu.

In light of the biostratigraphic evidence, ATN currently considers whether the Moe Fm alternatively may reflect the unnamed Skjerholmen and subsequent Spannslokket drownings. If so, it is unlikely that the clastics of the Jonstorp Fm derives from the Finnish land area. ATN suggests that the Ärina Fm reflects the Husbergøya Fm of Oslo (late Rawtheyan) and that the Porkuni succession reflects only the initial small-scale Langøyene Drowning (Hirnantian); the main upper part of the Hirnantian interval is represented by an omission surface in Estonia during which the area is likely to have emerged above sea level (later so in southern than in northern Estonia).

It is obvious that a biostratigraphical framework is essential for constraining the correlation of sea level cycles. The boundaries of several graptolite zones are rather arbitrary in both areas under consideration (see Männil 1976; Owen *et al.* 1990) and this could be one reason of the different interpretations. As apparent the disagreements described above are rather major at some levels, but we believe that they can be solved partly or entirely by a thorough revision of the existing biostratigraphical and lithological evidence, perhaps supplemented by new data from particularly critical intervals.

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A revised chronostratigraphic scheme for the Ordovician of Baltoscandia: potential for trans Baltoscandian correlation

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The detailed chronostratigraphic framework initially established for the Baltic States and western Russia provides a template for the correlation of biotic and geological events at high levels of precision during the Ordovician Period. These chronostratigraphic divisions were based on biotic data from condensed carbonate successions in a relatively shallow-water part of the Baltoscandian carbonate-ramp. Despite the many and obvious advantages of the scheme, it is only safely applicable within the type East Baltic area. In particular the correlation of upper Middle and Upper Ordovician outside the East Baltic is difficult and far from secure. Currently correlation within the Scandinavian successions is often achieved without reference to the East Baltic stadial units, whereas adjacent to the Caledonian mountain belt in Norway, authors have commonly adopted the British Ordovician chronostratigraphy. On an international scale the East Baltic units are of relatively short duration making long-range correlation outwith the Baltic palaeobasin difficult. For these reasons we suggest a rationalisation of parts of this chronostratigraphy with a view to developing the scheme as a standard for the whole of Baltoscandia and enhancing its usefulness outside the Baltic area. A possibility is to reduce the rank of some of the East Baltic stages to substages and focus on the definition of 8 to 10 stages that are more readily correlated across Baltoscandia. Clearly there are a number of key events - many triggered by sea level changes - that can be recognised across the entire Baltoscandian region, which can form the basis for a more generalised and applicable chronostratigraphy. If possible, these stadial boundaries should preferably be defined to match international boundaries, although this should not be a primary criterion.

The basal Ordovician Pakerort and Varangu stages can also be recognised outside the East Baltic area, but the latter interval is of rather short duration, and we suggest a combination of these units into one stage, for which the Estonian name Iru is available. The current Hunneberg, Billingen, Volkhov and Kunda stages are recognisable on the basis of a variety of biotas, both regionally and internationally and we support the continued use of these units. The Hunneberg and Billingen stages are occasionally ranked as substages of a Latorp Stage, particularly by eastern Baltic authors, but from a Scandinavian point of view it is useful to distinguish them as separate stages. The post-Kunda divisions are less easy to correlate outside the type area. Above the Kundan, there is no readily recognised biotic event below the base Idavere that affects both the shelly and pelagic faunas. The base of the Upper Ordovician is not clearly recognisable in the shelly succession, but probably correlates with a level at or just below the base of the Kukruse Stage. It is paradoxical that the GSSP for the Upper Ordovician is located in Baltoscandia but cannot be readily correlated into the regional chronostratigraphical classifications. Currently, however, correlations between the graptolite and shelly biofacies at this level remain tentative; for this reason the position of the GSSP is largely irrelevant to the development of this chronostratigraphic scheme. The case is similar to the base of the Darriwilian, which is located inside the Volkhov Stage and thus is not reflected in the Baltoscandian chronostratigraphic subdivision.

Above the base of Idavere (= base Haljala in some schemes), the base of the Oandu Stage is characterised by a conspicuous faunal turnover in the East Baltic area, and although the faunal changes are less striking in Scandinavia, the level is still potentially traceable. It correlates with the base of the *clingani* graptolite Zone, which, according to recent suggestions, may define the base of the second highest global stage in the Ordovician System. Hence this level appears an obvious candidate for a Baltoscandian stage. Upwards, the base of the Harju Series at the Rakvere/Nabala boundary represents another level that can be widely traced in Baltoscandia, followed by the Vormsi/Pirgu boundary.

We thus suggest two possible solutions:
(1) Three post-Kunda stages, comprising the combined Aseri–Keila, Oandu–Vormsi and Pirgu–Porkuni stages of current usage, or
(2) Five post-Kunda stages, comprising the current Aseri–Kukruse, Idavere–Keila, Oandu–Rakvere, Nabala–Vormsi [=Vasegaardian] and Pirgu–Porkuni stages.

The first solution creates a total of 8 Baltoscandian stages, the latter case 10. An alternative is to preserve the Porkuni Stage, since it probably correlates with the Hirnantian. Although relatively short, this latter interval has recently been nominated as the highest global chronostratigraphic unit of the Ordovician.

We have deliberately avoided naming any new units in order to encourage discussion of this controversial proposal. However, rationalisations are not uncommon to effect more long-range correlations. Recently the traditional stages of the British Caradoc Series were reduced from 8 to 4 (even with the addition of the Velfryan) in an attempt to make these chronostragraphic divisions more applicable outside the Anglo-Welsh borderlands.

Distribution of *Nemagraptus* in the East Baltic Ordovician

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Recently the appearance level (FAD) of *Nemagraptus gracilis* was considered as the lower boundary of the Upper Ordovician for the global scale. The type section for the "Golden spike" is the Fågelsång section in Scania, Sweden, from where detailed data on the distribution of graptolites, conodonts and chitinozoans in the black shale succession have been obtained (Finney & Bergström 1986, and references therein; Bergström *et al.* 2000).

The occurrence of *Nemagraptus* in the carbonate rocks of the East Baltic area is not as yet illustrated. Earlier genus level determinations from some Estonian and Latvian sections were given by Männil (1976, figs. 2–4; 1986, fig. 2.1.1.). As a rule, extremely rare specimens occur in beds correlated traditionally with the interval from the *Hustedograptus teretiusculus* up to the lowermost *Diplograptus foliaceus (=multidens)* graptolite zones or with the upper part of the *Pygodus serra* and the lower part of the *Amorphognathus tvaerensis* conodont zones, which correspond roughly to the upper Uhaku and Kukruse regional stages. The boundary between these stages coincides roughly with the upper boundary of the global Darriwilian Stage (see Nõlvak 1997, 1999).

So far *Nemagraptus* has been found in the limestones (mainly calcarenites) of the East Baltic area at three different stratigraphical levels. The lowermost finds of our collection are from the uppermost beds with *Gymnograptus linnarssoni* and just above them (Fig. 1). The latter graptoloid species has an important correlative value. Its range is restricted to the lower Uhaku succession in all sections investigated in the East Baltic and in the Siljan district, Sweden (Männil 1976, fig. 2). All nemagraptids belong to the *Nemagraptus subtilis* Hadding. Its range can be well followed also in the Fågelsång section, where it disappears below the appearance of *N. gracilis* (Bergström *et al.* 2000). Except one problematic find (No. 22, Fig. 1) all distinguished specimens from Leningrad District (Russia), northern Estonia, western Latvia and Siljan District (Central Sweden) are concentrated to the lower part of the Uhaku Stage representing the oldest known level of nemagraptids.

The second level with nemagraptids can be followed only in the Savala section, in the boundary beds of the Uhaku and Kukruse stages, well known by their content of marine oil shales – kukersites in beds, which dip to the south (see Männil 1986). However, the material available is too fragmentary and more precise identifications on the species level are at the moment not possible.

The third and the most remarkable occurrence of nemagraptids is in the beds of the upper Kukruse Stage, which are absent (wedge out) in the outcrop area in northern Estonia (Fig. 1). In these beds *N. gracilis* (Hall) has been identified in several sections of central and southern Estonia and western Latvia. A relatively rich graptolite fauna will be described elsewhere. However, despite detailed sampling this species was not found from the coeval Dalby Limestone beds in the Fjäcka section (Siljan District, Sweden; Nõlvak *et al.* 1999) – the stratotype section for the conodont zonation (Bergström 1971).

According to the appearance of *N. gracilis*, the base of the global Upper Ordovician Series could be traced in the East Baltic sections on the following levels: (1) near the upper boundary of the Kukruse Stage in NE Estonia (in the stratotype area of the Kukruse Stage); (2) in the middle of the Kukruse Stage in central and southern Estonia; (3) near the lower boundary of the Kukruse beds in western Latvia.

The correlation of the sections under discussion is based on the biomicrostratigraphy of chitinozoans in the East Baltic (Fig.) and detailed data from the Fågelsång section in Scania (Nõlvak & Grahn 1993).

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***Nemagraptus gracilis* (Hall)**
* 7 – Aizpute, 1064.4 m,
(see Männil, 1976, fig.2);
cf.* 8 – Ruhnu, 664.8 m;
cf.* 9 – Kamariku, 160.6 m;
* 10 – Valga-10, 407.8 m;
* 11 – Ruhnu, 662.8 m;
* 12 – Kamariku, 158.6 m;
* 13 – Palvere, 97.0 m;
* 14 – Palvere, 95.9 m;
* 15 – Laeva-13, 266.8 m;
* 17 – Palvere, 93.7 m;
* 18 – Rooküla, 70.5 m;
* 19 – Rooküla, 70.1-2 m;
* 20 – Kandava-25, 930.8 m,
(see Männil, 1976, fig.4);
* 21 – Kandava-26, 1037.5 m;
* 24 – Järva-Jaani, 165.2 m.
***Nemagraptus subtilis* Hadding**
1 - Kandava-25, 951.5 m,
(see Männil, 1976, fig.3);
2 – Lasnamägi (Tallinn), 0.18-23 m,
(see Männil, 1976, fig.2);
3 – Vikarbyn, 8 m, Siljan District,
Sweden, (see Männil, 1976, fig. 2);
4 - Ülemiste-12 (Tallinn), 23.6-7 m,
(see Männil, 1976, fig. 2);
cf. # 5 - Kandava-25, 947.3 m,
(see Männil, 1976, fig. 3);
cf. # 6 – Chudovo-50, 136 m,
Leningrad District, Russia;
cf. # 22 – Kandava-25; 930.5 m,
(see Männil, 1976, fig.4);
***Nemagraptus* sp.**
¤ 16 – Palvere, 94.8 m;
¤ 23 – Kandava-25; 930.4 m;

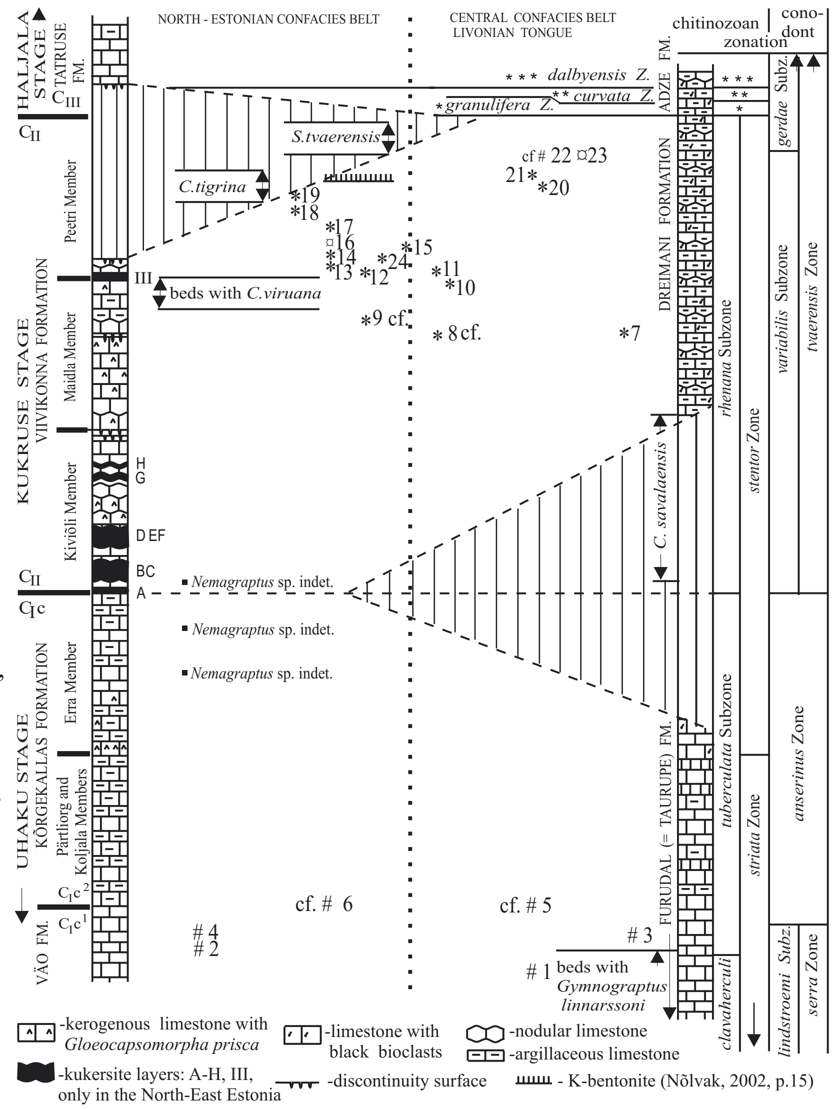

Fig. 1. Compound column (time-scale) of the upper Uhaku and Kukruse succession in the north-eastern Estonian (Savala core) and western Latvian (Kandava-25 core) sections showing the distribution of * *Nemagraptus gracilis* and # *N. subtilis.* The chitinozoans *Conochitina savalaensis* nom.nud., *C. viruana* nom.nud., *C. tigrina, Eisenackitina rhenana, and Spinachitina tvaerensis* are restricted only to the Kukruse Stage and the graptoloid *Gymnograptus linnarssoni* to the lower Uhaku Stage in all over Baltoscandia.

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New biostratigraphic and isotopic data from the Ordovician–Silurian boundary beds in the Copenhagen Canyon section, Nevada

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Ten samples from the Ordovician–Silurian boundary beds in the Copenhagen Canyon section in the Monitor Range, central Nevada, were collected in May 2003 by M. Harris, P. Sheehan and L. Ainsaar for the bio- and chemostratigraphical study. Mainly dark-grey calcareous mudstones and quartz-rich grainstones (CZ-9) belong to the Hanson Creek Formation and their detailed facies interpretation is given in Harris & Sheehan (1997). The average weight of our samples was 0.5-1 kg, except for CZ-8. The samples were processed using acetic acid.

Acid-resistant microfossils are very rare (see Fig.). Their preservation is bad to satisfactory: they are broken and very fragile but not flattened. Chitinozoans were found only from samples CZ-1 and CZ-5, and their fragments from CZ-7. Some very badly preserved scolecodonts were noticed in samples CZ-1, CZ-3, CZ-5, CZ-6 (*Oenonites* sp., *Mochtyella* sp.; identified by O. Hints). Their organic matter is changed, they are fragmentary (friable) and totally black. Obviously organic-walled microfossils — thin-walled chitinozoans, acritarchs, etc. — are not preserved, except in samples CZ-1, CZ-5, CZ-7, and barren beds are most probably influenced by thermal heat flow. Many fragments of black organic amorphous (bituminous) matter were found from samples CZ-3, CZ-4, CZ-6, CZ-7, CZ-10, however, without any certain figure or form and with unknown affinity. Dolomitization has also had certain influence on the preservation of the organic-walled microfossils, especially in the uppermost Ordovician layers. Well-developed rhomboidal crystals of dolomite were recognized in the insoluble residues from samples CZ-4, CZ-9, CZ-10, in lesser amount in the other samples. The grainstone (CZ-9) is full of well-rounded quartz. Similar layers are barren also in the sections of Baltica (in the Porkuni Stage) due to too active water facies for light microfossil groups. In the Monitor Range, the same is observed in the Martin Ridge section (Soufiane & Achab 2000, fig. 2).

Somewhat exceptional are some very rare light-brownish (phosphatic?) casts of ostracodes (2 specimens), micro-gastropods (2 sp.) and sponge spicules (4 sp.) from sample CZ-6. Other samples are almost barren except for calcitic crinoid ossicles (maybe a little silicified?) saved in weak acetic acid. A very interesting find is specimens of Muellerisphaerida (phosphatic, incertae sedis) from samples CZ-1 (1 specimen), CZ-5 (8 specimens), CZ-7 (fragments). Earlier finds include Ordovician forms from Baltoscandia and Tasmania, and Silurian ones mostly from Europe and Canada (Norford & Orchard 1985).

Conodonts were found in all samples except for two from the topmost Ordovician beds (Fig. 1). Samples CZ-1, CZ-2 and CZ-3 yielded only few specimens of typical Ordovician taxa. *Dapsilodus* sp. n. R appearing in sample CZ-5, is known from the Rhuddanian–lowermost Aeronian strata in Baltic. The fauna in sample CZ-6 is most similar to that occurring in the Rhuddanian–Aeronian transition interval in the Estonian sections. Sample CZ-7 shows similarity to the lower Telychian (*Pterospathodus eopennatus* Superzone), and CZ-8 to the upper Telychian (*P. a. amorphognathoides* Zone), but may lie even higher, in the lower Sheinwoodian (*P. pennatus procerus* Superzone). The uppermost three samples are characterized by abundant occurrence of conodonts, containing up to 450 specimens.

Chitinozoans in sample CZ-1 show similarities with those in the Hanson Creek Formation, Monitor Range composite section (Soufiane & Achab 2000, fig. 2). The level of that sample is approximately the boundary beds of the chitinozoan zones *Ordochitina nevadensis* and *Nevadachitina vininica* by Soufiane & Achab (2000, figs. 2, 3). The presence of zonal *N. vininica* itself can be proved after more careful work because *Nevadachitina* and *Ordochitina* by those authors are very similar and determination of so badly preserved specimens is difficult. The level of sample CZ-1 is most probably below the so-called beds with *Hirnantia* fauna, which in Baltoscandia can be compared with beds below the Porkuni Stage although chitinozoan assemblages are different. This correlation can be proved also by the character of the isotope curve (see Fig. 1).

The whole-rock carbon isotope analyses were made in Tallinn. The results (Fig. 1) coincide very well with earlier data from Copenhagen Canyon (Kump *et al.* 1999, fig. 1) and show exactly the same values. The overall positive excursion in $\delta^{13}C$ is well defined, which marks the interval of the Hirnantian glaciation (Brenchley *et al.* 2003, fig. 12). All revealed data support the conclusion that the Ordovician–Silurian boundary lies above the coarse-grained limestone layer, most probably between samples CZ-9 and CZ-5.

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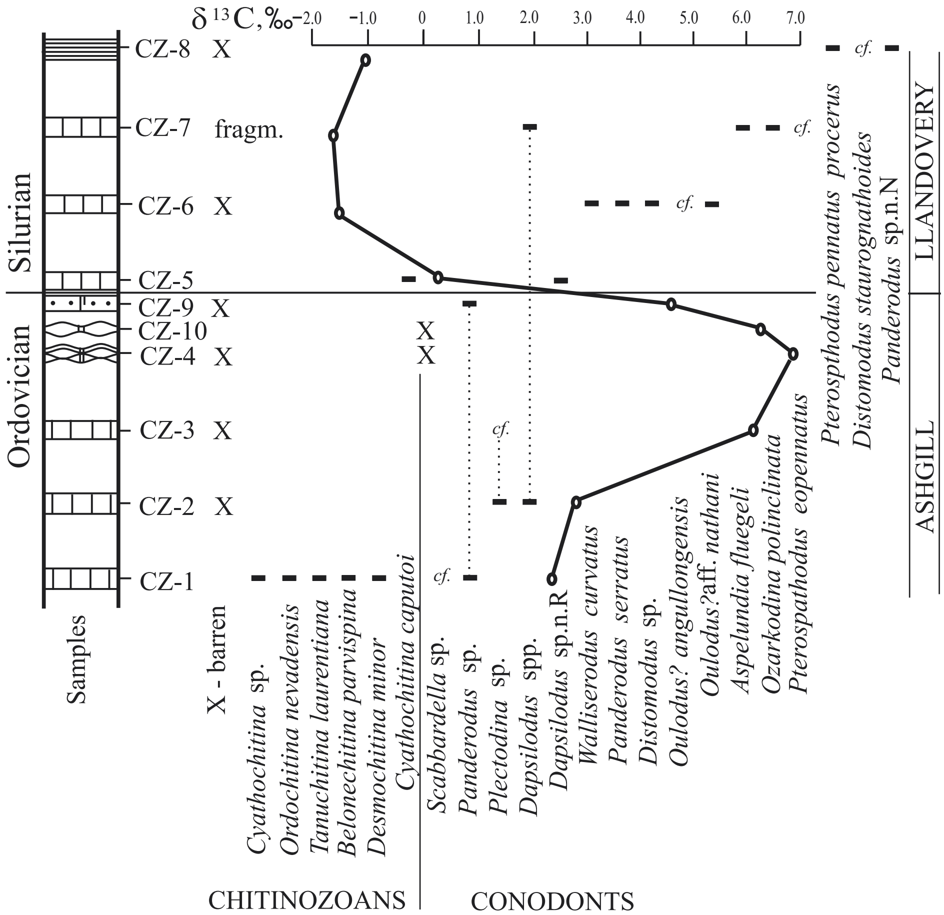

Fig. 1. Chitinozoan, conodont and isotopic data from the samples from the Ordovician–Silurian boundary beds of the Copenhagen Canyon section.

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The trilobite zonation of the Billingen Stage in the East Baltic

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In the northern East Baltic, the first Ordovician trilobites appear together with the carbonate in sediment in the Billingen Stage that embodies the Mäeküla, Vassilkovo and Päite members (see Fig. 1). Trilobites are very rare and usually poorly preserved in the Mäeküla Member, more common in the Vassilkovo Beds and abundant in the Päite Member. The first attempt to correlate the Billingenian rocks of the East Baltic with those of Scandinavia using trilobites ended up in the recognition of the Scandinavian *Megalaspis limbata* and *M. planilimbata* zones in the lowest limestone unit ($B_{II}\alpha$), and establishing the *Megalaspides* Zone in the topmost Glauconite Sandstone ($B_I\beta$) by Lamansky (1905). Balashova (1966) described five local trilobite zones within the current Billingen Stage and provided the correlation with the Swedish zonation, which was worked out by Tjernvik in 1956. Tjernvik's zonation was somewhat refined afterwards (Tjernvik & Johansson 1980) and is in general followed here.

Correlation of $B_I\beta$ with the ***Apatokephalus serratus*** **Zone** by Balashova (1966) (dotted lines in Fig. 1) was based on several erroneous identifications, like *A. serratus* [= *Remopleurides* sp.], *Euloma ornatum* [=? *Ptychometopus schmidti*], subspecies of *Pliomeroides primigenus* [= *Evropeites lamanskii* that is rather different for separation as an independent taxon, and was mistakenly identified as the Tremadocian *Triarthrus angelini* by Lamansky (1905)]. Also, a fragmentary pygidium attributed to the index species of the succeeding ***Megistaspis* (*Ekeraspis*) *armata*** trilobite Zone is too poorly preserved to allow a confident identification. Thus, these trilobite zones cannot be recognised in the East Baltic.

Recognition of the next Swedish zones in the East Baltic is somewhat complicated. In particular, the different growth stages of the isotelines (the index trilobites) vary considerably in their diagnostic characters (e.g. the number of pleural segments and length/width ratios of the axis or cephala and pygidia in general; see Nielsen 1995). Unfortunately, the hitherto published material on *Megistaspis* (*Paramegistaspis*) does not allow adequate comparison, and a special study of growth variation is needed. However, some dubiously identified pygidia of ***Megistaspis* (*P.*) *planilimbata*** from the Tosna locality indicate probable occurrence of that zone in the eastern Baltic. *Cybelopsis*? *linnarssoni* known from the *M. planilimbata* Zone in Skultrop, Västergötland, has also been identified from Lamoshka, which further supports that possibility. The trilobites, indicative of the ***Megistaspis* (*P.*) aff. *estonica* Zone** in Sweden, like *Pricyclopyge gallica* and *Raymondaspis brevicauda* are not recorded eastwards. However, there is potential to use some undescribed taxa mentioned by Tjernvik, belonging to *Geragnostus*, *Symphysurus* and *Megalaspides*.

In most of the studied localities the trilobite assemblage of the lowest part of the Mäeküla Member contains fragmentary *Rhinoferus* (*Popovkiaspis*) *leuchtenbergi* (synonymous with *Rhinoferus* (*Popovkiaspis*) *pogrebovi*), accompanied by *Evropeites lamanskii*, *Encrinuroides regularis* and *Pychometopus schmidti*. In addition, *Proasaphus primus* s.l. turns up, showing clear morphological evolution throughout the Billingen Stage. The agnostids *Geragnostus* and *Arthrorhachis* and raphiophorids *Globampyx* and *Ampyx*, close relatives of which are also common in the Swedish succession, appear in the overlying beds. Until further study proves the usability of Scandinavian isoteline zonation in the East Baltic, the earlier established ***Evropeites lamanskii*** **Zone** is accepted for this assemblage, as this taxon has been recorded from most localities. Yet, the last species is closely related to *E. toernquisti*, known only by the type specimen from the *M. dalecarlicus* Zone at Dalarna. If the appearance of this particular genus in the Baltoscandian Palaeobasin was synchronous, this assemblage could be correlated with the ***Megalaspides* (*M.*) *dalecarlicus*** **Zone**. However, *Evropeites* itself is apparently descendant from the Tremadocian *Anacheirurus* from Avalonia, showing its earlier appearance. Consequently, the study of trilobites shows that the faunal association of different beds of the Mäeküla Member is varying by localities, indicating diachrony of the lowest calcareous beds of the East Baltic as mentioned already by Pärnaste (2003).

Agnostids, pliomerids and encrinurids are lacking in the trilobite association of the more clayey Vassilkovo Beds, where both *M. dalecarlicus and M. paliformis* are common. The last is the index species for the **upper part of the *M. dalecarlicus* Zone** in Sweden. In addition, *Krattaspis*, *Proasaphus* and *Ottenbyaspis* are rather common, but *Rhinoferus* is rare.

A new, almost monospecific fauna comprising ***M.* (*P.*) *estonica*** appears in the Päite Member, marking the corresponding zone, which is well correlative all over Baltoscandia. Yet, some other poorly preserved and obscure isotelines were described by Balashova (1966) as *M.* (*P.*) *planilimbata rossica*, *M.* (*P.*) *putilovensis*, *M.* (*P.*) *popovkiensis* and *Megalaspides* (*Lannacus*) *popovkiensis*. *M.* (*P.*) *popovkiensis* and *M.* (*P.*) *scutata* show several similarities with *Megistaspis* (*M.*), being likely ancestors of the latter. Similarly, *M.* (*L.*) *popovkiensis* is the ancestor of *Rhinoferus* (*Lawiaspis*). *M.* (*M.*) *dalecarlicus balticus* apparently belongs to *Protoptychopyge*, bringing down the first appearance of that group. A few rare raphiophorids, niobinids, cheirurids and cybelines are recorded also from the Päite Member. The appearance of *Megistaspis* (*Megistaspis*) just above the discontinuity surface "Püstakkiht" marks well the Billingen–Volkhov boundary. Both, the index trilobite *M.* (*M.*) *polyphemus* (= *M. lata* synonymized by Nielsen in 1995) and *M.* (*M.*) *limbata baltica* are rather abundant in the overlying Saka Member of the Volkhov Stage.

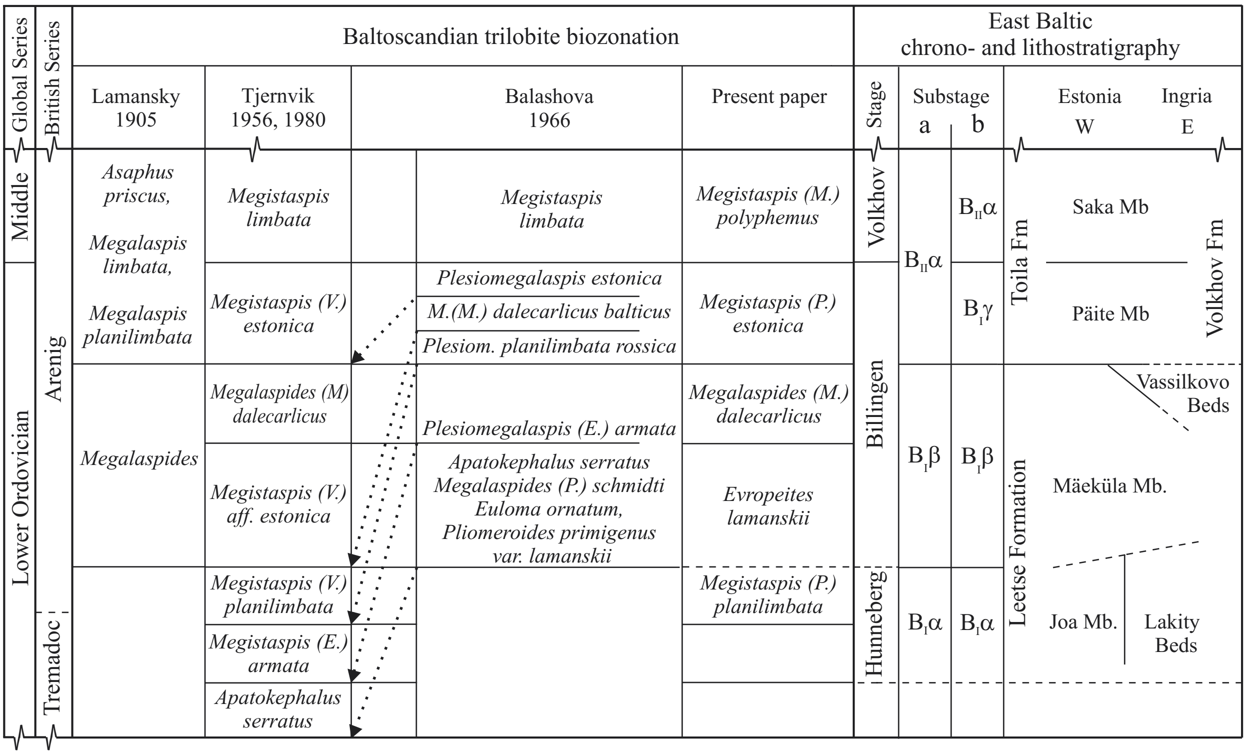

Fig. 1. The Baltoscandian trilobite zonation in correlation with chrono- and lithostratigraphy of the East Baltic. Dotted lines show the correlation of Balashova (1966), including B$_{I}\beta$ to the Tremadocian. Traditional substages: a, after Lamansky (1905); b, after Pärnaste 2003.

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Palaeomagnetic data of Early and Middle Ordovician carbonates in northern Estonia

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We present palaeomagnetic data of Early and Middle Ordovician carbonates from northern Estonia. The Ordovician sequence we studied is composed of limestones and dolomites of Arenig and Llanvirn age (485–464 Ma). Rocks of the Billingen to Uhaku Baltoscandian regional stages were sampled at 6 locations (see Figs 1 and 2).

To study the origin of the remanence components, we used alternating field (AF) demagnetization treatment up to 160 mT in steps between 2.5 and 20 mT. After each step the intensity and direction of the natural remanent magnetization (NRM) was measured with a superconducting (SQUID) magnetometer at the palaeomagnetism laboratory of the Geological Survey of Finland. The thermal treatments failed due to complex mineralogical changes during heatings at ~300°C.

The carbonates reveal mainly three remanence components. *First*, a component having a steep southeasterly downward direction yielding a pole that plots onto the Ordovician APW-curve of Baltica (Fig. 3). We interpret this component to be primary, because we observe a shift of poles from older to younger along the Ordovician APW-track when plotted in stratigraphic order. *Second*, a steep northeasterly downward directed component with its reversed counterpart represents most likely Permo-Triassic overprint. The Permo–Triassic overprinting increases and the Ordovician component vanishes towards northeastern Estonia, where the sediments are generally more dolomitized. *Third*, a steep northwesterly downward directed component with an unknown age.

The study is supported by the Estonian Science Foundation (grant No. 5500).

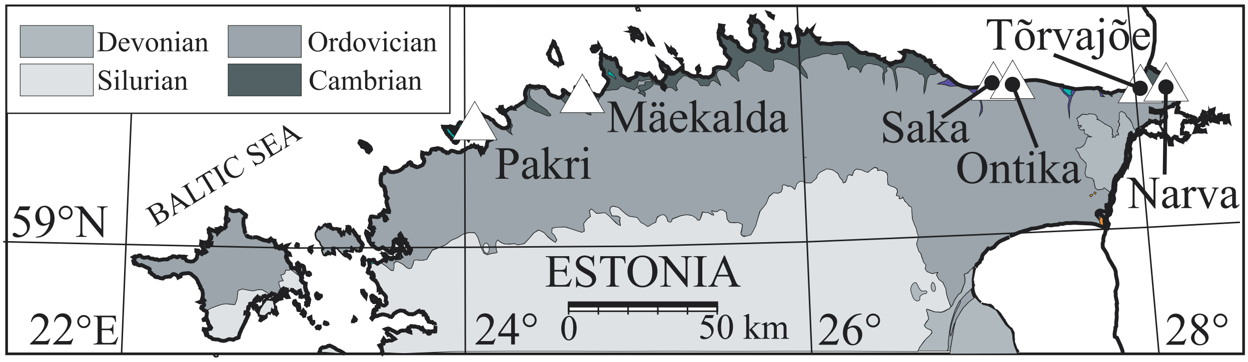

Fig. 1. Simplified geological map of northern Estonia. White triangles indicate the sampling locations.

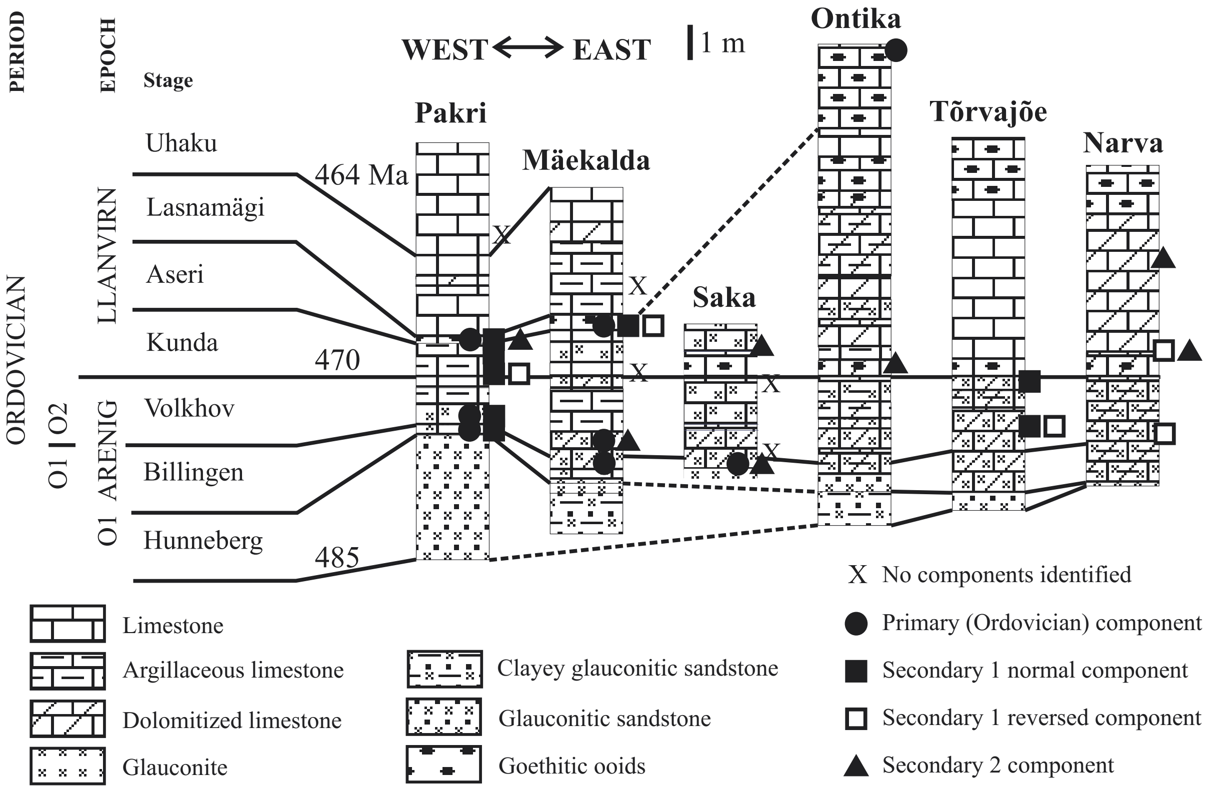

Fig. 2. Lithological sections and the identified remanence components at the sampled outcrops.

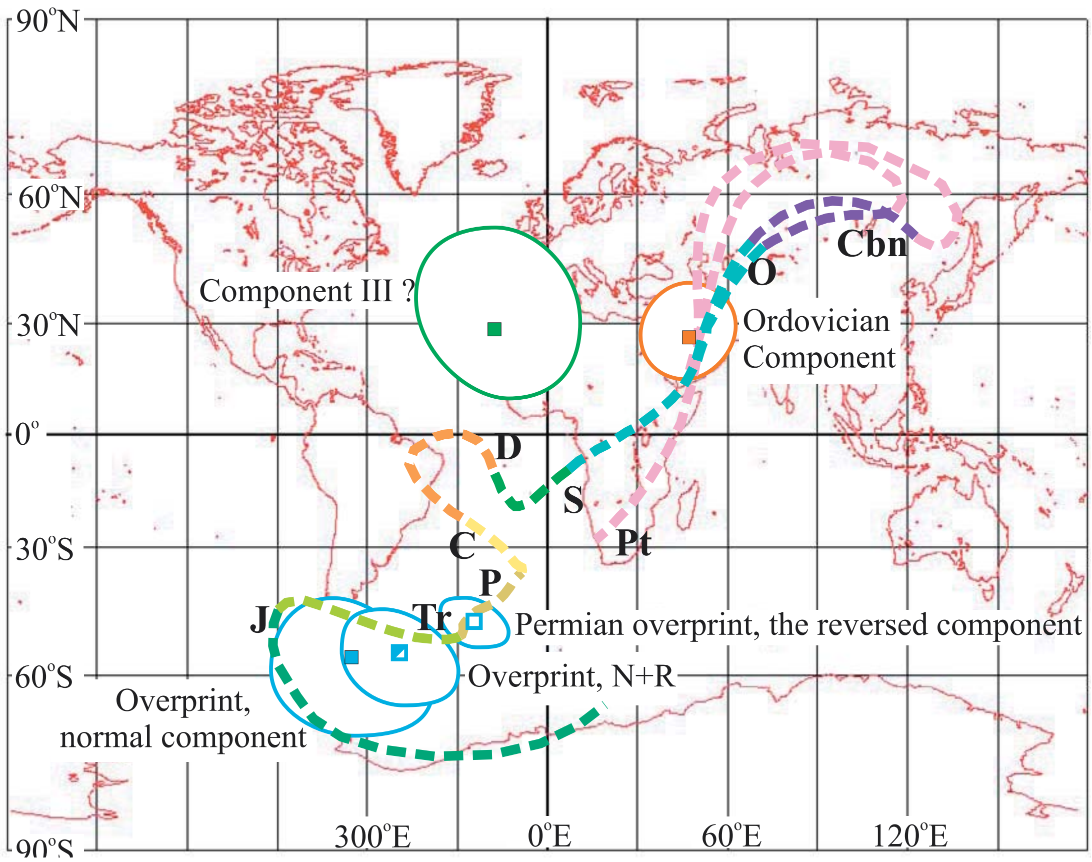

Fig. 3. The Apparent Polar Wander Path for Baltica from Late Precambrian to Mesozoic ages generated using the spherical splinefitting routine as implemented in the GMAP programme of Torsvik and Smethurst (http://www.geophysics.ngu.no). On top of the path three components, based on the present study, with 95% error ovals of confidence are plotted.

Tracing the base of the Ordovician System in Baltoscandia

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The GSSP of the base of the Ordovician System and the Tremadocian Stage of the Cambrian–Ordovician boundary defined by the first appearance of *Iapetognathus fluctivagus* in the Green Point Section, Newfoundland, was ratified by the International Commission on Stratigraphy in 2000 (Remane 2003). In this section, *I. fluctivagus* appears at the same level as *Cordylodus lindstromi.* In some other sections, e.g., Lawson Cove, Utah, *C. lindstromi* appears below *I. fluctivagus*, allowing defining a separate biozone. However, in the latter section, *Iapetognathus* appears in a relatively high stratigraphic position, only slightly below the appearance of *C. angulatus* (Dubinina 2000).

In the Ordovician correlation charts of the East Baltic published before defining the GSSP, the base of the Ordovician System coincides with the base of the Pakerort Stage, considered as the lowermost unit of the Ordovician in Baltoscandia since the definition of the Packerort Formation by Raymond (1916).

In the type area of the Pakerort Stage (the Pakri Cape), its basal beds beginning with a conglomerate correspond to the *Cordylodus proavus* Biozone. Except for occasional research visits, the Pakri Cape was an inaccessible military area before 1993 for more than 50 years. The concept of the Pakerort Stage was based on the eastern sections of Estonia, and all beds starting with the *C. proavus* Biozone were included to the Pakerort Stage. In a paper by Viira *et al.* (1987), part of the forms previously assigned to *Cordylodus proavus* were described as a separate species, *Cordylodus andresi* Viira et Sergeeva, the earliest representative of the *Cordylodus* lineage in Baltoscandia. Since then, the base of the Pakerort Stage has been defined by the first appearance of *Cordylodus andresi* in Estonia (Männil 1990; Männil & Meidla 1994). The *C. andresi* Biozone can be distinguished in some sections east of Tallinn (Heinsalu & Viira 1997), but not in the Pakri Cape, where *C. andresi* is reworked to the conglomerate formed in *C. proavus* time (Mens *et al.* 1996; Nemliher & Puura 1996).

At present, the base of the Ordovician defined by the GSSP correlates not with the base of the Pakerort Stage, but with a higher level within the stage. Puura & Viira (1999) suggested the subdivision of the Pakerort Stage into two substages. The base of the Pakerort Stage and the lower, Vihula Substage is defined by the first appearance of *Cordylodus andresi* in the Vihula section. The base of the upper, Karepa Substage is defined by the first appearance of *C. lindstromi* in the Toolse River section. This subdivision was accepted by the Estonian Commission on Stratigraphy in 1999.

The base of the *C. lindstromi* Biozone defining the base of the Karepa Substage is the closest traceable level in Baltoscandia approximating the base of the Ordovician as defined by the GSSP. Rare specimens of *Iapetognathus* sp. have been recorded in some Estonian sections and suggested as an additional criterion for tracing the Cambrian–Ordovician boundary (Heinsalu *et al.* 2003). Another additional criterion is the first appearance of graptolites *Rhabdinopora flabelliformis flabelliformis* and *R. f. sociale* occurring above *C. lindstromi* and *I. fluctivagus*, which is especially useful in sections with scarce conodont record.

In the western part of the Baltoscandian basin, where the Cambrian–Ordovician transition is represented by a monotonous sequence of the Alum Shale Formation, palaeontological criteria for defining the base of the Ordovician may be scarce. The first appearance of *C. lindstromi* can be recorded in some sections, e.g., Naersnes in the Oslo Region (Bruton *et al.* 1988), or Stenbrottet quarry at Orreholmen, Västergötland, Sweden (Löfgren 1996), where conodonts occur in stinkstone lenses. Where available, graptolites of the *Rhabdinopora flabelliformis* group refer to a traceable level approximating the base of the Ordovician. In many sections of the medium part of central Sweden and Öland, the base of the lowermost Ordovician beds recorded is usually referred to the *Cordylodus angulatus* or *Paltodus deltifer* biozones, whilst fossils from the Cambrian and *C. lindstromi* Biozone are often reworked to these beds (Puura & Holmer 1993).

In the outcrop area of the Cambrian–Ordovician boundary beds along and near the Baltic–Ladoga Klint, including northern Estonia and NW Russia (St. Petersburg Region), the closest level approximating the Cambrian–Ordovician boundary can be traced by the first appearance of *C. lindstromi* within a clastic sequence, e.g., within the Kallavere Formation in Estonia and in the Tosna Formation in NW Russia (Popov *et al.* 1989; Puura 1996). In core sections of Latvia, the basal beds of the Ordovician overlying a hiatus yield *Cordylodus angulatus* reworked to condensed beds together with conodonts from the *P. deltifer* Biozone (Ulst *et al.* 1982).

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The Kunda (early Mid Ordovician) brachiopod fauna from western Russia: data from Putilovo Quarry

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The Ordovician brachiopod fauna of the East Baltic are abundant, diverse and generally well preserved (Hints & Harper 2003). A number of groups, such as the distinctive clitambonitoids (Vinn & Harper 2003) are found almost exclusively in Baltoscandia and these together with a number orthide and pentameride groups have defined the Baltic brachiopod province. The Early Ordovician assemblages still retain some links with Gondwanan faunas, however the succeeding Volkhov assemblages are more individual. The Kunda brachiopod assemblage of western Russia reflects a major turnover in the faunas of the Baltic palaeobasin. The brachiopod fauna of the underlying Volhkov Stage is both abundant and diverse, dominated by the orthides *Paurorthis* and *Ranorthis*, together with the polytoechioids *Antigonambonites* and *Raunites* and the clitambonitoid *Apomatella* (e.g. Hansen & Harper 2003). Nevertheless close to the base of the Kunda Stage there is marked faunal change with the increasing dominance of *Orthis*, *Cyrtonotella*, *Lycophoria*, *Gonambonites*, *Ladogiella*, *Ingria*, *Ahtiella* and *Porambonites* in a variety of shallow-water, carbonate environments. New collections through the Kunda Stage, exposed in Putilovo Quarry, near St. Petersburg, have provided new data on the early Mid Ordovician brachiopod faunas of the region. The faunas are diverse although not particularly abundant. There was a striking diversification amongst the clitambonitides with species of *Antigonambonites*, *Clitambonites*, *Gonambonites*, *Hemipronites*, *Iru*, *Ladogiella* and *Neumania*? forming an important part of the fauna. The orthides are represented by species of *Orthis*, *Cyrtonotella*, *Orthambonites*, *Ranorthis*?, *Paurorthis*, *Platystrophia* and *Productorthis* together with the aberrant *Lycophoria*. But in addition there are a number of relatively common camarelloids, plectambonitoids and porambonitoids, some unusual at this level. Biogeographical analysis of the fauna suggests an increasing independence of the Baltic fauna during this interval, based on characteristic orthide-clitambonitide assemblages and associated with the shallow cool-water carbonate facies of the East Baltic. Nevertheless, of considerable interest are the Kunda plectambonitoid and porambonitoid associations that provide further endemic spikes in the region.

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Echinoderms of the bioherms of the Baltic Ordovician basin: comparison of the coldwater (Volkhovian) and tropical (Keila) communities

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At the beginning of the Middle Ordovician the echinoderms became the dominant group of the benthos in many biotopes of most of the epicontinental seas. Their evolution influenced the characteristics of the deposits at the sea floor and via that the development of other benthic animals. The Baltic palaeobasin can be used as an example of that. The first echinoderms appeared here at the end of the Billingenian. During the Volkhovian they quickly occupied the dominant position in the communities of the carbonate soft ground, hardground and bioherms. The echinoderms were especially diverse in the biotopes connected to bioherms because of the variety of environmental conditions.

Bioherms of the beginning of the Middle Ordovician were mud mounds covered by micrite with hard surface. The neighbouring bioherms often partially superimposed each other. There are several hypotheses of the origin of these peculiar bioherms, but the exact process of their formation is unclear. The superimposing of the mounds over each other, presence of the micrite with hardground and their erosion created in some places an intricate relief of the sea floor, influencing the water mobility and the character of the ground even in closely spaced areas. That is why the echinoderms settled mosaically, side by side, both on hard ground and soft ground, forming a single community. Almost all echinoderms, present in other two communities, are found on bioherms, but their numbers differ greatly. For example, in contrast to the soft ground community, the percentage of Rhombifera (*Echinoencrinites*), astroblastid diploporites and bolboporitids is less and the percentage of rhopalocystitid eocrinoids and perittocrinid crinoids is larger on bioherms. Hardground community members on bioherms are more difficult to characterise, as those are usually found outside their dwelling place and are more poorly preserved. Nevertheless, there are less rhipidocystids and more crinoids on bioherms than on hardgrounds outside bioherms. Some echinoderms are not found outside bioherms. These are endemic eocrinoids *Simonkovicrinus* and *Paracryptocrinites*. From the ecological aspect, the echinoderms on bioherms are passive suspension-feeders only, their tiering being from 0 to 1m over the surface of bioherms. During the Volkhovian the bioherm community changed insignificantly. By the beginning of Kunda time the bioherms stopped developing, and the bioherm echinoderm community disappeared. At the same time the hardground community disappeared as well. This was caused by the regression and the change in the currents flow. Palaeobaltica gradually shifted to the equator and in the course of time the climate became tropical. At that time the depth and territory of the Baltic basin changed. However, the fauna was not affected seriously by environmental conditions, which gradually changed from cold to warm. Palaeogeographical connections were changing: the basin separated from the Asian ones, and the connections with Laurentia increased. New immigrants were changing the taxonomic composition of the biota, including echinoderms. The local taxa also evolved. By the appearance of bioherms in Keila time, the composition of fauna changed considerably (in comparison to the Volkhovian fauna) on a rather high taxonomic level. The character of bioherm formation also changed. These tropical bioherms are well exposed in the quarries near Vasalemma. The bioherm system here consists of big carbonate bodies, formed around the edge of a large carbonate shoal. They are much larger than the Volkhovian bioherms, and the carbonate production of their inhabitants was 100–1000 times higher than that of the Volkhovian ones. Some parts of bioherms are built of algae, some of edrioasteroid echinoderm *Cyathocystis* and, sometimes, of various bryozoans. Separate bioherm bodies were set mosaically along the edge of the shoal, intercalated by soft grounds, protected from waves. Other places were filled by a great amount of carbonate debris, produced by destroyed skeletons of dead animals on bioherms. The most part of this debris was destroyed echinoderm ossicles. The volume of echinoderm debris exceeded the volume of all the remaining skeletal animals put together. The bryozoans occupied the second place.

The taxonomic diversity of echinoderms here was rather high, including members of 9 classes: Crinoidea (10 genera), Eocrinoidea (1 genus), Paracrinoidea (1 genus), Rhombifera (2 genera), Diploporita (2 genera), Edrioasteroidea (1 genus), Cyclocystoidea (1 genus), Homoiostelea (1genus) and Asteroidea (1 genus). Crinoids were the most diverse and made up half of the generic composition of echinoderms (Disparida – 4 genera, Hybocrinea – 1 genus, Cladida – 3 genera and Camerata – 2 genera). However, the major biomass and carbonate matter of echinoderms was produced by two genera: *Cyathocystis* (Edrioasteroidea) and *Hemicosmites* (Rhombifera). Thecae of *Cyathocystis* fusing to one another formed sometimes the frameworks of separate bioherms. *Hemicosmites* inhabiting bioherms was the main producer of carbonate debris. Fragments of its ossicles filled the gaps and reinforced bioherms. *Hemicosmites* debris formed large limestone beds as well.

Cyathocystis, being one of the framebuilders of Vasalemma bioherms, originally appeared in the Baltic basin during the Volkhovian. The earliest members of this genus were found in the middle Volkhovian mudstones in the east of the basin (Lynna River). These small (5–10 mm) cylindrical or conical animals were living on soft ground. Sometimes they are found in younger beds. For example, a group of *Cyathocystis* thecae fused to the upturned bryozoan colony was found in the Lasnamägi Regional Stage. However, mass settlements of *Cyathocystis* are found only in Vasalemma bioherms (Keila Regional Stage). The maximal size of the individuals from the Keila Stage is more than 10 times larger than that of Volkhovian individuals. Only two genera of edrioasteroids have been identified in the Ordovician of the Baltic basin, while in the North America their number is considerably higher.

Hemicosmites first appeared in the Baltic basin in the middle of the Kunda Regional Stage. These are the earliest members of this branch of Rhombifera. This genus is sporadically found in younger strata but its representatives became abundant in Vasalemma bioherms only.

The hybocrinid crinoid *Hoplocrinus* is characteristic of Vasalemma bioherms as well. It is one of the first echinoderms that appeared in the Baltic basin. One species of this genus is found in the Volkhovian bioherms. The individuals of this species are several times smaller than those of the Vasalemma species. Two genera of disparid crinoids (*Virucrinus* and *Ristnacrinus*) are, possibly, the descendants of the Volkhovian iocrinids, and two others, including a Calceocrinacea member, are immigrants from Laurentia. The cladids and camerats seem to be immigrant from Laurentia too.

Members of other echinoderm classes on Vasalemma bioherms are aboriginal. They appeared in the Baltic basin during the middle of Kunda time or little later due to warming and changing of the current configuration of the basin.

Thus, the echinoderm fauna of bioherms formed of the members of the communities of soft ground and hardgrounds around bioherms. However, their ratios were different from those outside bioherms. Most echinoderm genera of the coldwater bioherms of the Volkhovian sea disappeared by the end of Kunda time, that is, they only slightly survived the disappearance of bioherms themselves. Only the crinoid *Hoplocrinus* is an exception. Two echinoderm genera, which appeared in the Baltic basin originally on soft ground in the cold water of the Volkhovian sea (*Cyathocystis*) and in the slightly warmer waters of the Kunda sea (Hemicosmites), existed in the tropical Keila sea and reached an overwhelming abundance on the Keila sea bioherms.

The great increase in echinoderm diversity, individual size, quantity and carbonate production are the main features distinguishing the tropical Keila bioherms from the coldwater Volkhovian ones.

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Ichnofossils from the Lower–Middle Ordovician boundary interval in the St. Petersburg region, Russia

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The investigated region is located in the northwest of the Russian platform where Ordovician rocks occupy an elevated area called the "Ordovician Plateau". The plateau is bounded in the north by a prominent natural escarpment known as the Baltic–Ladoga Klint. The main natural outcrops follow the Klint line and are located in canyons and valleys of the rivers dissecting the plateau. The interval under consideration includes the Leetse Formation (Varangu, Hunneberg and partly Billingen regional stages) and the Volkhov Formation (uppermost Billingen and Volkhov regional stages). The Leetse Formation consists of quartz sandstone with high contents of glauconite grains, interbedded with glauconite-bearing clay limestone. The Volkhov Formation is traditionally subdivided into three units: (1) the "Dikary Limestone" ($B_{II}\alpha$); (2) the "Jeltiaki Limestone" ($B_{II}\beta$); and (3) the "Frizy Limestone" ($B_{II}\gamma$). The "Dikari Limestone" consists of up to 2.20 m thick hard, well-bedded, glauconitic limestone, varying from bioclastic packstone or grainstone to marlstone. The "Jeltiaki Limestone" consists of up to 1.6 m thick clayey limestone, yellow, red or variegated in colour, interbedded with clay. The 2.7 m thick "Frizy Limestone" consists predominantly of light grey nodular glauconitic limestone, intercalated with numerous lens-like beds of clay.

The Billingenian and Volkhovian beds in the vicinity of St. Petersburg provide an extremely rich fossil record of early deep bioturbation and early bioerosion. Twelve ichnogenera have been identified in the studied sections: *Gastrochaenolites*, *Trypanites*, *Circolites*, *Bergaueria*, *Chondrites*, *Palaeophycus*, *Thalassinoides*, *Dolophichnus*, *Phycodes*, *Planolites*, *Macaronichnus* and *Arenolites* (Dronov *et al.* 2002). The top of the Leetse Formation (Vassilkovo Member), interpreted as a transgressive systems tract of the Latorp depositional sequence (Dronov & Holmer 1999), is dominated by *Thalassinoides*, *Planolites*, *Palaeophycus* and *Chondrites.* The Billingenian part of the Volkhov Formation (highstand systems tract) is dominated by *Dolophichnus*. The previous phase in the substrate development showed basically the same genera as the top of the Leetse Formation (*Thalassinoides*, *Chondrites*, *Planolites*). The Volkhov Regional Stage represents a full cycle of deposition and is interpreted as a single depositional sequence. The hardground surfaces at the base and top of the Volkhov Stage are interpreted as the lower and upper sequence boundaries. The ten upper beds of the "Dikari limestone" represent a lowstand systems tract. The "Jeltiaki" and "Frizy" limestones represent the transgressive and highstand systems tracts, respectively. The hardground at the base of the sequence demonstrates a complex boring history with *Gastrochaenolites* aff. *oelandicus* as the prevailing component. The lowstand systems tract starts with beds having numerous, heavily bored hardgrounds, remnants of *Thalassinoides* ichnofabric and vertical *Dolophichnus*-like structures. The top of the "Dikari Limestone" (Bratvennik and Butok beds) shows a very specific trace fossil record including large *Bergaueria*, the system of *Phycodes* and network of *Thalassinoides*. The ichnologic record of the transgressive systems tract starts with a prominent hardground at the top of the "Dikari Limestone". It is marked with small *Trypanites* and probably also small rounded pits of *Circolites*. The following "Jeltiaki" section of muddy limestones intercalated with claystones has basically two common ichnotaxa, *Thalassinoides* and *Chondrites*, which often penetrates the fill of *Thalassinoides*. Specific beds are dominated by *Palaeophycus*, *Planolites*, *Macaronichnus* and *Arenolites*-like ichnofabrics. The ichnofabric of the highstand systems tract consists chiefly of *Thalassinoides*. The upper portion of the "Frizy Limestone" demonstrates, in addition, several beds with *G.* aff. *oelandicus*, *Bergaueria* and *Trypanites/Circolites* borings. The unconformity at the top of the sequence is marked by a hardground surface with *Trypanites*-like borings. Ichnofabric distribution patterns across the studied interval show a close relationship between the ichnofabric and the sea-level change.

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Geostatistic modelling of primary and diagenetic processes in the Ordovician sedimentary basin of Estonia

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Geological processes in Ordovician carbonate rocks were studied using multidisciplinary data including chemical, physical, geological and geographical parameters of samples from drill cores and outcrops. Geostatistic methods were applied to different Ordovician formations from Estonia. Examples from the Öland, Viru and Harju regional series of the Ordovician are shown and compared.

Twelve variables, related to latitude (Y in metres), longitude (X in metres), bulk and grain densities, porosity, SiO_2, Al_2O_3, K_2O, Fe_2O_3total, MnO, MgO and CaO contents of Ordovician rocks were examined statistically using regression and R-mode factor analysis to investigate relationships between them. For mapping the spatial distribution of the parameters reflecting geological processes computer gridding by the Kriging method was applied.

The Volkhov Stage (Öland Series, Arenig, Middle Ordovician) is distinguished in the Baltic Basin by wide regional distribution of carbonate-argillaceous rocks with glauconite impurites, regional dolomitization and a thin goethite ooid layer in northeastern Estonia (Meidla 1997). The dataset from the Volkhov Stage was represented by 37 rock samples from 19 localities (Fig. 1), including 10 limestones and calcareous marlstones and 27 dolomitized rocks. The three most important factors that accounted for about 77% of parameters variation were interpreted in terms of geological processes. The first factor was interpreted as primary sedimentation processes. They controlled the accumulation of siliciclastic and clay components (high loadings on SiO_2, Al_2O_3 and K_2O), which have positive correlation with primary porosity and negative correlation with bulk density. The second factor was interpreted as the "early diagenetic mineralization" process resulting in the formation of glauconite grains and iron (goethite) ooids (high loadings on total iron content). Total iron content had positive correlation with MnO content, which was higher in dolostones. The third factor was interpreted as dolomitization (high loading on MgO and negative – on CaO content). Dolomitization had regional control (high loading on longitude and significant on latitude). Dolomitization of the Volkhov Stage spread from north-east to south-west and was absent in the south-east and north-west of Estonia (Fig. 2). Dolomitization caused an increase in grain density.

The Lasnamägi Stage (Viru Series, Llanvirn, Middle Ordovician) is represented by the lower part of the Väo Formation in the north and by the Stirnas Formation in the south of Estonia. The Lasnamägi part of the Väo Formation is composed of the relatively argillaceous limestones of the Rebala Member, dolomitized Pae Member and hard limestones of the lower part of the Kostivere Member (Hints 1997). The data set from the Lasnamägi Stage was represented by 113 carbonate rock samples (14 localities), including 9 dolostone samples from the Pae Member taken from the Laagna outcrop in Tallinn and three dolostone samples from boreholes located in transitional and northern facies zones (Fig. 1, right). The three most important factors accounted for 85% of parameters variation. The first factor was interpreted as the dolomitization process. Dolomitization of the studied samples caused an increase in bulk and grain densities, dolomitization of the Pae Member caused an increase in Fe_2O_3total and MnO. The second factor was interpreted as the primary sedimentation process. The increase in clay content was facies-dependent and caused an increase in primary porosity. The third factor was interpreted as facies differentiation in trace element content (significant positive loadings on MnO and high negative loadings on latitude and longitude).

The Upper Ordovician rocks of the Harju Series are represented by argillaceous carbonate rocks of the Nabala, Vormsi, Pirgu and Porkuni stages. Sometimes glauconite and more rarely iron oolite impurities occur. Rocks may be dolomitized and are eroded in northern Estonia (Hints & Meidla 1997). The data set of the Harju Series is represented by 134 samples of carbonate rocks (12 localities), including 21 variously argillaceous dolostones. The first factor was interpreted as the primary sedimentation process, which controlled total iron content, primary porosity and bulk density. The second factor was interpreted as dolomitization, which was absent in western

Estonian islands (significant loading on longitude). The third factor was interpreted in the same way as in the Lasnamägi Stage as it had significant positive loadings on MnO and negative on latitude (increase in MnO content with a decrease in latitude).

Comparative analysis of the results of statistical analysis permitted us to draw the following conclusions. Primary sedimentation processes controlled siliciclastic-clay material deposition in the studied Ordovician carbonate rocks. The increase in clay content caused an increase in primary porosity and decrease in bulk density. Total iron content was controlled by the clay content of limestones. Glauconite and iron oolites could form during early diagenesis. The increase in iron content associated with dolomitization in most of the Ordovician dolostones, being the most significant in the Pae Member of the Väo Formation. The accumulation of manganese was facies-controlled in the Harju and Viru series and the MnO content was the highest in the south of Estonia. The porosity of dolostones was lower than primary porosity of limestones and the lowest among studied dolostones (about 2%) in the Pae Member. Both primary and secondary porosities of dolostones were the highest (up to 21%) in the rocks of the Harju Series (Fig. 2, left). Secondary porosity was determined in the drill cores of all ages from the North Estonian and transitional facies zones (Fig. 2).

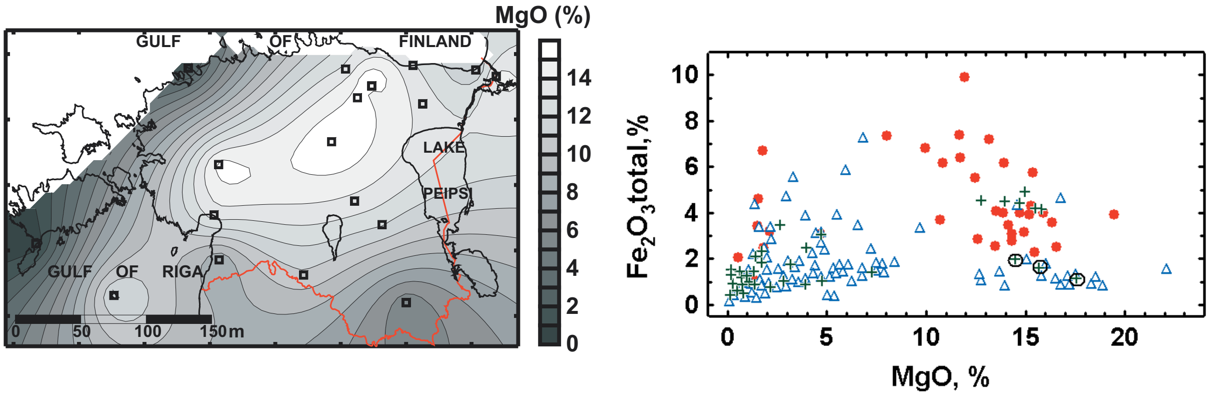

Fig. 1. Left – average MgO content in the rocks of the Volkhov Stage, as an indicator of dolomitization. Location of boreholes and outcrops is shown by squares. Right – total iron content versus MgO content in all studied rocks. Samples with MgO>10% are dolostones and dolomitic marlstones, those with MgO<10% are limestones and calcareous marlstones. Legend: Volkhov Stage (filled circles), Lasnamägi Stage (crosses), Nabala–Porkuni stages (triangles). Dolostones from the Pae Member (Laagna outcrop) are shown by crosses, dolostones from boreholes in the northern and transitional facies zones are shown by crosses in circles.

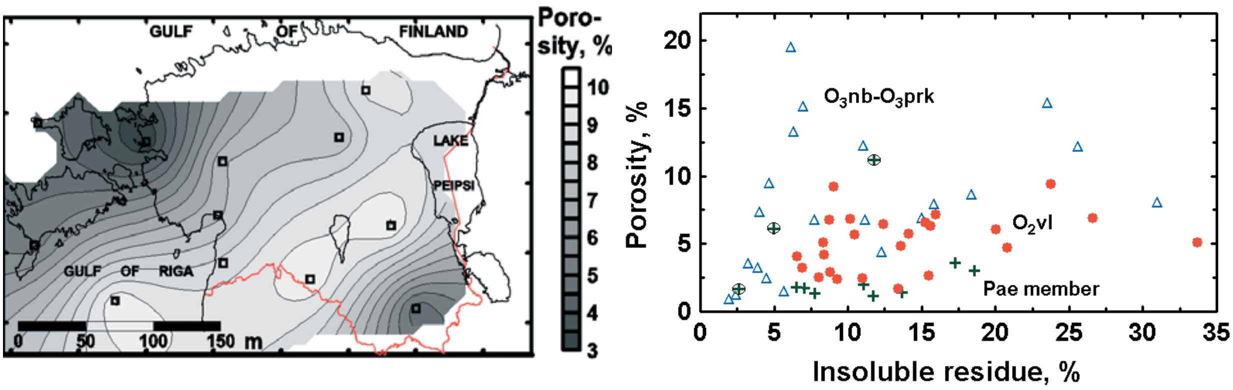

Fig. 2. Left – average rock porosity of Upper Ordovician rocks (Nabala–Porkuni stages). Right – porosity versus insoluble content of dolomitized rocks (MgO>10%). For legend see Fig. 1.

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Lower and Middle Ordovician conodonts of the Sutkai-86 borehole: a preliminary report

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In Lithuania multielement classification has not been used in the study of Lower and Middle Ordovician conodonts. No preserved conodont apparatuses have been found and often it is difficult to tell to which apparatus the particular specimen belongs. Now an attempt was made to apply multielement taxonomy to the Lower and Middle Ordovician conodonts of the Sutkai-86 borehole core (southwestern Lithuania). Eighteen samples were studied, two of which (from depths of 1128.4 and 1128.0 m) belong to the Lower Ordovician Latorp Regional Stage (RS). *Cordylodus angulatus*, *Paltodus deltifer pristinus*?, *P.* cf. *P. subaequalis*, *Oistodus lanceolatus*, *Drepanoistodus forceps*, *D. suberectus*, *Paroistodus proteus*, *P. numarcuatus*, *Scolopodus rex* and *Protopanderodus rectus* were identified at a depth of 1128.4 m.

Five samples (from 1127.6–1121.3 m) belong to the Volkhov RS. *Scolopodus rex*, *Protopanderodus rectus* and *Drepanoistodus suberectus* coming from the above mentioned depth extend their range here. *Cornuodus* sp.? and *Drepanodus arcuatus* enter the sequence at 1127.6 m, *Paroistodus originalis*? and *Baltoniodus parvidentatus*? at 1122.7 m, and *Drepanoistodus basiovalis* appears at 1123.6 m,

The species *Scalpellodus gracilis*?, *Scolopodus peselephantis* and *Semiacontiodus cornuformis* occur first at a depth of 1119.8 m (Kunda RS). *Drepanoistodus basiovalis*, *D. suberectus* and *Drepanodus arcuatus* range over from the older rocks. *Lenodus variabilis*? was established only at 1118.9 m. *Baltoniodus prevariabilis*? and *Protopanderodus greaei*? are characteristic of the upper part of the Kunda RS.

Scalpellodus gracilis? (1113.7 m) and *Protopanderodus rectus* (1111.2 m) range until the Aseri RS. *Drepanoistodus contractus*? was discovered at 1113.7 m. *Walliserodus ethingtoni*? was found slightly higher (at 1111.2 m). In these rocks also *Protopanderodus rectus*, *Drepanoistodus suberectus*, *Semiacontiodus cornuformis* and *Drepanodus arcuatus* were discovered.

The last occurrence of *Scolopodus peselephantis* (at 1108.8 m) was in the Lasnamägi RS.

Four samples were studied from the Uhaku RS. *Protopanderodus greaei*? extends to a depth of 1105.2 m, *Walliserodus ethingtoni*? to 1101.0 m and *Semiacontiodus cornuformis* to 1099.0 m. *Eoplacognathus pseudoplanus*? was established only at 1104.6 m and *Panderodus* sp. A? at 1099.0 m.

These are only preliminary data, but nevertheless it is obvious that the variety of conodonts in the studied section was largest at a depth of 1128.4 m (Leetse Formation). In the Baltic region the occurrence of *Protopanderodus rectus* is linked to the late Latorp *Oepicodus evae* Zone, and *Scolopodus rex* ranges to the Latorp RS. The Kunda RS is also distinguished by increased diversity of conodonts. *Semiacontiodus cornuformis* and *Lenodus variabilis*? are characteristic of the *Lenodus variabilis* Zone of early Kunda time (Löfgren 2000). *S. cornuformis*, according to published data on other regions, is found also in the older rocks of the Volkhov RS (Dzik 1994; Löfgren 1999).

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Ostracod biofacies in the Arenig Baltoscandian Palaeobasin

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Ordovician ostracods of Baltoscandia are the oldest known representatives of the group and among the most thoroughly studied ostracod faunas in the world (Williams *et al.* 2003). Biofacies differentiation among ostracods was first recognised in the Arenig (Tinn 2002) and the present study is aimed at documenting and dating this important event and grouping species with a similar spatial and temporal distribution pattern in the Baltoscandian Palaeobasin.

Quantitative data for 49 ostracod species from 14 sections and 295 samples were clustered using the unweighted pair-group average linkage rule and Euclidean distance measure. The analysis resulted in 13 clusters, defined by (one or several) dominant species and treated herein as ostracod assemblages. The distribution and alterations of ostracod assemblages are shown in a profile along the facies gradient, reaching from northern Estonia and the St. Petersburg region to western Latvia and Sweden (Dalarna and Västergötland) (Fig. 1).

The successive late Billingen–early Volkhov *Tallinnellina primaria*, *Rigidella mitis* and *Brezelina palmata* assemblages are of low diversity and show basinwide distribution, being documented in most studied sections in the middle (Väike-Pakri, Nõmmeveski, Kaugatuma and Lava sections) as well as in the outer ramp regions (Jurmala section). In early mid-Volkhov time the low-diversity *Ogmoopsis bocki* assemblage prevailed, being also documented almost all over the Palaeobasin. Biofacies differentiation first appeared in middle–late Volkhov times, resulting in two distinct ostracod biofacies and increasing diversity of the assemblages. The middle ramp region was inhabited by the *Incisua ventroincisurata* assemblage (Nõmmeveski, Kaugatuma and Lava sections). At the same time, the outer ramp region (Jurmala section) was inhabited by the *Glossomorphites digitatus* assemblage. Interfingering of strata with these two specific assemblages is recorded in the middle ramp sections.

Most of the Estonian sections are discontinuous in the Volkhov–Kunda transition. Detailed data from Dalarna (from Hessland 1949) demonstrate the prevalence of the *G. digitatus* assemblage in the Volkhov–Kunda transition, with some intervention of the *I. ventroincisurata* assemblage in the middle part of the composite section, corresponding to the early Kundan regression.

According to the present data, the Early Palaeozoic ostracods were benthic crawlers or swimmers, some probably adapted to infaunal life-styles (Siveter 1984). Ostracod distribution was controlled by substrate type and energy conditions (=depth). The same relationship is characteristic of the distribution pattern of modern ostracods. In this light, the basinwide distribution of ostracod assemblages in early Volkhov time is noteworthy. Although the Baltoscandian Palaeobasin was a considerably flat area without major depth differences (as suggested already by Jaanusson 1982), the principal facies belts still differ in substrate properties. Starting from mid-Volkhov time, distinct ostracod biofacies differentiation can be documented in the Baltoscandian Palaeobasin. This happened simultaneously with tectonically differentiated subsidence of the basin floor. Since late Volkhov–early Kunda time, the sea-level fluctuations and facies shifts can be traced by changing ostracod assemblages.

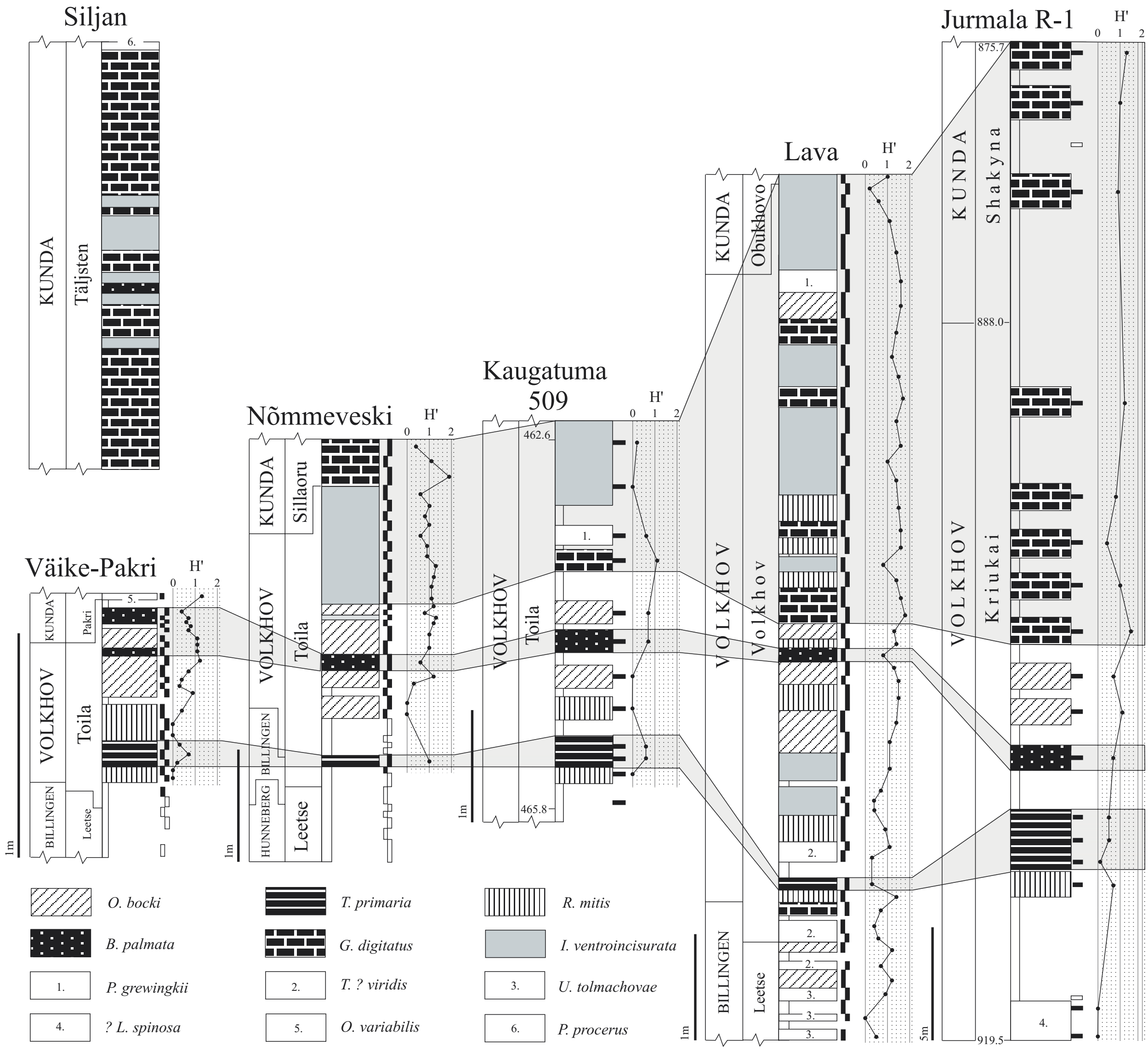

Fig. 1. The distribution of ostracod biofacies in selected sections of the Baltoscandian area. Composite data for the Siljan district by Hessland (1949). H' – Shannon-Wiener index.

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Pander’s heritage: finds of the undescribed collection

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One of the sad moments in the history of conodont studies is the loss of the type conodont collection of Christian H. Pander, the first discoverer of conodonts (Pander 1856). In his famous monograph he described and figured twenty-six new species and nine new genera of conodonts from the Lower Ordovician strata in the vicinity of St. Petersburg. Despite the absence of the type material, eight of the genera are still in use. The revision of Pander’s species is an important task of conodont studies, but it is seriously hampered by difficulties with the identification of Pander’s species in conodont collections from the corresponding stratigraphic level in the type area. Therefore the examination of any conodont material collected by Pander is important.

A small collection of Lower Ordovician conodonts evidently collected by Ch. Pander was recently found in the Mining Institute, St. Petersburg, Russia. It consists of two small test tubes both containing not less than 500 conodont elements, but there are no documents that could provide information on the sampled locality or conodont identifications. In spite of this fact the following observations are possible.

Both test tubes contain conodont elements of thirteen species typical of the upper part of the *Prioniodus elegans* Zone. This is the same stratigraphic interval where Pander (1856) described conodonts. The most numerous elements in the collection are the large hyaline elements of *Scandodus furnishi* Lindström, *Oistodus lanceolatus* Pander and *Scolopodus rex* Lindström, which constitute not less than 80% of the collection. *Tropodus* cf. *T. comptus* Branson & Mehl, *Tripodus* sp., *Drepanodus arcuatus* Pander, *Paltodus subaequalis* Pander, *Paroistodus proteus* (Lindström) and *Drepanoistodus forceps* (Lindström) are also relatively abundant whereas *Prioniodus elegans* Pander, *Protoprioniodus costatus* (Van Wamel), redeposited elements of *Cordylodus angulatus* Pander and gerontic elements of unknown genera are represented by single elements. The species number in the collection does not reflect the whole species diversity that was recorded in the sandstone layers of the upper part of the *P. elegans* Zone. The conodont assemblages of that time interval are very diverse in the type area and usually comprise not less than 20 species.

The majority of conodont elements in the test tubes are large and unbroken. This fact suggests that Pander selected conodonts by size and completeness during the picking. The collection contains considerably more elements of relatively larger species, in contrast to the actual conodont assemblages from the *Prioniodus elegans* Zone in several localities mentioned by Pander. Representatives of small species or juveniles are less common in the collection or even absent. Moreover, the actual conodont elements are not very well preserved; most of them have broken cusps and chipped edges of the base.

The new collection shed light on some strange observations of Ch. Pander. For example, he has mentioned that *Scolopodus* elements are white, whereas in the material collected recently from the type area hyaline elements of *Scolopodus* are always yellow. However, a quarter of *Scolopodus* elements in the collection are white.

Ordovician sea-level changes in the Małopolska Block (south-eastern Poland)

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The sections located across the southern part of the Holy Cross Mts., i.e., the Kielce tectonic unit (KU), were studied. The unit is considered to be the northern, exposed part of the Małopolska Block (MB) (Fig. 1). There are two different points of view on geotectonic provenance of the MB: the first one treats it as a proximal terrain detached from Baltica before the Ordovician (Dadlez *et al.* 1994; Narkiewicz 2002), the second one considers it as Gondwana-derived terrain (Belka *et al.* 2000). In the Early Ordovician the MB lay at about 60° S latitude (e.g. Lewandowski 1993) close to Baltica and moved gradually towards the north, reaching the middle latitudes of the Southern Hemisphere in the Late Ordovician (e.g. Dzik & Pisera 1994; Trela 2004). According to Cocks (2002), the MB was a part of Baltica in Early Palaeozoic times. Palaeomagnetic data indicate that the MB was located in the present position at the SW margin of Baltica as early as the Late Silurian (Nawrocki 2000).

A model involving three highstand and three lowstand sea-level intervals correlated with 2nd order sea level changes established by Nielsen (2003) for the Ordovician of Baltoscandia is outlined for the MB. The following intervals are distinguished:

- The **late Mid-Late Tremadoc highstand** interval, recorded on the whole MB by condensed, glauconite-rich facies that is a diachronous horizon reflecting the landward migration of an environment of slow net accumulation. This horizon exhibits the onlap pattern, which is associated with the ensuing transgression and is common along passive continental margins. The traces of multiple reworking by various processes, chiefly storm waves and currents, led to sediment bypassing and winnowing and subsequently to sediment condensation (Trela 2002).
- The **latest Tremadoc–Early Arenig lowstand** interval coeval with the Ceratopyge Regressive Event (Erdtmann 1986). The related facies are associated with the coastal deltaic fan system along the margins of the tectonically active basin. The emersion of some localities together with the ensuing later transgression generated the erosional unconformity and hiatus representing a relatively long time interval.
- The **late Early Arenig–earliest Late Arenig highstand,** recorded by fine-grained terrigenous and carbonate facies, e.g., heterolites, marlstones and limestones of the open shelf environment (Trela 2002). A thin shell-rich glauconitic lag deposit (associated with rapid flooding) occurs at the base of this interval. The maximum flooding may be correlated with the *gibberulus* graptolite and *navis* conodont zones.
- The **Late Arenig–Early Llanvirn lowstand** that resulted in the deposition of storm-dominated quartzitic sandstones (poor arenite). The final stage of this interval is recorded by a sedimentary discontinuity in Mójcza and a short-time hiatus (lower Llanvirn) coeval with the Viru unconformity in Baltoscandia (Dzik & Pisera 1994).
- The **latest Late Llanvirn–Late Caradoc** long-lasting highstand interval with a short lowstand period. In the southern part of the KU (Zbrza-Brzeziny area) it correlates with the *teretiusculus*/*gracilis* and *multidens*/*clingani* black to dark grey shale horizons, in the northern part of the KU (Mójcza) with phosphatized skeletal grainstones and (phosphatic) ooidal grainstones related to the maximum flooding period (Trela 2004).
- The **latest Late Caradoc–Ashgill** lowstand interval characterized in the southern part of the KU (Zbrza) by the occurrence of green mudstones/claystone shales exhibiting an upward trend of increasing levels of dissolved oxygen in the bottom waters (or sediment–water interface), and finally by coarser-grained siliciclasic deposits (quartz wackes and marly siltstones) of the late Ashgill. In the northern part of the KU grainstones pass upward into lowstand-related skeletal and oncoidal packstones (Trela 2004). The facies turnover was coeval with changes in palaeoceanographic circulation related to gradual closure of the western part of the Tornquist Sea during the Late Ordovician due to collision of E Avalonia with Baltica and glacioeustatic sea-level drop (Trela 2004).

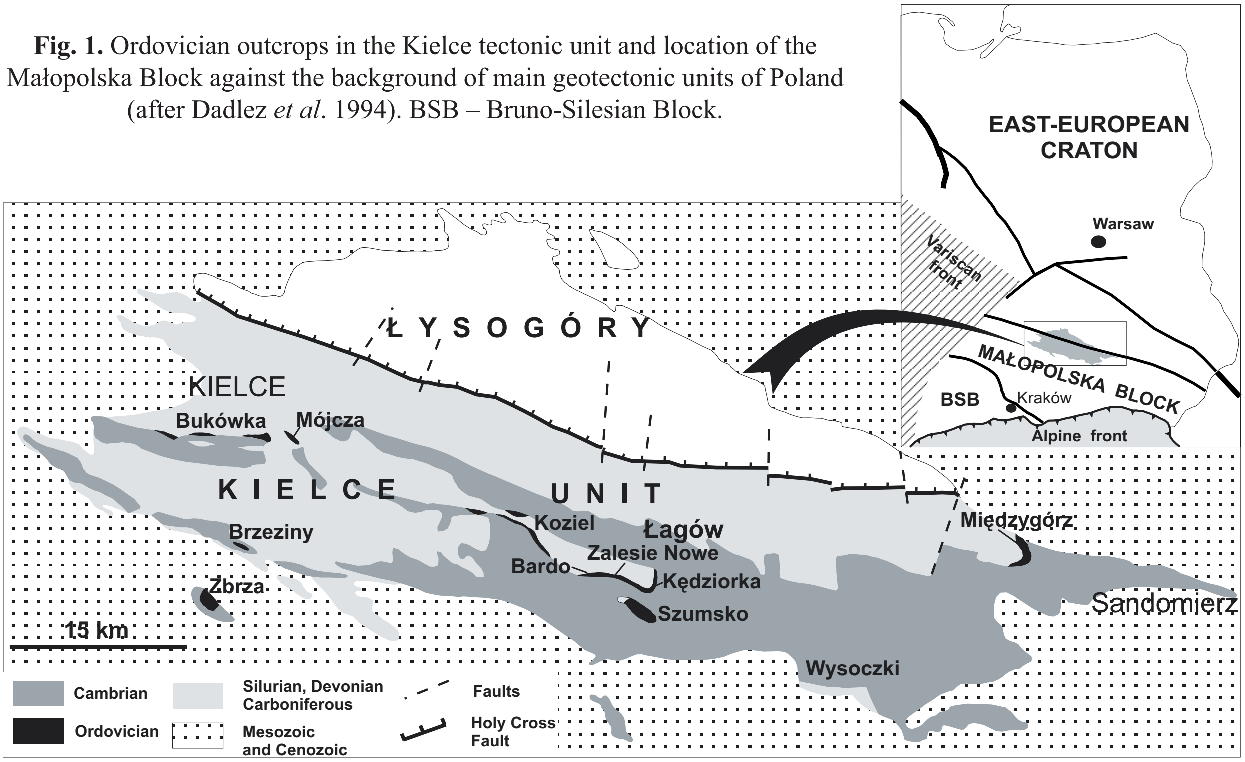

Fig. 1. Ordovician outcrops in the Kielce tectonic unit and location of the Małopolska Block against the background of main geotectonic units of Poland (after Dadlez *et al*. 1994). BSB – Bruno-Silesian Block.

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Sedimentation, erosion and redeposition of sediment and conodont elements in the upper Tremadoc boundary beds of Cape Pakri, NW Estonia

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Cape Pakri is situated c. 40 km west of Tallinn in northern Estonia. The Pakri coast has been protected as a National Landscape Reserve since 1998. Along the coast of Cape Pakri and the adjacent Pakri Islands the escarpments expose a succession of sedimentary beds ranging from the Lower Cambrian into the Middle Ordovician, c. 25 m in all. In the mid-1990s several geological research projects were initiated in this area. One of these, the investigation of the biostratigraphy (based on conodonts) and sedimentary history of the upper Tremadoc boundary beds at Uuga Pank in the west-central part of the peninsula is presented here.

A combination of sedimentological and biostratigraphical criteria has clarified the tempo and mode of the processes that formed these glauconitic sandstone boundary beds at Cape Pakri. The sediments include some quartz grains derived from the nearby Cambrian deposits, glauconitic sand grains formed on the seafloor during sedimentation or redeposited slightly later, and clay and lime mud as well as a mixture of redeposited and indigenous conodont elements. Apatitic conodont elements are particularly well suited for tracing the geological history of the surrounding sediment, since they can be repeatedly included in the sediment, eroded and redeposited. These processes often leave telltale marks on the elements that are nevertheless identifiable. We have sorted out the indigenous elements from different generations of redeposited elements, and located the local upper boundary of the Tremadocian to slightly more than 1 m above the base of a c. 4 m thick sandy interval, which is the main object of our study.

In order to solve the sedimentological and biostratigraphical problems, a "redeposition index" was used by combining the relative percentages of:

(a) conodont elements with a source below and beyond the section at Pakri,

(b) conodont elements derived from strata laterally equivalent to those in the investigated section, but below the stratigraphical level of the actual sample, and

(c) elements derived from conodont animals that lived at approximately the same time as when the sediments with their enclosed elements were finally deposited.

The relative abundance of the different kinds of redeposited elements was subsequently analysed. This revealed that the mixed sandy-clayey sediments and argillite (clay) at the base of the section generally included less than 50% redeposited elements and confirmed the fact that this part belongs to the *Paltodus deltifer* Zone, Varangu Stage. The succeeding sandy interval (Leetse Formation), which makes up the *Paroistodus proteus* Zone and includes 58–97% redeposited elements, was formed during several depositional phases. The proportion of redeposited conodont elements is thus clearly correlated with the degree of sorting (grain size) of the clastic material.

The upper part of the section, belonging to the *Oepikodus evae* Zone, consists of limestone beds with successively diminishing amounts of redeposited conodonts.

Trypanites borings in the Estonian Caradoc oil shale brachiopods

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During Kukruse (Caradoc) time, light-brown organic matter forming oil shale interlayers in calcareous sediments accumulated in the Baltoscandian Basin (Nestor & Einasto 1997). The oil shale is composed of almost equal portions of carbonate, and clayey siliciclastic and organic matter. The existence of a rich and diverse normal marine bottom fauna in the Estonian oil shale (Rõõmusoks 1970, p. 172) and a very low content of pyrite suggest the absence of anoxic conditions in the bottom waters (Bauert & Puura 1990). It is supposed that most of the organic matter was derived from algal mats (Kõrts & Veski 1994) covering extensive tidal flat areas, from where it was transported to shallow subtidal environments (Puura *et al.* 1988). The organic matter accumulation took place during the Late Llanvirn–Early Caradoc regression (Puura *et al.* 1988).

Abundant *Trypanites* borings were found in brachiopods *Clitambonites*, *Estlandia*, *Nicolella*, in the Upper Ordovician (Caradoc) oil shale in northern Estonia. About 43% of the 21 studied brachiopod genera are bored. *Trypanites* is host-specific, and the rate of bored valves varies from 6.5% in *Bekkerina* to 51% in *Estlandia*. The majority of the borings are oriented, and living hosts were preferred to dead shells. The few bored valves bear the blister-like shell repair structures in their interiors.

The brachiopods of the oil shale taphocoenosis presumably are a mixture of at least two separate assemblages. Brachiopod taxa without borings could belong to an autochthonous assemblage in a region of organic-rich mud. The bored taxa are possibly allochthonous and lived together with boring organisms in the inner sublittoral zone, just below the low tide mark on hardgrounds or areas of very low sedimentation. Later they were transported into deeper parts of the basin, where organic-rich sediment was deposited.

An alternative explanation would be the mixed taphocoenosis of brachiopod shells derived from different time intervals. Then the bored shells should have undergone re-sedimentation and represent the time interval of very high boring activity.

According to this explanation the boring activity should have been high during the time intervals separating the periods of oil shale accumulation. The obviously smaller number of brachiopod shells affected by *Trypanites* borings in the limestones surrounding the oil shale deposits does not support this explanation.

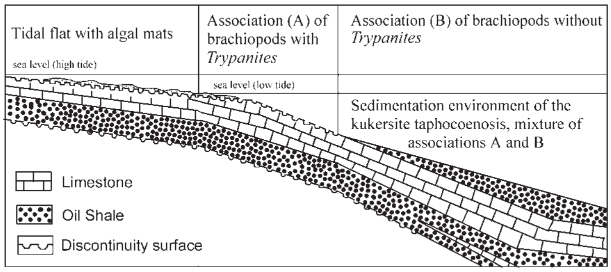

Fig. 1. Depositional model of the oil-shale basin showing distribution of brachiopods with *Trypanites* borings.

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Cornulitids, problematic Ordovician calcitic tubicolous fossils

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In 1820 cornulitids were first described by Schlotheim as the Palaeozoic tubicolar annelids from the Silurian of Gotland (Schlotheim 1820). They range from Middle Ordovician (Öpik 1930; Fisher 1962) to Carboniferous (Fisher 1962) and have variable morphology. They achieved cosmopolitan distribution in the Late Ordovician. The macroscopic details of the external morphology were relatively well known already by the end of the 19th century (Hall 1847; Nicholson 1872a, b; Nicholson 1873; Vine 1882; Hall 1888). Fisher (1962) included four genera – Cornulites, Conchicolites, Cornulitella and Kolihaia – into the family Cornulitidae. The ultrastructure of cornulitid shell has never been systematically studied. The only published material is the abstract by Blind (1972) on cornulitid shell structure, in which he mentions that a nacreous layer occurs in the shell wall. That could not be supported by our study because cornulitids obviously had original calcitic shells.

The following differences were found between the members of two cornulitid genera, *Cornulites* spp. and *Conchicolites* spp. Both genera have egg-shaped embryonic shells, which presumably calcified after the settling of larva to the substrate, but the embryonic shells in *Cornulites* are larger than in *Conchicolites*. *Cornulites* has the foliated shell ultrastructure and pseudopuncta, whereas the shell ultrastructure in *Conchicolites* is prismatic. In *Cornulites* the outer part of the shell contains more or less numerous vesicular cavities that were never observed to cross the borders of the surface annulae, indicating a cyclic shell formation.

Our studies show that extant serpulids can produce vesicular wall structure on their shells in contact with the substrate. However, this structure is not homologous with the vesicles in *Cornulites* because it consists of calcareous lamellae projected from the aperture and leaving some empty spaces between successive sets of lamellae. Besides, the aragonitic shell is made up of two oriented layers: an irregular spherulitic prismatic layer and a fine complex crossed lamellar layer.

In *Conchicolites* the vesicular shell structure is absent and the calcitic prisms are deposited at the shell aperture more or less at right angles to the longitudinal shell axis.

The shell structure and shell secretion of *Cornulites* have most in common with those of lophophorates, comprising brachiopods, bryozoans and probably also tentaculitids. The only unique feature for *Cornulites* is the cyclic structural change in each annulae. On the other hand, they seem not to have been closely related to *Conchicolites*.

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The Upper Ordovician and lower Silurian *Platystrophia*-like brachiopods from Sweden

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Occurrences of orthid brachiopods, usually identified as *Platystrophia*, have repeatedly been reported from the Upper Ordovician (Caradoc, Ashgill) and lower Silurian (Wenlock) of Sweden (Lindström 1880, 1888; Wiman 1907; Troedsson 1921; Jaanusson 1960, 1963, 1982). However, detailed studies of their morphology and taxonomy are lacking. It has already been demonstrated in the East Baltic and St. Petersburg Region that platystrophiid brachiopods are useful for detailed biostratigraphy. In particular, Paškevičius (2000) subdivided the Ordovician deposits of Lithuania (LCB) into nineteen biostratigraphical units based on brachiopods, whereas six of them were based on a taxon of *Platystrophia* (s.l.) as the index species. The biostratigraphical significance of platystrophiids for the North Estonian Confacies Belt (Estonia, St. Petersburg region, north-western Moscow Basin) is also well established (Alichova 1953, 1960).

This report presents preliminary results of the study of Swedish *Platystrophia*-like brachiopods from the Kullsberg and Boda limestones, housed in the Palaeontological Museum of Evolution, Uppsala University, and in the Department of Palaeozoology, Swedish Museum of Natural History, Stockholm. These are traditionally referred to the genus *Platystrophia* (s.l.).

Differences in the morphology of the dorsal cardinalia of Swedish platystrophiids allow recognition of two different morphological groups. Shells from the Kullsberg Limestone (Caradoc) show a dorsal cardinalia, which is similar to most of the contemporaneous species from other parts of Baltoscandia and eastern Avalonia. The species *Platystrophia lynx lynx* Eichwald (1830) can be definitely identified. Probably it is the westernmost occurrence in Baltoscandia of the species otherwise widespread in the East Baltic. However, the Ashgill Boda Limestone of the Siljan district (Province of Dalarna, Sweden) contains also shells with a different morphology of dorsal cardinalia somewhat comparable to the cardinalia of *Gnamptorhynchos globatum* (Twenhofel 1928) from the Ashgill of Anticosti Island, Canada (Jin 1989; Jin & Zhan 2000) and of an unnamed species from the Llandovery of UK. The Silurian species, known as *Platystrophia jaaniensis* Rubel (1963) from Gotland, also differs markedly from the typical *Platystrophia* that occur in the Caradoc of Sweden and the East Baltic in characters of cardinalia and dorsal muscle field. Taxonomically the shells from the Boda Limestone were recently assigned to a new genus (Zuykov & Egerquist, in press.) whose affinities to the family Platystrophiidae, however, need further confirmation.

Platystrophia-like brachiopods belong to the most easily recognizable and widespread brachiopods in the Middle and Upper Ordovician of Baltoscandia. A sequence of short-lived species of these brachiopods can be potentially used as a basis for the biostratigraphical zonal scheme for the shallow-water facies in the Baltic basin, as an alternative to the microfaunal, trilobite or other classical biostratigraphical subdivisions.

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Excursion Guidebook

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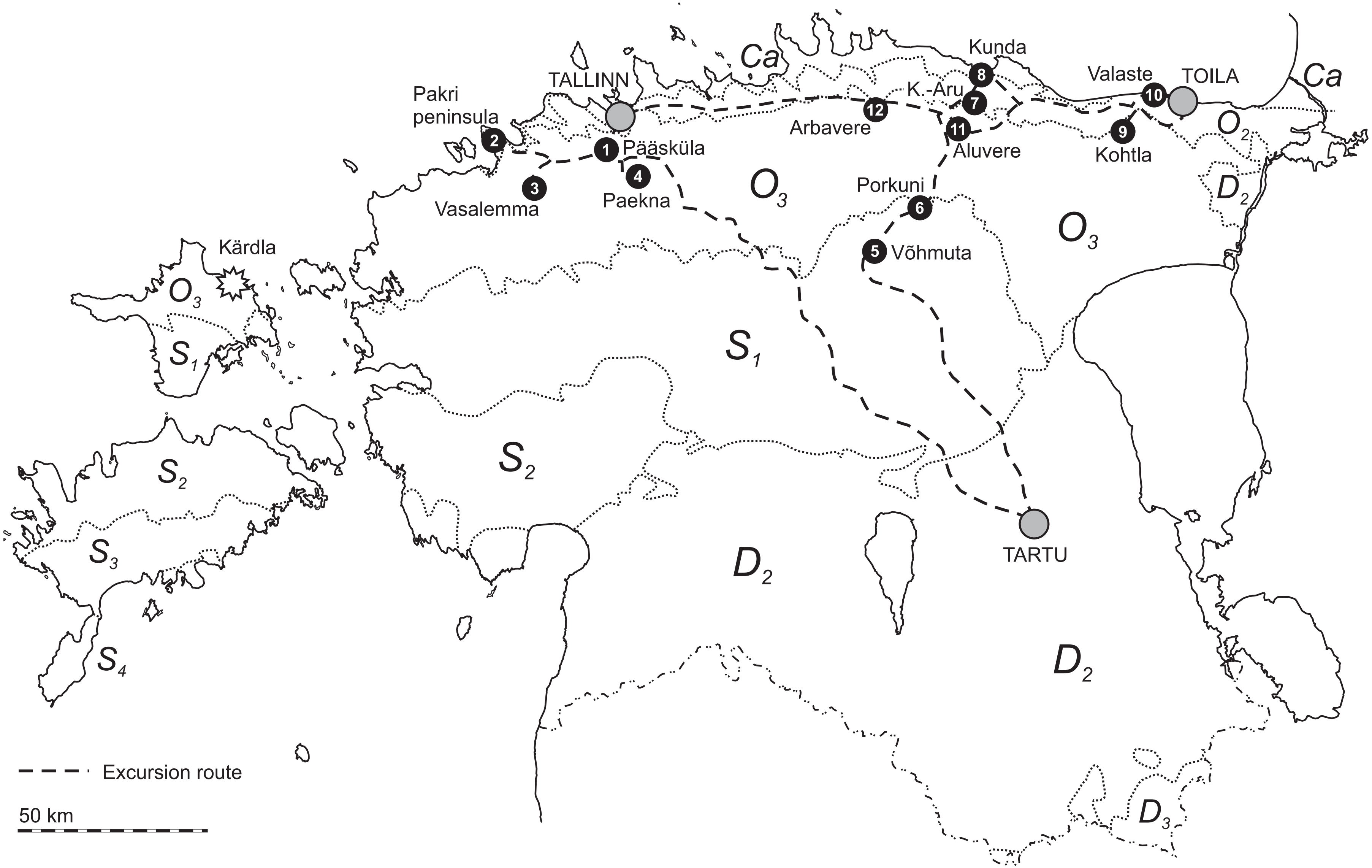

Fig. 1. Schematic geological map of Estonian bedrock with excursion routes.

On the Ordovician System in Estonia

Tõnu Meidla and Leho Ainsaar

The main distribution area of the Ordovician strata in the East European Platform extends from the Gulf of Finland in the north to Belarus and Poland in the south, and from the Baltic Sea islands in the west to the vicinity of Moscow in the east. Within this area, beds are exposed in the magnificent sections of the Baltic–Ladoga Klint, in several river bank sections, old and new limestone quarries and open cast pits of northern Estonia and north-western Russia. Good accessibility of strata, excellent preservation of fossils and sedimentary structures and perhaps also the characteristic succession of the Cambrian to Middle Ordovician, represented by several distinctive rock units (like phosphatic brachiopod coquina, *Dictyonema* argillite, dark green glauconite sandstone, etc.), attracted the attention of investigators already in the early 19th century (Engelhardt 1820; Strangways 1821; Eichwald 1825).

The main features of the Ordovician stratigraphy were brought to attention already by F. Schmidt in his thorough monographic paper of 1858. The general pattern of his geologic map, presented in the same volume, is well recognised in the modern bedrock maps of Estonia. The generally simple geologic structure of the area – with almost horizontal strata, only 2–5 m/km dipping to the south – results in nearly latitudinal orientation of the outcrop belts of the Ordovician stages in northern Estonia (Fig. 1).

The main part of the Ordovician succession in northern Estonia is composed of various kinds of limestones, with some intercalations of kukersite oil shale, concentrated mainly in the Kukruse Stage. Only the basal strata of the Ordovician comprise a relatively thin succession of clastics – sandstones, argillites and clays of the Pakerort and Varangu stages, overlain by the glauconitic sand- and siltstones of the Hunneberg and Billingen stages. The transition from the terrigenous to carbonate rocks in the Billingen Stage is marked by the appearance of calcareous interbeds in the siltstones, which grade into the first limestone/dolomite unit, the Toila Formation. The appearance of the first representatives of the numerous characteristic Middle Ordovician fossil groups is recorded in the same transition interval or in the overlying Volkhov Stage. The Ordovician limestone succession in Estonia and adjacent areas begins with cold-water carbonates deposited in a sediment starving shallow marine basin. Upward the sedimentation rates have increased. The corals make their first appearance in the Upper Ordovician, and the first carbonate buildups can be recorded, emphasising a striking change in the overall character of the palaeobasin. Generally the change in the type of sedimentation and in the character of biofacies is ascribed to a gradual climatic change resulting from the northward drift of the Baltica Palaeocontinent from the temperate climatic zone to the (sub)tropical realm (Nestor & Einasto 1997).

The details, but also the problems of the Ordovician geology in the subsurface area, in central and southern Estonia, were first revealed only in the 1950s. A high number of drill cores, obtained in the course of an extensive drilling programme in the 1950s–1980s, revealed a marked difference between the stratigraphic sucessions in the outcrop area and southern Estonia. As a result of the comparison of the eastern Baltic and Scandinavian successions, the concepts of the structural-facies zones (by Männil 1966) or confacies belts (by Jaanusson 1976; see Fig. 2) were introduced for the Ordovician of Baltoscandia. The micropalaeontological and macrofaunal studies of the core sections also revealed the distinctive biogeographic differentiation pattern, characteristic of the Ordovician rocks (Männil 1966; Männil *et al*. 1968; Meidla 1996, etc.). Although the biofacies pattern is generally described for the eastern Baltic area, the facies zonation of the entire Baltoscandian area is still imperfectly known. The seismic investigations of the Baltic Sea area, performed in the last decades (Tuuling 1998 and references therein), but also detailed (micro)palaeontological investigations (e.g. Tinn 2002) might produce valuable new information in this field.

The total thickness of the Ordovician varies from 70 to 180 m, being maximal in central and eastern Estonia and considerably less in the outcrop area.

Several correlation problems still persist in the Ordovician of Estonia, due to marked biofacies differences between northern and southern Estonia. In

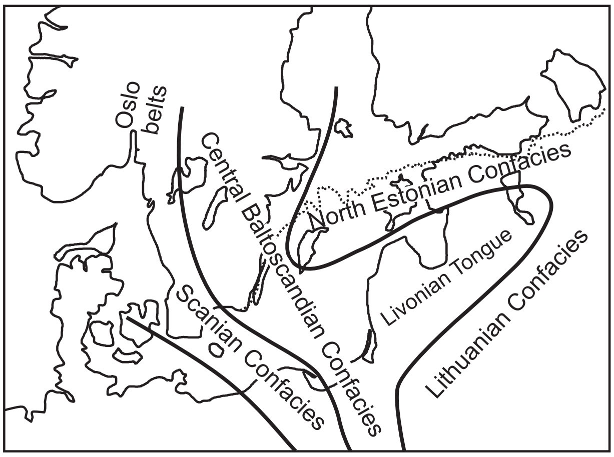

Fig. 2. Post-Tremadoc Ordovician confacies belts in Baltoscandia as defined by Jaanusson (1976).

part, they are discussed also in a recent monographic overview of Estonian geology (see Heinsalu & Viira 1997, Meidla 1997, Hints 1997, Hints & Meidla 1997 in Raukas & Teedumäe 1997). New prospects in this field have already been opened by stable isotope studies, as the stable carbon isotope curves have demonstrated a good correlation potential (Kaljo *et al*. 1999, 2001, and references therein; Ainsaar *et al*. in press; see also the papers of the same authors and Meidla *et al*. in the present volume).

The development of the stratigraphic classification of the Ordovician strata in Estonia, from the "beds" (*Schichten*) by Schmidt (1858) to the stages in modern meaning is documented in detail in Männil (1966), Rõõmusoks (1983) and Rõõmusoks *et al*. (1997). The term "Ordovician" was introduced for Estonia by Bassler (1911). A number of regional series and subseries for the Ordovician System in Estonia and neighbouring Russia were introduced by Schmidt (1881) and several subsequent authors. Raymond (1916) introduced the traditional American three-fold subdivision of the Ordovician System for this particular area, but this classification was subjected to repeated changes until 1987. Also the terms "Oeland Series", "Viru Series" and "Harju Series" have been widely used as a basic classification for the Ordovician System of the area since the 1950s (introduced by Kaljo *et al*. 1958 and Jaanusson 1960 in a nearly recent meaning). The subseries have partly been introduced as well (see Männil & Meidla 1994 for a summary), but they are rarely used today. The modern three-fold classification of the Ordovician System (IUGS 2004) was first used for the Estonian succession by Webby (1998) and is presented here in detail (Fig. 3). Some problems of the correlation of the Estonian units with the global scale will be discussed in the present meeting as well (see the papers by Puura & Viira and by Nõlvak in the current volume).

In relation to the definition of the GSSP for the base of the Ordovician System in the Green Point section, Newfoundland (Remane 2003), a revision of the traditional position of the Cambrian–Ordovician boundary at the base of the Pakerort Stage in Estonia turned out to be necessary. According to conodont data, the system boundary in the northern Estonian sections lies some metres higher than previously suggested, i.e. in the middle of the Pakerort Stage, within the Kallavere Formation (Puura & Viira 1999 and in this volume).

The term "Stage", first applied by Bekker (1921), has become the principal category in the chronostratigraphic classification of the Ordovician System in Estonia. Main features of the chronostratigraphic classification of the Ordovician System were established already by Männil (1966). Only minor changes were introduced in the later decades: the Ceratopyge Stage was renamed the Varangu Stage (Männil 1990), the Latorp Stage was replaced by the Hunneberg and Billingen stages (Hints *et al*. 1993) and a new unit, the Haljala Stage, is used instead of the Idavere and Jõhvi Stages (following Jaanusson 1995 and Nõlvak 1997). Hints & Nõlvak (1999) brought the concept of boundary stratotypes ("golden spike") into the Estonian stratigraphy, proposing a stratotype – the Pääsküla outcrop – for the lower boundary of the Keila Stage. However, as stratigraphic hiatuses on the stage boundaries are very common in northern Estonia (the best faunal changes are usually related to hiatuses), wide usage of this concept for the stage boundaries in this area looks rather complicated.

The lithostratigraphic classification of the Ordovician rocks was introduced by Orviku (1940) for the upper Middle Ordovician. This approach was widely accepted by subsequent authors and led to compilation of a series of detailed correlation charts approved by the Interdepartmental Stratigraphic Committee of the former USSR (Resheniya... 1965, 1978, 1987 and a related paper by Männil & Rõõmusoks 1984). The last version of such a formal correlation chart (the edition of 1987) was, in a slightly emended form, published also in English, in the series of the IUGS publications (Männil & Meidla 1994). The correlation chart in Fig. 3 contains only minor improvements compared to this publication, the most recent ones being introduced in Ainsaar & Meidla (2001). Some

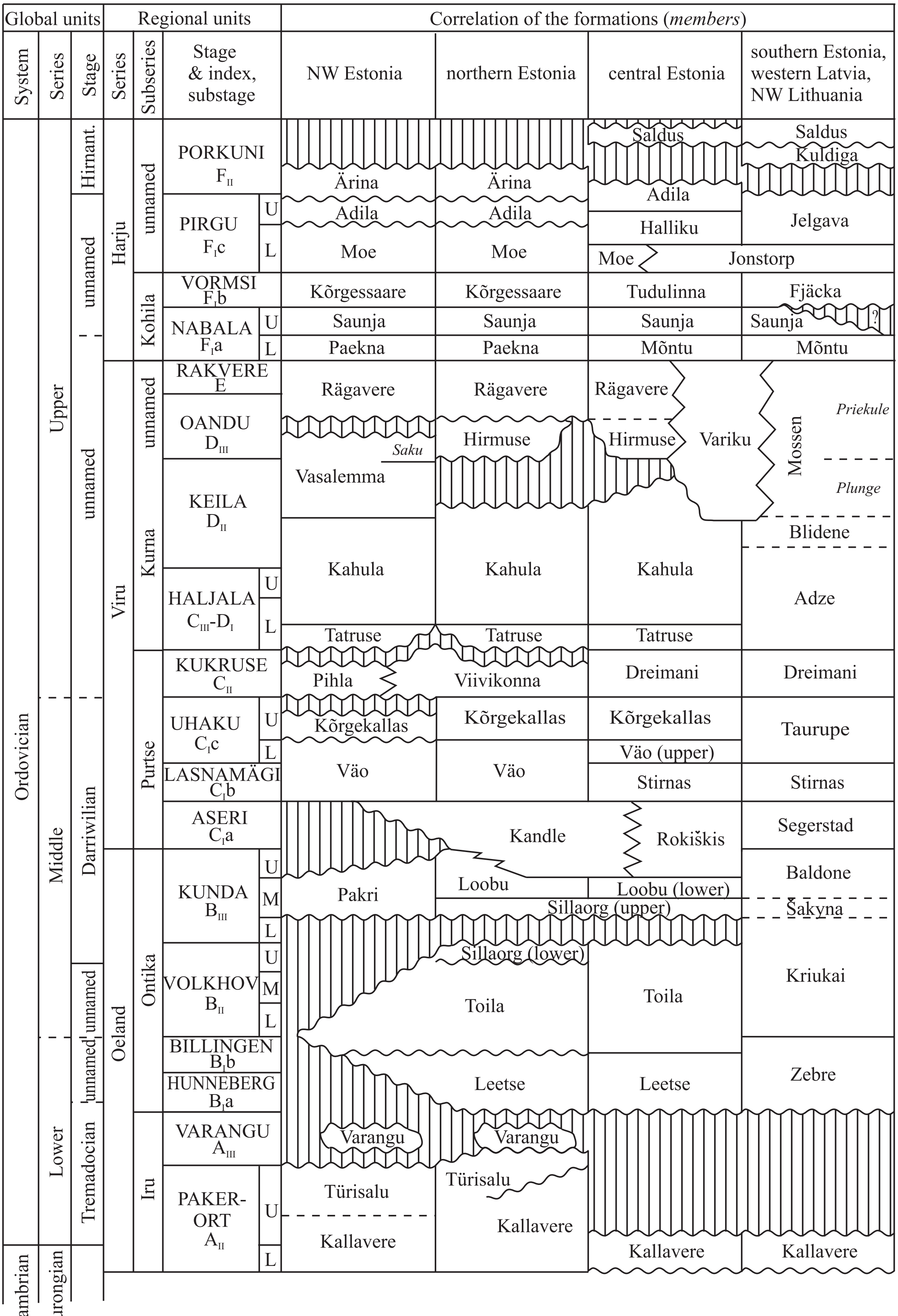

Fig. 3. The Ordovician of Estonia.

more modifications of the Ordovician correlation charts for Estonia have been published by Hints et al. (1993) and Nõlvak (1997). The composition and textures of the Ordovician carbonate rocks and the principal differences between the confacies belts were summarised by Põlma (1982 and references therein).

The monographic studies on the Ordovician palaeontology started already in the 19th century. After the comprehensive review on the Ordovician and Silurian strata (in modern meaning) by Schmidt (1858 and several subsequent monographic papers), a number of important monographic papers were published by F. B. Rosen, W. Dybowski, A. Pahlen, G. Holm, A. Mickwitz, O. Jaeckel, J. H. Bonnema and R. F. Bassler. The tradition of palaeontological investigations on the Ordovician material of Estonia was continued by A. Öpik (1930, 1934 and others) and, later on, by the recent generation of palaeontologists. Monographs and extensive monographic papers were published on the Ordovician brachiopods, corals, stromatoporoids, chitinozoans, scolecodonts, ostracods, conodonts, etc. Summaries on the palaeontological investiations on virtually all fossil groups recorded from the Ordovician of Estonia are published in the recent monograph *Geology and mineral resources of Estonia* (Raukas & Teedumäe 1997).

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A glance at the history and geology of Tallinn

Jaak Nõlvak

Based on the historical background, relief, and size and location of buildings, Old Tallinn is divided into two main parts: the fortress of Toompea (*Dome Hill*) and the lower town. Toompea, the oldest part of the town on a high limestone elevation rises like an island dozens of metres above the surrounding quarters, dominating even today among new tall buildings. According to an old Estonian epic, Toompea is the tomb of Kalev, the ruler of Old Estonians, heaped up of stones by his widow Linda.

Some historical remarks

Ancient Estonians built their probably wooden stronghold named *Lyndanise*, later known as the Danish *Castrum Danorum*, or the German *Ordensburg*, on the top of the elevation, which offered natural protection. It is not known whether the stronghold was inhabited all the year round or only during the navigation period and in case of war. It also served as a guarantee for the port's development in its site. The harbour was supposedly in the estuary of the Härjapea River, nowadays closed in underground pipes, from where the seamen could quite easily get fresh water from Lake Ülemiste. In the course of time the port was shifted seaward because of the constant withdrawal of the sea. At the present time, the area is rising at a rate of 2–2.5 mm per year.

The lower town (*suburbium, Unterstadt*), known (from 1280) as the Hanseatic town Reval, was a trade centre, populated by craftsmen, merchants and lower-class people depending on them throughout the Middle Ages.

The first written record of Tallinn dates from 1154, when Al-Idrisi, an Arab geographer, entered it in his map of the world. Due to its position at the crossroads of the land and sea routes, the stronghold of Toompea turned into a major centre of northern Estonia in the turn of the 10th and 11th centuries.

In June 1219, in the course of the crusade, proclaimed to christianise and subordinate Estonia, the stronghold was conquered by Danes, who pulled it down and started to build their own. The first stone stronghold was completed about ten years later in 1229 by the German Order of the Brothers of the Sward who had subjugated Danes. The material for the construction was taken from the first Ordovician limestone quarry, which is now under the Toompea Cathedral. The quarry seems to have existed only for a short time, because there are data saying that as early as 1233 the construction of the cathedral at the same place was under way.

In the 13th century the fortress of Toompea was thoroughly rebuilt and the cathedral has maintained its Late Gothic countenance. At the same time, the development of the lower town was also active. The population consisted of Germans and Estonians, but also Swedes and Finns, and a smaller number of Russians and other nations. Important were the privileges granted by the Lubeck Law, which became the main legal act adopted in *civitas Revaliensis* between 1238 and 1248. In 1265 the order to secure the city with a wall was given. The most noteworthy buildings of that period include the monasteries (St. Catherine, started in 1246 by Dominicans, also Cistercians, etc.), the Town Hall (finished just 600 years ago (!), in 1404), the town wall with more than 45 (26 preserved) defence- and gate-towers, guild's houses, basilicas (Niguliste, first mentioned in 1308, and Oleviste in 1267), numerous dwelling houses with a specific layout and other stone buildings.

Later the main structure of the old town was preserved, however, adapted to new tasks and changes in the mode of life and taste. All buildings were reconstructed, especially during the Late Gothic period, i.e. in the 15th century and at the beginning of the 16th century. The advancing war technique impelled improvement of the Toompea stronghold and modernisation of the town wall, which was repeatedly reinforced and made higher, up to 14 m. For all these works building material was needed, and a magnificent material – very specific and unique light grey limestone for blocks, plates and lime was, and is still available.

Some geological remarks

In the Tallinn area and in the whole of northern Estonia, the Proterozoic crystalline basement is covered with Upper Proterozoic and Lower Palaeozoic sedimentary rocks, up to the lowermost Upper Ordovician (some 456 million years ago) near the North Estonian Klint. The upper boundary of the basement is at a depth of 130–150 m b.s.l. The sedimentary rock layer forming the upper surface of the basement has a gentle southward inclination, on average

3–4 m per km. On the basis of the lithology of sedimentary rocks, the Lower Palaeozoic in Tallinn can be divided into two main parts: (1) the uppermost Precambrian (Vendian), Lower and Upper (Forungian) Cambrian and Lower Ordovician (Tremadocian), composed mostly of terrigenous rocks, and (2) the Ordovician (upper Lower–lower Upper, = from Arenig to lowermost Caradoc), formed mainly of carbonate and fine-terrigenous rocks.

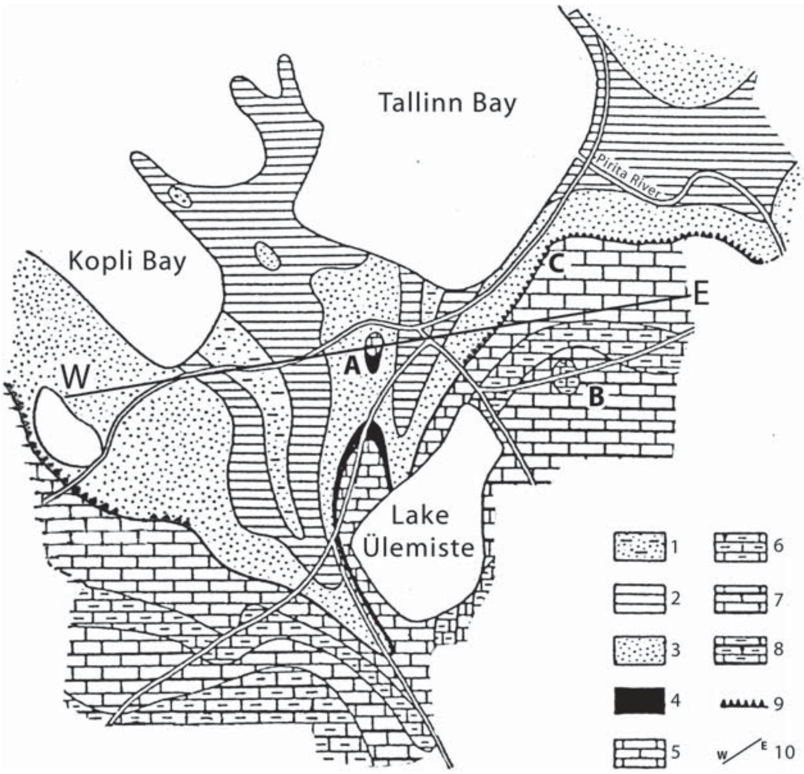

Fig. 1. Geological map of Tallinn (modified from Müürisepp 1976). **1** – Lower Cambrian sandstones, **2** – Lower Cambrian clays, **3** – Upper Cambrian-Lower Ordovician sandstones, **4** – Lower Ordovician argillites, **5** – Lower-Middle Ordovician limestones, **6** – Middle Ordovician limestones, **7** – Middle-Upper Ordovician limestones with kukersite, **8** – Upper Ordovician argillaceous limestones with marls, **9** – North Estonian (Baltic) Klint, **10** – profile (Fig. 2). **A** – erosional island of Toompea, **B** – Sõjamägi, **C** – Suhkrumägi.

Lower Cambrian sandstones and clays crop out at Rocca-al-Mare and in the valley of the Pirita River. The clays of the Lontova and Lükati formations are of economic significance. These so-called blue clays crop out on the Kopli Peninsula and served for centuries as raw material in the manufacture of rough ceramic products such as roof-tiles, bricks, etc., and cement. Nowadays the quarries are abandoned.

A hiatus corresponds to the Middle Cambrian. Yellow sandstones, typical of the Upper Cambrian occur in the lowermost Ordovician as well. The section continues with dark brown kerogenous argillites, which can be observed also in the lowermost parts of the Toompea sections (e.g. near Nunne Street). These argillites pose a serious threat to the environment because they tend to self-ignite when disposed in waste dumps and contain some radioactive elements. This was the situation in the Maardu area (east of town) before the open-cast pits for phosphorite mining were closed in 1991. In the limits of the town the layer with phosphorite-bearing valves of fossils (lingulate brachiopods) and fragments rich in P_2O_5 (up to 35%) is absent (e.g. at Toompea), or some lenses reach about 10 cm in size (e.g. at Mäekalda). The most representative layer (0.7 m) can be observed in the Iru section. Oil shale (kukersite), one of the main mineral resources in North-East Estonia, is represented in Tallinn sections only by very thin layers (some cm) related to the limestones of the Kukruse Stage, which are observable in some temporary cavings near Sõjamägi, but not in the klint sections.

Within the town, the bedrock topography is characterised by several relatively small positive (Toompea, Sõjamägi) and nowadays not visible negative features – three up to 140 m deep valleys fulfilled with Quaternary sediments (Figs 1, 2). The bedrock topography is of complicated genesis due to the contribution of different relief-forming forces. The most important factor is the heterogeneous lithological composition, i.e. the unequal resistance of the rocks to denudational processes.

A steep escarpment, the North Estonian Klint (part of the Baltic Klint), is the most notable bedrock relief

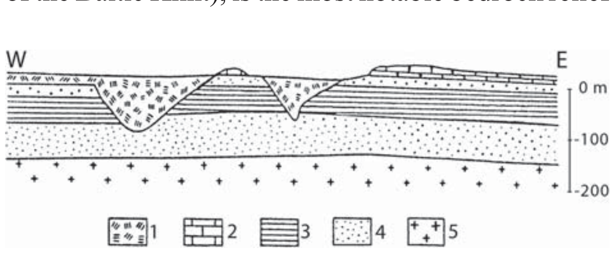

Fig. 2. Geological section (W-E) of the Tallinn area. **1** – Quaternary sediments, **2** – limestones, **3** – clays, **4** – sandstones, **5** – crystalline basement.

form. The development of the klint started, in all likelihood, during the Cenozoic. It was formed by denudation during a prolonged continental period. Later on, it was transformed by glacial erosion and marine abrasion according to the hardness of the rocks (Figs 1, 2). The klint emerged from the waters of the Baltic Sea during the Yoldia regression, some 10 300–9 500 BP. The lower areas of the present-day town became dry land later according to how the sea withdrew. The klint consists of rocks with unequal resistance. The section at Suhkrumägi displays pure hard limestones and dolomites in the topmost 6.5 m, which are underlain by soft Upper Cambrian–Lower Ordovician terrigenous rocks, mainly sandstones.

Higher in the section, among much harder Middle Ordovician carbonate rocks there are limestones, some 462 million years in age, known as Building Limestone. Of the numerous quarries used at different times in the limits of the town, the most significant ones were situated in the Lasnamägi area. From times when work was done mainly by hand, 58 limestone layers have their specific names, reflecting the quality of a certain rock layer and its suitability for different purposes, e.g. for making walls, stairs, floor plates, facing panels or for lime production. There were also "bad" layers – more argillaceous intercalations, which were unfit for building, but could be used as road metal.

The topmost layers on Toompea Hill (480 x 220 m; 47.5 m a.s.l.) belong to the lowermost Uhaku Stage (see Fig. 1, point A), which is a little higher stratigraphical level than the topmost limestone beds at Suhkrumägi (Fig. 1, point C), where the klint is the highest in the limits of the town, rising up to 47 m a.s.l. In many places, at the foot of the klint, there are terraces of different stages of postglacial seas and therefore the klint is partly covered, not occurring always as an abrupt escarpment.

To the south of the klint the bedrock relief continues as a limestone plain (*alvar*) with a very thin Quaternary cover, or without it, reworked also by Quaternary glaciers and waters of the Baltic Sea and glacial lakes. Such plains are characteristic of the northern Estonian landscape as a whole.

Lake Ülemiste on the outskirts of Tallinn provides the town with drinking water. According to an old Estonian epic, the lake was formed from the bitter tears of Linda, when she was crying over old Kalev's death. Actually, the lake isolated from the Yoldia Sea ca 9500 BP.

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Stop 1. Pääsküla Hillock

Olle Hints

Pääsküla Hillock is situated at the southern boundary of Tallinn, close to the Tallinn–Pärnu road. A 3 km long, 2.4 km wide and 18 m high bedrock elevation in the discussed area is in most part occupied by the outcrop of the Keila Stage. The Quaternary cover is seldom thicker than 0.3 m on the hillock and the area is rich in bedrock exposures.

Most of the outcrops have been created artificially during the construction of Russian defensive positions between 1912 and 1918. The Pääsküla position was part of a huge system of defensive positions known as the naval fortress of Peter the Great, the primary aim of which was protecting St. Petersburg. The whole system, however, was never completed and none of the cannons fired in battle. The Pääsküla position consisted of six concrete shelters built into the bedrock, which were connected with subsurface tunnels and some other constructions. Some time after World War II, the major part of the hillock was occupied by a Soviet military base and was closed for geological research until the 1990s.

The argillaceous limestones of the Kahula Formation are exposed in various trenches, excavations, shelters and subsurface tunnels in a total thickness of over 10 m. Several of the sections in Pääsküla Hillock were described by R. Männil in the 1950s (published by Rõõmusoks 1970, pp. 257–260; see Fig. 1 herein). The limestones are rich in fossils that have been collected, especially in the 1930s–1940s. Common macrofossils include *Ristnacrinus* sp., *Cyclocrinites*, *Sowerbyella trivia*, *Porambonites* sp., *Dalmanella kegelensis*, *Asaphus nieszkowskii*, *Clinambon anomalus* and many others (see Rõõmusoks 1970 for more details).

One particular excavation in Pääsküla Hillock constituting the stratotype for the Laagri Substage was visited during the Third Baltic Stratigraphical Conference in 1996 (see description in Nõlvak 1996). The Pääsküla Member has been named after the Pääsküla locality. However, according to the present-day concept of this lithological unit, it is actually not cropping out in the hillock (see discussion in Põlma *et al*. 1988).

The composite section embodies two volcanic beds (K-bentonites, metabentonites) — the **Kinnekulle Bed** (previously known also as “thick bed”, “bed d”, “main metabentonite” and “bed BBB”) and the **Grimstorp Bed** (see Bergström *et al*. 1995 for an overview of these names, distribution patterns, references, etc.).

The subsurface tunnels, more than 2 km long in Pääsküla Hillock, are the only outcrops of the Kinnekulle Bed in the East Baltic region. The lithological properties and mineralogical composition of the Kinnekulle Bed in Pääsküla are discussed by Hints *et al*. (1997). In brief, the Kinnekulle Bed is some 25–30 cm thick and consists of 0–1 cm of hard rock variety in the lower part, 15–18 cm of light yellowish-grey clay in the middle part and 8–11 cm of yellowish hard rock with burrows in its upper part (Fig. 2). In some places, the upper hard part is thicker, making up nearly the entire thickness of the bed. The lower boundary of the bed is sharp; the underlying limestones are often impregnated with pyrite. The contact with the overlying limestones is, on the contrary, more or less transitional, making it difficult to measure the total thickness of the bed. Mineralogical analysis revealed potassium feldspar and illite-smectite in varying proportions as the main minerals. The maximum content of potassium feldspar reaches to about 80–90% and the maximum K_2O content determined is about 14% (Hints *et al*. 1997).

As the base of the Kinnekulle Bed has long been used to trace the lower boundary of the Keila Stage, a proposal was made by Hints & Nõlvak (1999) to

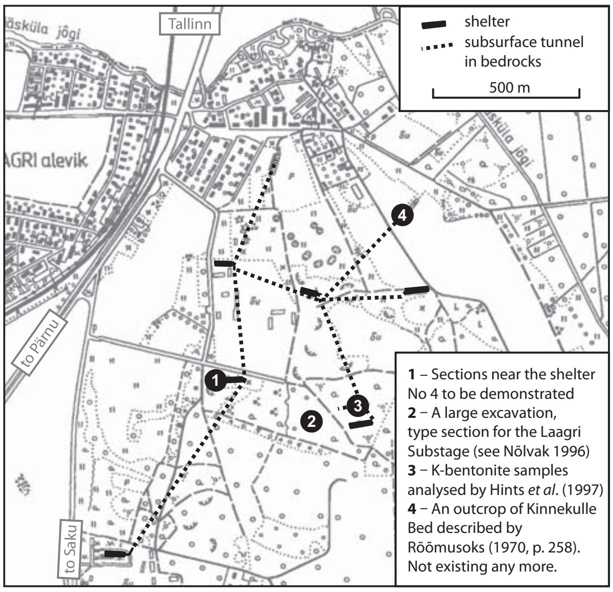

Fig. 1. Locality map.

fix the stratotype section and point for the Keila Stage in these tunnels.

Recent studies on microfossils in the Pääsküla composite section (Hints *et al*. 2003) have shown that the biotic effects of the Kinnekulle ash-fall can be most clearly recorded in ostracod assemblages. Just above the K-bentonite, the frequency and diversity of ostracods drop dramatically, the community gets reorganised and several species prevailing below the K-bentonite (e.g., *Tetrada memorabilis* and *Cytherellina* cf. *jonesi*) disappear. For some other groups the effect has not been that harsh, and only some quantitative changes in the assemblages can be noted. It is intriguing though that the Kinnekulle Bed itself is in places (not in every section studied) rich in microfossils, scolecodonts in particular (see Hints *et al*. 1997).

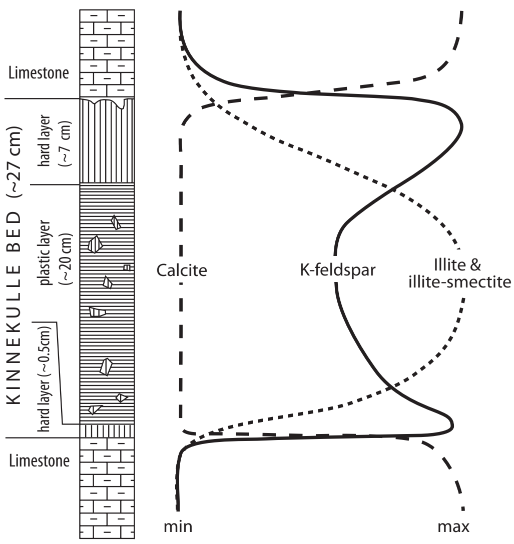

Fig. 2. Composition of the Kinnekulle K-bentonite in Pääsküla Hillock sections (after Hints *et al*. 1997).

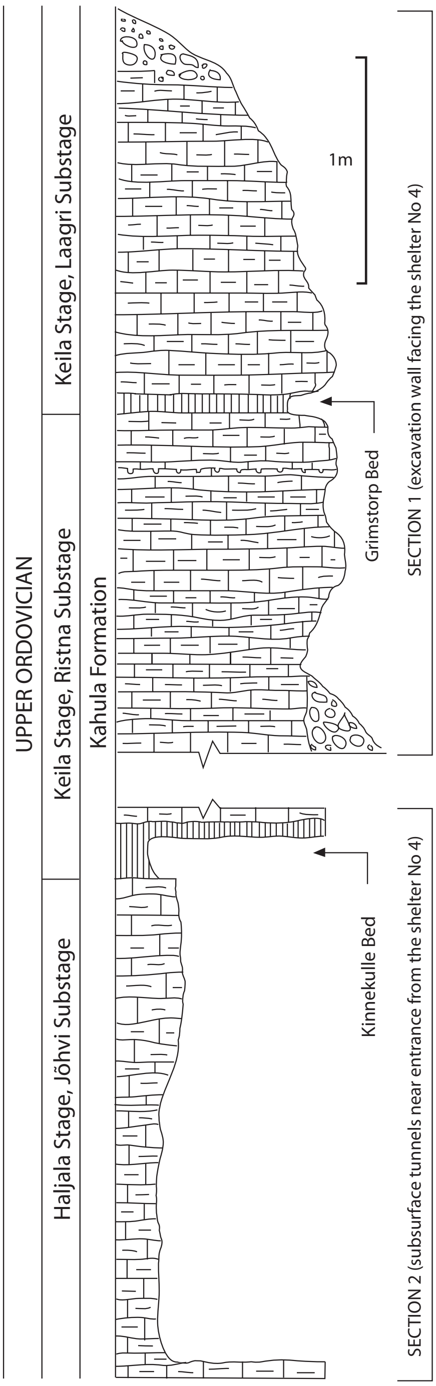

Fig. 3. Pääsküla Hillock sections.

Two separate sections will be demonstrated during the WOGOGOB-2004 excursion (see Fig. 3): **Section 1** is the wall of an excavation facing shelter No. 4 and exposing a ca 3 m succession of the Keila Stage. Also, in the middle part of the section, the Grimstorp K-bentonite can be observed.

Section 2 exposing the Jõhvian limestones and the Kinnekulle K-bentonite will be observed in a subsurface tunnel. The entrance into the tunnel goes through the above-mentioned shelter No. 4.

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Stop 2. North Estonian Klint on the Pakri Peninsula

Ivar Puura

The North Estonian Klint is the central part of a larger structure, the 1200 km long Baltic Klint. Some authors have regarded the klint as a tectonic fault terrace, but, considering the recent geological evidence including drilling data and undersea studies, it is rather a sub-aerial denudation or an abrasion terrace. Opinions vary with regard to the age of the formation of the klint: Mid-Devonian to Miocene–Pliocene. The age of the rocks exposed in the klint wall ranges from the Lower Cambrian to the Middle Ordovician. The Estonian national committee of the UNESCO has accepted the Baltic Klint as a candidate for a geological heritage site.

In the context of a study of the Baltic Klint as a nature monument, Suuroja (2003) has suggested names for klint regions, and his classification is followed here. In the west, the Baltic Klint begins under the sea tens of kilometres south of the island of Öland, continues along its western coast (the 150 km long Öland Klint), then proceeds under the sea and appears on the Estonian islands of Osmussaar, Suur-Pakri and Väike-Pakri (the 500 km long klint in the Baltic Sea). It continues on the mainland of Estonia at or near the south coast of the Gulf of Finland, from the Pakri Peninsula in the west to the Narva River in the east (the 300 km long North Estonian Klint). East of the Narva River, in the Leningrad district of Russia, known as Ingria or Ingermanland, the klint extends to Lake Ladoga and is exposed in river valleys and as a series of inland terraces (the 250 km long Ingermanland Klint, also called the Ladoga Klint). The highest point of the Ingermanland Klint and the Baltic Klint plateau (140 m above sea level) is located in the environs of Koporye. In Estonia, the klint is the highest in the Ontika–Valaste area, another klint locality described in this excursion guide by O. Tinn.

The Cambrian and Lower to Middle Ordovician rocks exposed in the sections along the North Estonian Klint and in the river valleys cut into the klint plateau have attracted the geologists' interest since the early 19th century and their study has been significant for the development of the Ordovician stratigraphy of Baltoscandia.

The Pakri Peninsula

The klint section on the Pakri Peninsula has been studied by geologists over two centuries (see Rõõmusoks *et al.* (1997) for the historical review). Intensive modern biostratigraphic studies on the Pakri Peninsula began in August 1993, when the former military area became accessible for visitors. Since 1998, the coast of the Pakri Peninsula has been protected as a national landscape reserve, with temporary restrictions on access to some klint areas near the promontory of the peninsula, where the only nesting area of the black guillemot (*Cepphus grylle*) in Estonia is located.

Key sections

The composite section (Fig. 1) gives a general overview of the stratigraphic subdivision of the westernmost part of the North Estonian Klint. The promontory of the Pakri Peninsula points to the north-west, and its margins facing west and north-north-east are here referred to as western and northern margins of the peninsula. The cliff, extending along the western margin of the Pakri Peninsula from the lighthouse at the peak almost to the northern border of the Paldiski town, is like a textbook illustration of the slight southward dip (0.1–0.3 degrees or 3–4 m per km) typical of the Lower Palaeozoic rocks of Estonia lying on the southern slope of the Baltic Shield (Puura & Vaher 1997). Thus, the oldest rocks, including the Cambrian–Ordovician boundary beds, are exposed near the lighthouse at the promontory of the Pakri Peninsula and are submerged below the surface in the southward direction, where successively younger beds can be conveniently reached for investigation.

The Pakri Cape section (The section at the lighthouse – Raymond 1916; The Pakri Cape section – Mens *et al.* 1996; Pakri Peninsula, point A – Mens & Puura 1996). The section at the lighthouse described by Raymond (1916, p. 186, plate 2) as the "most instructive section" of his Packerort Formation, comprising "Ungulitensand" (A2) and "Dictyonemaschiefer" (A3) of Friedrich Schmidt, is now considered as the stratotype of the Pakerort Stage. The interval from the top of the Lower Cambrian Tiskre Formation to the top of the basal part of the Lower Ordovician Türisalu Formation is exposed and reachable for the study. The results of the biostratigraphic study of the section 1 km southeast of the promontory are available in Mens *et al.* (1996).

The Leetse section. The section near the Leetse village at the northern margin of the Pakri Peninsula, 5 km SEE of its promontory is the stratotype of the Leetse Formation, comprising the glauconite sandstones of the Klooga and Joa members and the glauconitic limestones of the Mäeküla Member. It has been a well-known locality since the 19th century, as is evident from publications and palaeontological collections.

The Uuga cliff section (Pakri Peninsula, point C – Mens & Puura 1996; Uuga Pank – Viira *et al*. 2004). In this section the clays of the Varangu Stage and the sandstones of the Hunneberg Stage can be conveniently reached for investigation. The results of the pilot studies of fossil distribution have been reported by Mens & Puura (1996) and Viira *et al*. (2004). The southernmost part of this section will be visited during the excursion. In the klint wall, the Lower to Middle Ordovician sequence from the Hunneberg Stage to the Uhaku Stage is exposed.

Description of the composite section (from base to top)

Lower Cambrian

Tiskre Formation

The exposed thickness above the surface in the Pakri Cape section 4 m. (The total thickness according to the Põllküla core F-317 is about 18 m). Light grey silty sandstone with intercalations of argillaceous siltstone and clay (Mens & Puura 1996).

Furongian (upper Cambrian)–Lower Ordovician

Pakerort Stage.

Thickness 8.5 m. In the Pakri Cape section, the Pakerort Stage is represented by the sandstone of the Kallavere Formation (thickness 4 m) and the overlying monotonous black shale (Türisalu Formation). The lower boundary of the Kallavere Formation varies along the klint. About 200–300 m east of the Pakri Cape section, it is marked by distinct lenses of basal conglomerate, incorporating pebbles of different ages and sources and even boulders from the underlying Tiskre Formation (Nemliher & Puura 1996). Acritarchs from the interior and of the pebbles (Mens *et al.* 1999) indicate different levels in the upper Cambrian as the sources. The lenses yield fragments of lingulate brachiopods *Ungula ingrica* (Eichwald).

In the Pakri Cape section, the basal beds yield conodonts of the *Cordylodus proavus* Biozone, as does the 2.75 m thick basal interval of intercalating sandstone and kerogenous shale beds corresponding to the Maardu Member. No conodonts have been found from the overlying 1.25 m thick interval of the sandstones of the Suurjõgi Member, although it yields reworked tiny shell fragments of lingulate brachiopods. From the basal part of the overlying dark brown kerogenous argillite of the Türisalu Formation, *Rhabdinopora flabelliformis flabelliformis* and *R.* cf. *R. flabelliformis desmograptoides* have been found (Mens *et al.* 1996).

The base of the Ordovician. The base of the *Cordylodus lindstromi* Biozone is the closest traceable level in Baltoscandia approximating the base of the Ordovician as defined by the first appearance of *Iapetognathus fluctivagus* in the GSSP in Green Point, Newfoundland (Puura & Viira 1999, in this volume). Rare specimens of *Iapetognathus* sp. recorded in some Estonian sections have been suggested as an additional criterion (Heinsalu *et al.* 2003). Another additional criterion is the first appearance of graptolites *Rhabdinopora* ex gr. *flabelliformis* (Puura & Viira in this volume).

In the Pakri Cape section, the graptolites indicate that the monotonous black shale of the Türisalu Formation can be safely assigned to the Ordovician. The finds of conodonts from the underlying Suurjõgi Member could give information if part of the member can be assigned to the Ordovician. According to Viive Viira (personal communication 2004), in an outcrop some hundreds of metres south of the promontory at the western margin of the Pakri Peninsula, a sample from the basal 0.75 m of the Suurjõgi Member (0.5–1.25 m below the base of the monotonous black shale), has yielded *Cordylodus angulatus* and *Iapetognathus* sp. Thus, the base of the Ordovician is definitely not higher than 0.75 m from the base of the Suurjõgi Member. Considering the low resolution of sampling, the base of the Ordovician is tentatively drawn below the Suurjõgi Formation, because in all other sections, the Suurjõgi Member is biostratigraphically not older than the *C. angulatus* Biozone.

Lower Ordovician

Varangu Stage

Thickness 0.5 m. In the Uuga Cliff section, the Varangu Stage is represented by greenish-grey to beige clay and silty sandstone with glauconite grains that has yielded conodonts of the *Paltodus deltifer* Biozone (Viira in Mens & Puura 1996).

Hunneberg Stage

Thickness 3.9 m. In the Pakri Peninsula, the Hunneberg Stage reaches its maximum thickness. In the

Uuga Cliff section, the Hunneberg Stage is represented by greenish-grey fine-grained glauconitic sandstone, with intercalations of light grey clay assigned to the Klooga and Joa members of the Leetse Formation. It also contains a diverse lingulate brachiopod fauna characteristic of the *Thysanotos siluricus* Zone (Mens & Puura 1996).

Billingen Stage

Thickness 0.3 m. Greenish-grey glauconitic silty sandstone which is replaced upwards by calcareous silty sandstones and glauconitic packstone. The lower 20 cm is assigned to the Mäeküla Member and the upper 10 cm to the Päite Member. The lower part of the Billingen Stage contains conodonts of the *Prioniodus elegans* Biozone and the upper part conodonts of the *Oepikodus evae* Biozone (Viive Viira, personal communication 2004; Viira *et al.* 2004).

Middle Ordovician

As compared to the eastern sections of the klint (e.g., Valaste waterfall section, Tinn in this guide), especially in the environs of the Volkhov River in Russia, the Middle Ordovician sequence in the Pakri Peninsula is very condensed. These carbonates, mainly packstones and wackestones, have formed in temperate climate low sedimentation rate conditions on shallow to middle ramp settings (Nestor & Einasto 1997). The following brief description of the Middle Ordovician units in the Pakri Peninsula is given after Einasto & Mens (1996).

Volkhov Stage

Thickness 1.3 m. In the Uuga Cliff section, the Volkhov Stage is represented by light grey limestone, with intercalations of marls, with glauconite grains. The lower boundary is marked by a distinct discontinuity surface. The conodonts of the *Oepikodus evae* Biozone have been found from the lower part of the stage (Viira *et al.* in this volume).

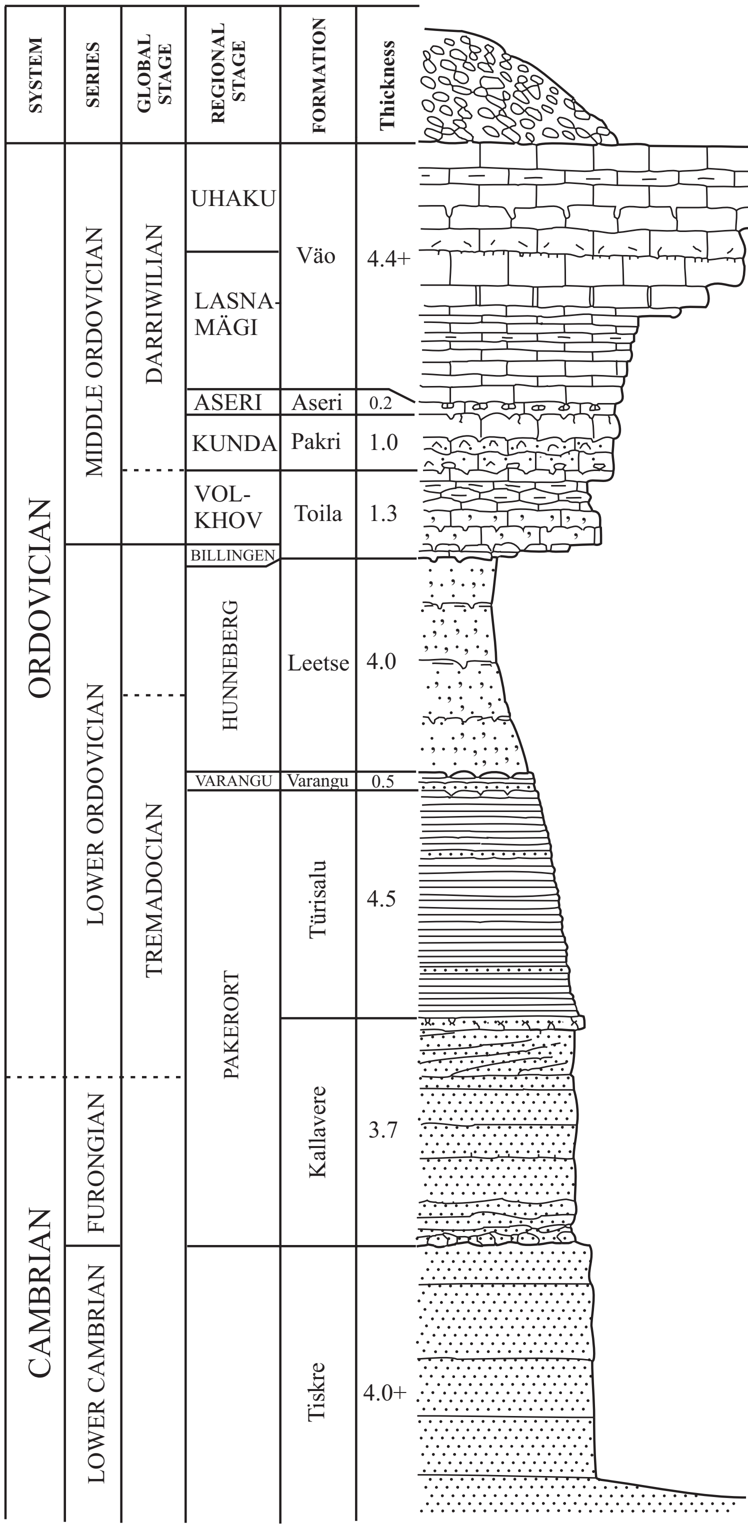

Fig. 1. The composite section of the klint in the Pakri Peninsula (after Mens & Puura 1996).

Kunda Stage

Thickness 1 m. The Kunda Stage is represented by the Pakri Formation in the peninsula. It is composed of kukersite-containing sandy limestone and limy sandstone. The rocks of the Pakri Formation are brecciated and penetrated by limy sandstone injections. This breccia (The Osmussaar Breccia), distributed in NW Estonia, has been explained as a result of a devastating earthquake (Suuroja *et al.* 2003). The lower boundary of the Pakri Formation is marked by a series of discontinuity surfaces, with rounded flat pebbles above them. The uppermost 0.1 m is represented by light grey fine-grained limestone with phosphatic discontinuity surfaces. The upper boundary is marked by an even discontinuity surface with deep vertical burrows.

Aseri Stage

Thickness 0–0.2 m. Argillaceous limestone with intercalations of marl and brown iron ooiids.

Lasnamägi Stage

Thickness 2.4 m. Brownish-grey dolomitic, thin- and medium-bedded limestone.

Uhaku Stage

Thickness about 2 m. Medium-bedded limestone with numerous discontinuity surfaces.

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Stop 3. Vasalemma quarry

Linda Hints, Peep Männik and Helje Pärnaste

In the vicinity of the Vasalemma settlement (first mentioned already in 1241), Ordovician limestones have been quarried since the 13th century. The limestone has been used as building material for the Padise monastery, Risti and Madise churches and Vasalemma and Laitse castles. At the beginning of the 20th century the Paldiski–Tallinn–Narva–Gatchina railway (opened in 1870) was connected with Haapsalu via a new section running through Vasalemma. The increasing need for limestone caused the opening of a new quarry close to the railway in 1920. Now there are five abandoned old quarries in the limits of the Vasalemma settlement. Some of these quarries are filled with water and are used as swimming places. In one quarry a rally tract has been built.

The quarry to be visited by our excursion is located east of Vasalemma. Here a small quarry was opened in 1931. From 1946, exploitation of the quarry increased considerably. Nowadays, the quarry is owned by the company Nordkalk Corporation, which is acting in Finland, Sweden, Poland and Estonia and is the leading manufacturer of limestone-based products in Northern Europe. The production of crushed limestone for building works is the main business of the Nordkalk Corporation in Vasalemma.

The size of the quarry is about 2x2 km (Fig. 1). The strata of the Kahula and Vasalemma formations are exposed in the quarry walls. The Kahula Formation and the main part of the Vasalemma Formation correspond to the Keila Stage, and the uppermost part of the latter formation exposed in the quarry belongs to the Oandu Stage. In the deepest places of the quarry the upper part of the Pääsküla Member (Kahula Formation) is exposed. Nowadays the mining is conducted about 5–6 m deeper than the previous quarry floor where the earlier studied drill cores Nos 36, 81, 812 and 816 are located (see Hints 1996, fig. A11). The excavating depth reaches the top or the uppermost part of the Pääsküla Member (Fig. 2). The upper surface of the Pääsküla Member is inclined to the south for about 4 m per kilometre (Fig. 2), which is close to the medium inclination of the Ordovician rocks in Estonia.

The quarries in the vicinity of Vasalemma have been visited by many excursions and descriptions of sequences are available at least in three earlier guides (Põlma & Hints 1984; Hints 1990, 1996). In the Vasalemma quarry the Ordovician section is exposed in a thickness of up to 10 m. The section of the uppermost Pääsküla Member can be studied only in the deepest part in the NW corner of the quarry (Fig. 1, locality 1). The measured thickness of the exposed part of the member is here about 1.3 m, but based on the data from cores drilled close to Vasalemma, the total thickness of the Pääsküla Member reaches up to 6–7 m in the region.

Lithologically, the Pääsküla Member consists of micritic pelletal seminodular limestones rich in aleuritic to very fine sand-size quartz (Fig. 3). Spotty distribution of hemispheric bryozoan colonies in life position and large valves of the strophomenid *Keilamena occidens* (Oraspõld) on some bedding planes are characteristic of the Saue Member. Also the trilobite *Conolichas* cf. *sjoegreni* Warburg can be found in the quarry. Some argillaceous bedding planes are covered by numerous valves of *Sowerbyella,* sometimes concentrated in large horizontal burrows. These burrows are presumably traces of

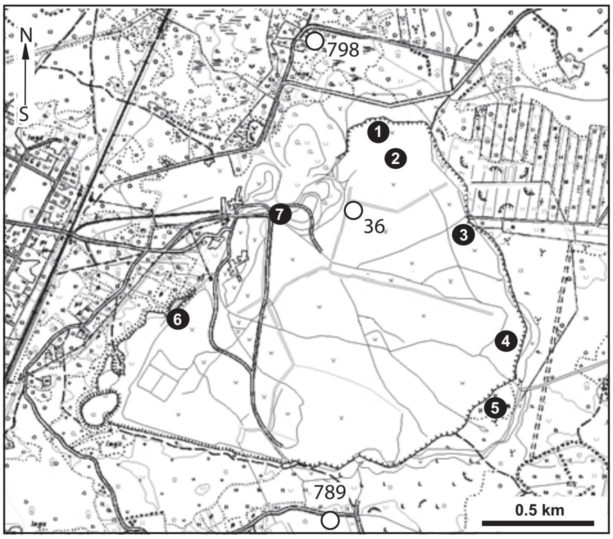

Fig. 1. Map of the Vasalemma quarry. Locality **1** – Section of the Pääsküla and Saue members of the Kahula Formation (see Fig. 2); **2** – quarry floor with ripple marks; **3** – northernmost carbonate buildups; **4** – reef-like structures with corals; **5** – Pleistocene ice scratches on the upper surface of the Vasalemma Formation; **6** – carbonate buildups in the bedded limestones on the quarry wall (Hints 1990, fig. 25); **7** – entrance to the quarry. For the sections of cores 798, 36 and 789 see Fig. 2.

some benthic animals, which were filled with bioclasts as a result of water activity. Also some other brachiopods, e.g. *Keilamena occidens* (Oraspõld), *Horderleyella kegelensis* (Alichova) and *Clinambon anomalus* (Schlotheim), are present in the Saue Member, the first two species occurring mainly as coquinas on some bedding planes.

At least one level with large ripple marks has been recorded in the Pääsküla Member (Fig. 3). The upper contact of the Pääsküla Member is also marked by large, impregnated and partly eroded ripple marks. In the northern part of the quarry the surface with ripple marks is exposed in a wide area (Fig. 1, locality 2). The dense set of burrows indicates active bioerosion by different organisms. The occurrence of *Trypanites*-type burrows refers to very shallow water conditions in the region at the end of Pääsküla time. The ripples are of N–S orientation and seem to be asymmetrical (for more details see Hints & Miidel in this volume). Southwards, close to the carbonate buildups, the ripple marks become indistinct, the quarry floor starts to undulate and in some places the corresponding bedding plane disappears under younger strata.

The boundary between the Pääsküla and Saue members becomes, according to the drill core data, less distinct north of the quarry. The transitional nature of this boundary and cyclic intercalation of argillaceous and less argillaceous limestones outside the distribution area of the Vasalemma Formation is described in Ainsaar (1992).

Most of the section in the northern part of the quarry (Fig. 1, locality 1) corresponds to the Saue Member of the Kahula Formation (Fig. 2). In the surrounding areas of the Vasalemma Formation the Saue Member is characterised by the occurrence of grainstone interlayers and lenses in more or less argillaceous limestones. The thickness of these interlayers and the amount of the bioclastic material in the limestone increase to the south. This can be followed in the eastern wall of the quarry (Fig. 1, from locality 1 to 3). The sequence in the southern part of this wall is mainly represented by bedded fine- to coarse-crystalline grainstones. Here, small carbonate buildups appear in the section.

In the lowermost part of the Saue Member the trilobites *Pseudobasilicus kegelensis* (Schmidt), *Asaphus* (*Neoasaphus*) *nieszkowskii* Schmidt and *Leiolichas illaenoides* (Nieszkowski) have been identified. The biodetrital limestones and bedding planes rich in fragments of different fossils have yielded, together with the above mentioned species, *Conolichas deflexus* (Angelin), *Atractopyge kutorgai* (Schmidt), *Atractopyge* aff. *errans* Öpik, *Keilapyge laevigata* (Schmidt) and *Chasmops marginatus* (Schmidt). The last two taxa and *Toxochasmops* (*Schmidtops*) *maximus* (Schmidt) and *Conolichas* cf. *sjoegreni* (Warburg) seem to be more characteristic of the argillaceous interlayers.

The transition from the Saue Member to the lower Vasalemma Formation has been studied also in several core sections (Fig. 2). Towards the south in the region, the argillaceous biodetrital limestones of the

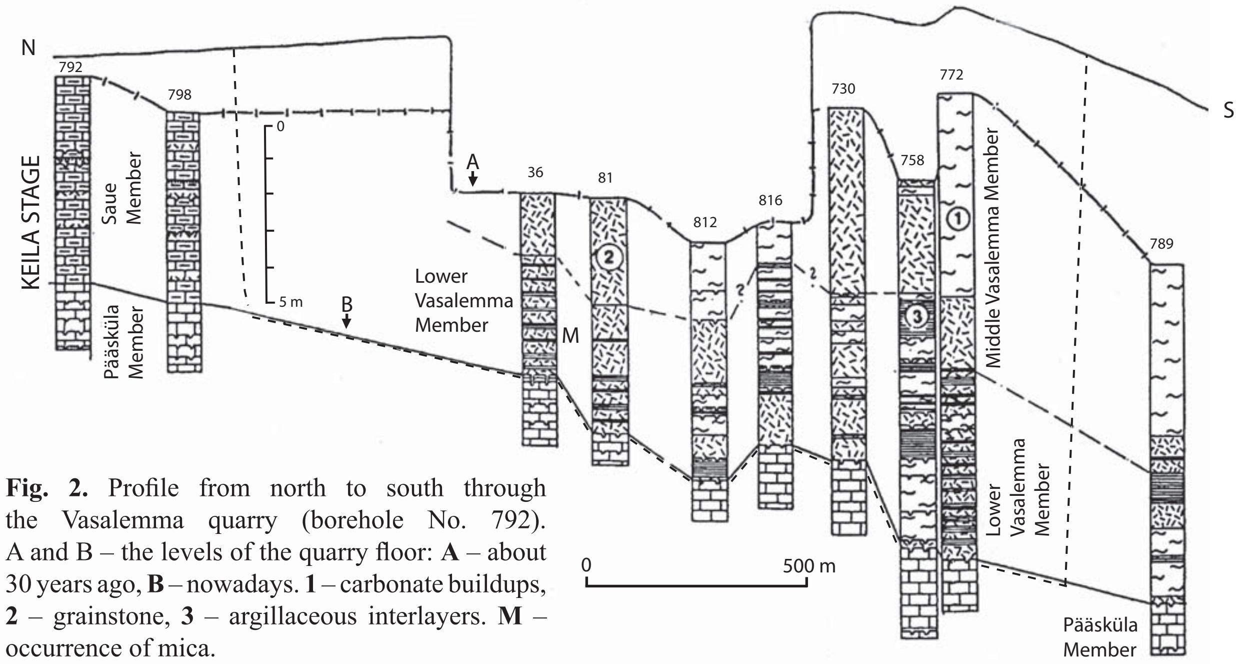

Fig. 2. Profile from north to south through the Vasalemma quarry (borehole No. 792). A and B – the levels of the quarry floor: **A** – about 30 years ago, **B** – nowadays. **1** – carbonate buildups, **2** – grainstone, **3** – argillaceous interlayers. **M** – occurrence of mica.

Saue Member (cores 792 and 798 in Fig. 2), with *Sowerbyella* coquinas at two or three levels (Fig. 3), are gradually replaced by grainstones in the lower part of the Vasalemma Formation containing also a bed with numerous *Sowerbyella*.

The Vasalemma Formation with its typical characteristic features is exposed in the middle and southern parts of the Vasalemma quarry (from locality 3 to 6 in Fig. 1). The formation consists of white to dark grey, fine- to coarse-crystalline bedded grainstones. The grainstones include distinctive irregular bodies of different size formed of massive pure limestones. The lower part of the formation contains an argillaceous interlayer, sometimes with fossils indicative of Keila age (for example the brachiopod *Estlandia*). The massive limestones, tentatively named here as carbonate buildups, have been interpreted as reefs (Raymond 1916), bioherms (Männil 1960; Rõõmusoks 1970) or mud mounds (Põlma & Hints 1984; Hints 1996). Younger buildups contain sometimes *Cyatocystis* (Edriasteroidea), which seems to form their frames. Also several other fossils are present,

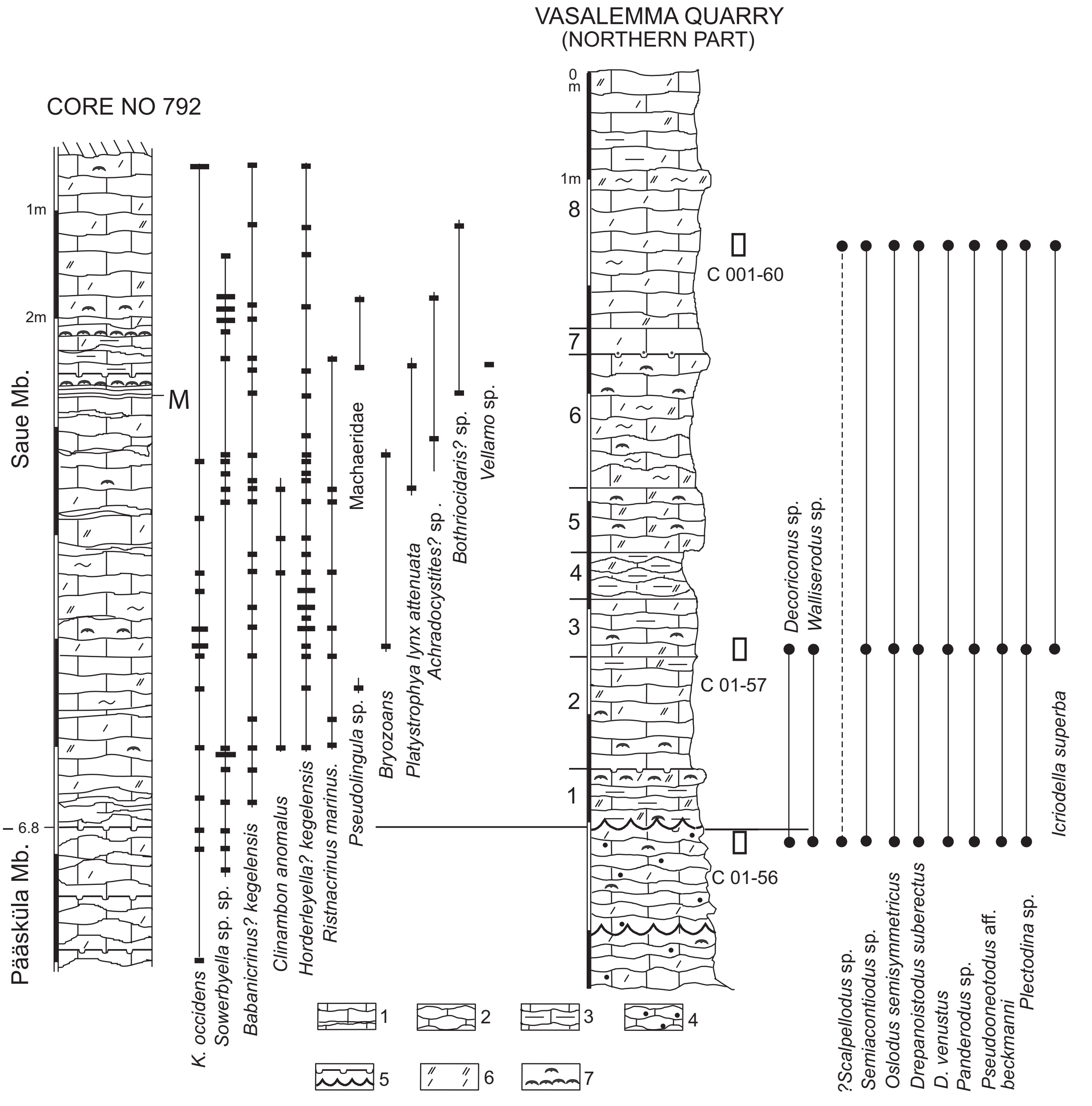

Fig. 3. Distribution of selected fossils in drill core No. 792 (located about 300 m north of borehole No. 789 in the upper part of Fig. 2) and conodonts (samples C on the right side of the log) in the northern quarry wall section. **1** – limestone (above) and argillaceous interlayer in limestone (below); **2** – nodular limestone; **3** – argillaceous limestone; **4** – pelletal limestone; **5** – discontinuity surface (above) and occurrence of mica (below); **6** – skeletal fragments (biodetritus) (above) and fossils, mainly brachiopods and burrows (below).

such as sponges, corals, stromatoporoid-like structures, the alga *Solenopora,* in some "caves" cephalopods, etc., which all together form the reef-like carbonate bodies. The Oandu age of the uppermost Vasalemma Formation in the quarry indicates the appearance of the oldest corals in Estonia (*Lyopora tulaensis* (Sokolov), *Eofetcheria orvikui* Sokolov) and brachiopods (*Rhynchotrema parva* Oraspõld and *Rostritsellula nobilis* (Oraspõld)), common in the Oandu marls (Rõõmusoks 1970).

The carbonate buildups contain a relatively low-diversity association of trilobites. The illaenid *Stenopareia ava* (Holm) is the most common, forming small coquinas (e.g. one about half a metre in diameter) in pockets on the outer surfaces of the buildups. The illaenimorph lichid *Leiolichas illaenoides* (Nieszkowski) and the cheirurid *Paraceraurus* sp. are also represented by a few findings in this association.

Conodonts were studied in three samples from the northern part of the quarry (Fig. 1, locality 1). In all samples the fauna is strongly dominated by simple-cone taxa. The most common taxon is *Semiacontiodus*, which accounts for up to 50% (or even more) of the total number of specimens in samples C 01-57 and C 01-60 (Fig. 3). *Drepanoistodus* and *Oslodus* are also well represented. Other taxa are less common. Ramiform taxa are extremely rare. Only two specimens of *Icriodella* and *Plectodina* have been found. It is also noteworthy that despite the relatively large size of the samples processed (samples C 01-57 and C 01-60 both weighed more than 15 kg), not a single specimen of *Amorphognathus* was found.

In the absence of *Amorphognathus*, extremely rare occurrence of other ramiform taxa and dominance of simple-cone taxa the conodont fauna is similar to that recognised in the Mid-Caradoc Event part of the Ruhnu (500) core section, in the interval from upper Haljala to lower Oandu stages (Männik 2003).

In the older literature the Vasalemma Formation has been described as *Hemicosmites*-limestone. In the Vasalemma quarry visited by the excursion, *Hemicosmites* has been found in the argillaceous limestones surrounding the carbonate buildups. However, in the Rummu quarry, located to the west of Vasalemma, the limestone consists mainly of the polygonal plates of the *Hemicosmistes* calyx (Fig. 4).

The possible upper boundary of the Vasalemma Formation was preserved only in few limited areas in the westernmost part of the quarry (locality 5 in Fig. 1). In some places the depressions between the convex upper surfaces of the reefs and grainstone bodies contain up to 5–7 cm thick lenses of grainstone, rich in rounded bio- and lithoclasts and impregnated discontinuity surfaces on their lower and upper contacts. The upper surface of the rocks of the Vasalemma Formation is marked by the ice scratches from the Pleistocene period.

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Stop 4. Paekna quarry

Leho Ainsaar and Tõnu Meidla

The Paekna quarry is situated north of the Tõdva–Nabala road. It is an old quarry, which has been used intermittently for more than half a century. In the early 1990s a new attempt was made to restart limestone exploitation and a new, 2.8 m deep excavation was opened in the centre of the quarry. The attempt was unsuccessful and most of the monoliths extracted are still lying around. Today the quarry exposes the upper part of the Rakvere Stage in a thickness of 5 m (Fig. 1).

In northern Estonia the Rägavere Formation (Rakvere Stage) is characterised mainly by pure micritic limestone (calcilutites) with a low content of skeletal debris (Hints & Meidla 1997). However, the uppermost part of the formation (Tudu Member), exposed also in Paekna, may contain up to 20% fine skeletal material, mainly debris of echinoderms, trilobites, algae and gastropods (Põlma 1972; Põlma *et al.* 1988). The limestone in the Paekna quarry is mainly packstone or wackestone, with *Vermiporella* algae prevailing among macrofossils. On the western margin of the quarry, the weathered argillaceous packstone-wackestone of the Paekna Formation (Nabala Stage) could have been occasionally exposed or excavated (Nõlvak & Meidla 1990), but in early 2004 this part of the section was covered.

Macrofossils are relatively rare in the limestone, but microfossils are rather abundant. Detailed micropalaeontological logs (ostracods, chitinozoans) are published in Nõlvak & Meidla (1990), together with the detailed range charts of the same groups in the nearby Nabala-16 borehole, which penetrates into the basal part of the Rakvere Stage.

The limestones of the Rägavere and Paekna formations were formed during the sea level highstand on warm-water carbonate shelf. The change of mud-supported mudstone-wackestone to grain-supported packstones in the upper part of the Rakvere Stage may indicate shallowing and seaward facies progradation during the Rakvere–early Nabala highstand.

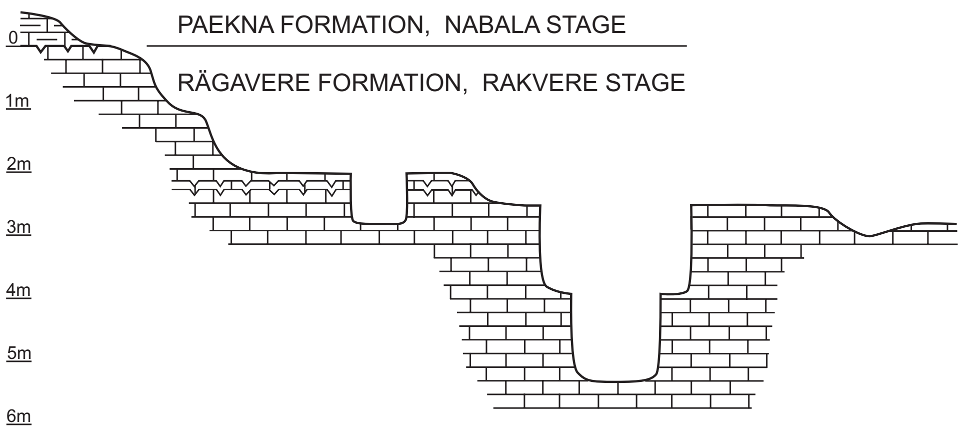

Fig. 1. Schematic cross-section through the Paekna quarry (Nõlvak & Meidla 1990, updated).

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Stop 5. Võhmuta quarry

Leho Ainsaar

A few years ago a new quarry exposed the coquina limestone of the Tamsalu Formation, Juuru Stage, Llandovery, near the Järva-Jaani–Tamsalu road in Võhmuta village. The section is very similar to that in the older and well-studied Karinu quarry situated 4 km south of Võhmuta (Nestor 1990).

The main part of the section exposes white massive coquinoid rudstone of *Borealis borealis* (Tammiku Member) in a thickness of 5 m (Fig. 1). The rock is composed of complete and fragmental valves of the brachiopod *B. borealis*. Interestingly, *B. borealis* is represented in the Tamsalu coquina almost exclusively by ventral valves (M. Rubel pers. comm.). The valves of *B. borealis* may comprise up to 64% of the coquina rock of the Tammiku Member (Jürgenson 1966). The matrix contains pellets and rounded skeletal particles in sparry calcite. The bedding planes are usually stylolites, with occasional thin marl coatings. The upper boundary of the *Borealis*-bank is a wavy erosional surface (Nestor 1990). The coquina contains occasional stromatoporoids.

The Karinu Member covering the *Borealis*-bank (Tammiku Member) is exposed in the north-eastern and south-western corners of the Võhmuta quarry in a thickness of up to 0.8 m. It comprises a yellowish biostromal stromatoporoid bank of irregular nodular coenostea (lower part) and fine-grained pelletal grainstone with stromatoporoids (upper part). According to the data of Nestor (1990) from the Karinu quarry, the most common stromatoporoids in the Karinu Member are *Clathrodictyon boreale* Riab., *C. kudriavzevi* Riab. and *Ecclimadictyon microvesiculosum* Riab. Heliolitid corals are also widespread in the biostromal limestone.

The shell bank of the Tamsalu Formation was deposited in a shallow-water high-energy environment in the shoal belt of the Silurian Baltoscandian Sea. The *B. borealis* accumulation zone is distributed in the east–west direction all over central Estonia, being about 200 km long and 30 km wide. The *Borealis*-bank is up to 13.5 m thick (Nestor 1997). It represents the first Silurian shallowing in the Baltoscandian Basin, lying on the argillaceous limestones (wackestones) of the Varbola Formation, Juuru Stage, about 10–20 m higher than the Ordovician–Silurian boundary. The massive *Borealis*-bank, resistant to erosion, has later given shape to the core of the Pandivere Upland in this part of Estonia.

The limestone of the *Borealis*-bank contains minor amounts of dolomite (1–10%) and siliciclastics (clay; usually <2%) in the Karinu–Tamsalu area (Jürgenson 1966). It has been used for lime production for over a century as pure calcitic material. The Võhmuta quarry is exploited by the enterprise AS EDK and the material extracted is used in chemical industry.

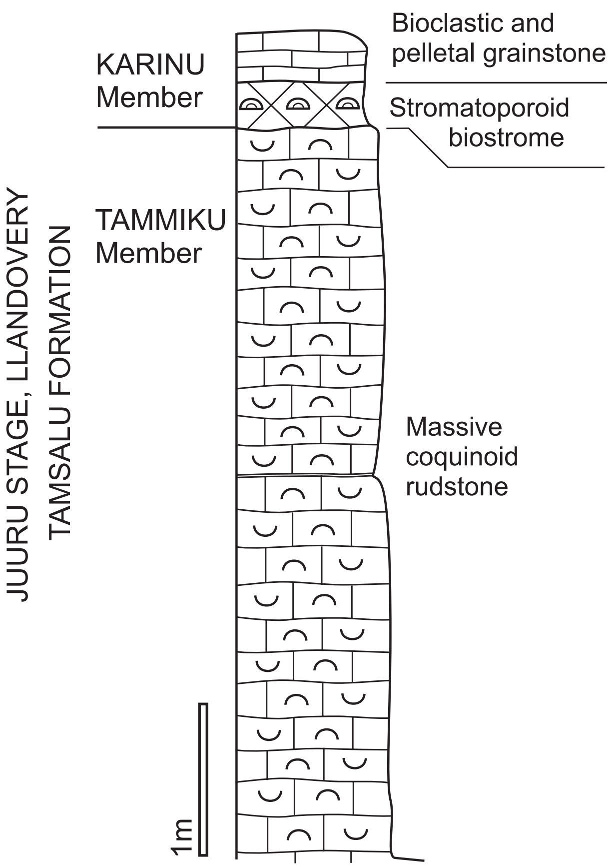

Fig. 1. Võhmuta quarry section.

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Stop 6. Porkuni quarry

Linda Hints and Asta Oraspõld

The old Porkuni quarry is located near the Porkuni village by the Tamsalu–Kullenga road, some 20 km SW of Rakvere. The locality was known already to Eichwald (1854), who first mentioned the Borkholm dolomite as a specific type of rocks (Borkholm is the old German name for Porkuni). Schmidt (1858) established the "Borkholm´sche Schicht" (=Porkuni Stage), which forms the roof of his "Untersilurische Formation" (=Ordovician). Since then the Porkuni quarry has served as the (unit-) stratotype for the Porkuni Stage.

Schmidt recognised a succession of four different types of rocks in the Porkuni quarry: "Encriniten-lager", followed by dolomitic limestones, brownish marls and white limestones in the upper part. Later on, geographical names were provided for these units: the Röa, Vohilaid, Siuge and Tõrevere members, respectively. All four members belong to the Ärina Formation which represents the youngest Ordovician rocks in northern and central Estonia. In southern Estonia, the Porkuni Stage is represented by the Kuldiga and Saldus formations. Based on the biostratigraphical and stable isotope data the Ärina Formation exposed in the Porkuni quarry most likely correlates with the lower part of the Kuldiga Formation, thus representing only early Porkuni time (Nõlvak & Grahn 1993; Kaljo *et al*. 2001).

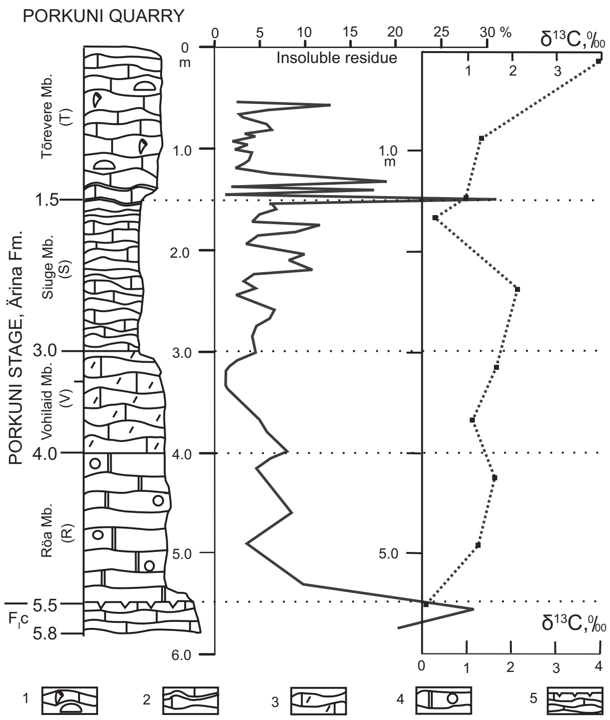

Fig. 1. Porkuni quarry section after Hints *et al*. (2000).

The description of the Porkuni quarry section according to Hints *et al*. (2000) is as follows (from base to top, with thickness shown in parentheses):

Adila Formation

5.47–5.83+ m (0.36+ m): yellowish-grey to yellow micro- to fine-crystalline variably calcitic and argillaceous dolostone with numerous yellow horizontal patterns, which may mark the weathered discontinuity surfaces. Skeletal debris consists of fragments of bryozoans, brachiopods and echinoderms. The interval comprises brachiopods, bryozoans and rugose corals (e.g., *Thaerodonta nubila* (Rõõmusoks), *Encrinurus moe* Männil, *Pirgumena martnai* Rõõmusoks, *Vellamo* cf. *silurica* Öpik, *Similoleptaena* sp.). These lowermost beds were described in a pit dug into the quarry floor.

Ärina Formation

Röa Member

4.00–5.47 m (1.47 m): yellowish, thick-bedded, predominantly fine-crystalline dolostone. Pores are rare and fine, representing the casts of skeletal debris. Unevenly distributed crinoid ossicles are poorly preserved. In the middle part of the interval (depth 4.6–5.1 m) the casts of ossicles or stem fragments (up to 4 cm long) are more frequent and the rock has fine- to middle-crystalline or evenly coarse-crystalline texture. Dolomitisation is the lowest in the upper part of the interval; the content of insoluble residue is low (4–10%). The upper boundary of the Röa Member is marked by a wavy discontinuity surface without impregnation.

According to the published data (Schmidt 1858; Rõõmusoks 1991), the brachiopods *Elsaella bekkeri* (Rosenstein), *Hindella*? sp., *Thaerodonta nubila* Rõõmusoks, trilobite *Valdariops eichwaldi* (Schmidt) and some other fossils occur in the upper half of this interval.

Vohilaid Member

3.30–4.00 m (0.70 m): yellowish, light grey, middle- to thick-bedded slightly dolomitic skeletal grainstone. The skeletal debris consists mainly of echinoderm fragments and a smaller amount of unevenly distributed fragments of bryozoans, corals and brachiopods. On some levels occur carbonate clasts 2–3 mm in diameter. The content of insoluble residue varies from 1.4 to 6%. Fine- to coarse-crystalline pure calcite forms the rock matrix.

3.00–3.30 m (0.30 m): light grey limestone, partly recrystallised and dolomitised. The lower half of the interval is similar to the previous interval in rock texture, the upper half – to the overlying strata in the decreased content of skeletal debris and increased content of insoluble residue (1.3–5%). In some places concentrations of crinoid ossicles occur, shells and skeletal fragments are partly silicified.

The shelly fauna of the Vohilaid Member consists of rugose and tabulate corals, bryozoans and brachiopods, but skeletal fragments of echinoderms are also present. The most complete list of fossils (about 40 species) was given in a manuscript by Rõõmusoks, who listed the corals *Kodonophyllum rhizobolon* (Dybowski), *Palaeophyllum fasciculum* (Kutorga), *Priscosolenia prisca* (Sokolov), brachiopods *Platystrophia* cf. *humilis* Oraspõld, *Streptis undifera* (Schmidt), *Ilmarinia ponderosa* Öpik, *Vellamo silurica* Öpik, *Aphanomena luna* (Lindström), etc. The occurrence of large tabulate corals (*Paleofavosites dualis* (Sokolov); the corallum of the holotype of this species can be observed in the quarry wall) in the southern wall of the quarry shows the beginning of the formation of reefs, which continue also in the overlying strata.

The Siuge Member

1.50–3.00 m (1.50 m): yellowish-grey, beige to brownish-grey, dolomitic micro- to fine-crystalline wackestone and packstone. The nodular or wavy structure is caused by the wavy or branching thin interlayers of slightly kerogenous marl (2–10 cm in thickness). The content of predominantly fine-grained skeletal debris (crinoids, ostracods, brachiopods) reaches up to 25%; at some levels it is orientated according to the horizontal lamination. The upper boundary of the unit is lithologically sharp.

The Siuge Member contains a diverse association of corals and brachiopods including *Palaeofavosites porkuniensis* Sokolov, *Porkunites amalloides* (Dybowski), *Sclerophyllum sokolovi* Reiman, *Schmidtomena acuteplicata* (Schmidt), *Streptis undifera* (Schmidt), *Laticrura* sp. n., *Pirgumena* sp., *Reushella* sp.; characteristic are also delicate reticulate or branching bryozoans and dendroid graptolites *Callograptus kaljoi* Obut et Rytzk, *Dictyonema delicatulum* Lapworth, *Mastigograptus crinitus* Obut et Rytzk. Most of the fossils are silicified and well preserved.

Tõrevere Member

1.5–0.00 m (1.5 m): brownish-grey, micro- to fine-crystalline coral-stromatoporoid limestone with wavy to massive structure. Fine-grained skeletal debris makes up 25–30% of rock and consists mostly of echinoderm fragments. The matrix is microcrystalline or biomicritic (wackestone). The stromatoporoid-coral reefs are well represented in this interval. The most characteristic fossils of these reefs are *Clathrodictyon gregale* Nestor, *Ecclimadictyon porkuni* Nestor, *Eocatenipora parallela* (Schmidt), *Paleofavosites nikitini* (Sokolov), *Rhabdotetradium frutex* Klaamann, *Holacanthia tubula* (Dybowski), *Strombodes middendorfi* (Dybowski). Most of the brachiopods occurring in the micritic limestone of the Tõrevere Member are common also in the underlying Siuge and/or Vohilaid members.

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Stop 7. Kunda-Aru quarry

Rein Einasto and Linda Hints

The Kunda-Aru quarry is located about 2.5 km south of Kunda, to the east of the Põdruse–Kunda road. It extends for about 1.8 km in the NS and 1.4 km in the WE direction (Fig. 1). The quarry is owned by the company Kunda Nordic Cement, which was founded in 1992 and is today a member of the international Heidelberg Cement Group. Cement production in Kunda has a 130-year history. It was developed thanks to the local raw material, the Ordovician high-quality limestone, quarried in the vicinity of the town and Estonian oil shale used as fuel for the kilns. The limestone is used for the manufacture of cement and crushed limestone. On average 650 000 tonnes of cement are produced in a year, which enables its exportation to many European countries.

Cement manufacture in Kunda caused serious dust pollution from clinker kiln and cement mills by about the end of the 20th century. The reduction of the pollution has been an essential task facing the company, and nowadays the filters installed to clean the exhaust guarantee that the production process meets the environmental requirements.

The Ordovician in the Kunda-Aru quarry is about 20 m thick and comprises the section from the upper Kunda to the middle Uhaku stages. The formations and members of the sequence are shown in Fig. 2. The excavated part belongs mainly to the Lasnamägi and Uhaku stages exposed in the quarry walls. The Aseri Stage is partly exposed in the walls of the drainage ditches. The lowermost part of the section, the Aseri and Kunda stages are best observable in the pumping-station in the NW part of the quarry.

The characteristic North Estonian sequence with some specific features in the central part of the area, including the vicinity of Kunda, is represented in the quarry. In this area glauconite grains are absent or sparse in the Loobu Formation, the oolitic limestones of the Napa Formation are missing in the top of the Kunda Stage, and the limestone of the Aseri Stage contains brown ooids mainly in its lower half.

Following the stratigraphical practice of the last decade, the lower boundary of the Uhaku Stage is identified by the appearance of the graptolite *Gymnograptus linnarssoni* Moberg. In the type area of the Lasnamägi Stage in the Tallinn–Lasnamägi section this species appears about 1.5 m above the Pae Member of the Väo Formation. By the bed-by-bed correlation of the stratotype Lasnamägi and Kunda sections using the bed names given by quarry workers, the boundary between the Lasnamägi and Uhaku stages falls into bed 46 (Estonian *Raudsüda*) within the Kostivere Member.

The sedimentary structures, such as current or slide marks, can be followed sometimes on the bedding planes on the quarry floor. The upper surface of the carbonate rocks is wavy in the areas surrounding the quarry. The Pleistocene ice scratches have been observed on this surface after the removal of the Quaternary deposits during the expansion of the excavating area.

Among the fossils occurring as loose specimens

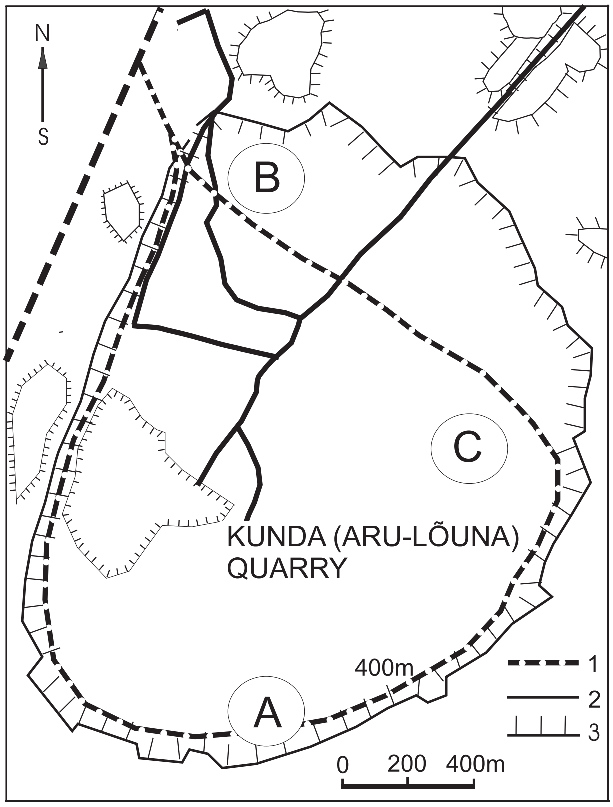

Fig. 1. Sketch map of the Kunda-Aru (Lõuna-Aru) quarry. **1** – road, **2** – railway, **3** – limit of the quarry. **A** – the uppermost part of the sequence is exposed (Lasnamägi and Uhaku stages); **B** – the lowermost part of the sequence (Kunda and Aseri stages) is exposed; **C** – bedding planes with the flow marks are exposed on the quarry floor.

cephalopods (from the Aseri and Kunda stages), bryozoans, trilobites and brachiopods are common. In the upper part of the sequence the cystids *Echinosphaerites* and *Heliocrinites* are rather numerous. The brachiopods *Lycophoria, Antigonambonites* and *Estlandia* are found in the lower part of the sequence, and *Clitambonites, Platystrophia, Cyrtonotella, Hesperorthi* and *Bekkerina* and others in the upper part. Trilobites, identified by Helje Pärnaste, are represented by *Megalaspidella* spp., *Pliomera fisheri* (Eichwald) (from the Kunda Stage), *Platyasaphus*? *latisegmentatus* (Nieszk.), *Cornasaphus* spp., *Pseudoasaphus* spp. (Aseri Stage), *Illaenus* spp., *Pseudobasilicus*? sp., *Paraceraurus* sp. (Lasnamägi Stage), *Xenasaphus* sp. (Uhaku Stage) and others. The well-known mass perishing of *Xenasaphus devexus devexus* (Eichwald) in bed 28 (named *Tige seitsmene;* Jaanusson 1939, 1940) or in beds 30–31 (*Muldvalge*; Orviku 1940) near Tallinn is documented in Kunda in the upper part of bed 31 (*Alumine muldvalge,* see Fig. 2). A similar level is known in the Volkhov River section at about the same stratigraphical level in the middle part of the Valimskaya Formation.

Stage boundaries

The boundary between the Kunda and Aseri stages on the stratotype area is marked by an even pyritized discontinuity surface, with deep (8–15 cm) burrows (pockets). It is underlain by thick-bedded skeletal fine-detrital packstone with rare small glauconite grains of the Loobu Formation and overlain by medium-bedded unsorted skeletal grainstone with brown chamosite oolites and rare well-rounded pyritized pebbles (10 cm) of the underlying limestone of the Loobu Formation. A few centimetres thick interbed of detrital marlstone lies above the boundary.

The lower boundary of the Lasnamägi Stage is marked by an uneven limonitized discontinuity surface. It is underlain by the medium-bedded oolitic limestone with frequent brown chamosite oolites of the Aseri Stage and overlain by thick-bedded hard unsorted skeletal limestone with rare redeposited chamosite oolites and numerous small white lime oolites in the basal bed (0.3 m). The boundary discontinuity surface includes an interbed of marlstone (3–5 cm) with unsorted biodetritus and chamosite ooids.

The lower boundary of the Uhaku Stage is by Männil (1976) in the middle of the Kostivere Member on the level of bed 46 (*Raudsüda*). It is marked by a sharp phosphatised discontinuity surface. The basal bed of the Uhaku Stage (3–8 cm) which marks the boundary of minicyclites and above which *Gymnograptus linnarssoni* Moberg appears is represented by skeletal grainstone with a sharp uneven pyritic discontinuity surface on the top of it.

Formations exposed in the Kunda quarry

The Loobu Formation represents the upper part of the Kunda Stage. It consists of thick-bedded unsorted to well-sorted skeletal fine-detrital light grey limestones with some distinct pyritized discontinuity surfaces and numerous nautiloids. In the middle part of this interval more argillaceous fine nodular interbeds (20 cm) occur. In the quarry an about 3 m thick part is visible, the lower part is under water (1.5 m), or covered by loose material (0.5 m) (Fig. 2).

The Kandle Formation of the Aseri Stage is represented by oolitic limestone, medium-bedded brownish-grey skeletal packstone to grainstone with some sharp goethitized discontinuity surfaces. The formation is subdivided into the Malla (rare ooids, numerous nautiloids and trilobites) and Ojaküla (numerous ooids in more argillaceous interbeds) members, both representing a minicyclite (Fig. 2). The thickness of the formation is 1.3 m.

The Väo Formation of the Lasnamägi and lower Uhaku stages consists of clear medium- to thick-bedded light grey limestone with a large number (over 90!) of distinct phosphatized discontinuity surfaces with numerous deep burrows (pockets) penetrating the underlying bed (Fig. 2). Thin interbeds (0.2–4 cm) of wavy microlaminated marlstone intercalate with limestones. These subvertical (in the lower part sometimes horizontal) pockets are filled with dark grey argillaceous silty dolostone. The pyrite in these dolostones is goethitized because of weathering, and the dark grey colour of the rock has turned reddish-brown. The reddish-brown pattern of deep pockets is a characteristic feature of the Lasnamägi building stones. The discontinuity surfaces mark the boundaries of mini- and microcyclites. The total thickness of the formation is 10.8 m. The Väo Formation is subdivided into three members, the middle and upper one representing the classical Lasnamägi Building Limestone.

(a) The Rebala Member forms the lower part of the Väo Formation. The uppermost Rebala Member (2.8 m thick) is represented by thin- to thick-bedded argillaceous limestone, very seldom used as building stone. The thick-bedded lower part (0.75 m) is considered as a good building stone but excavating it below the argillaceous limestones is complicated. In the quarry the Rebala Member is exposed only partly.

(b) The Pae Member is 0.8 m thick in the Kunda quarry. The rocks of this member are commonly sec-

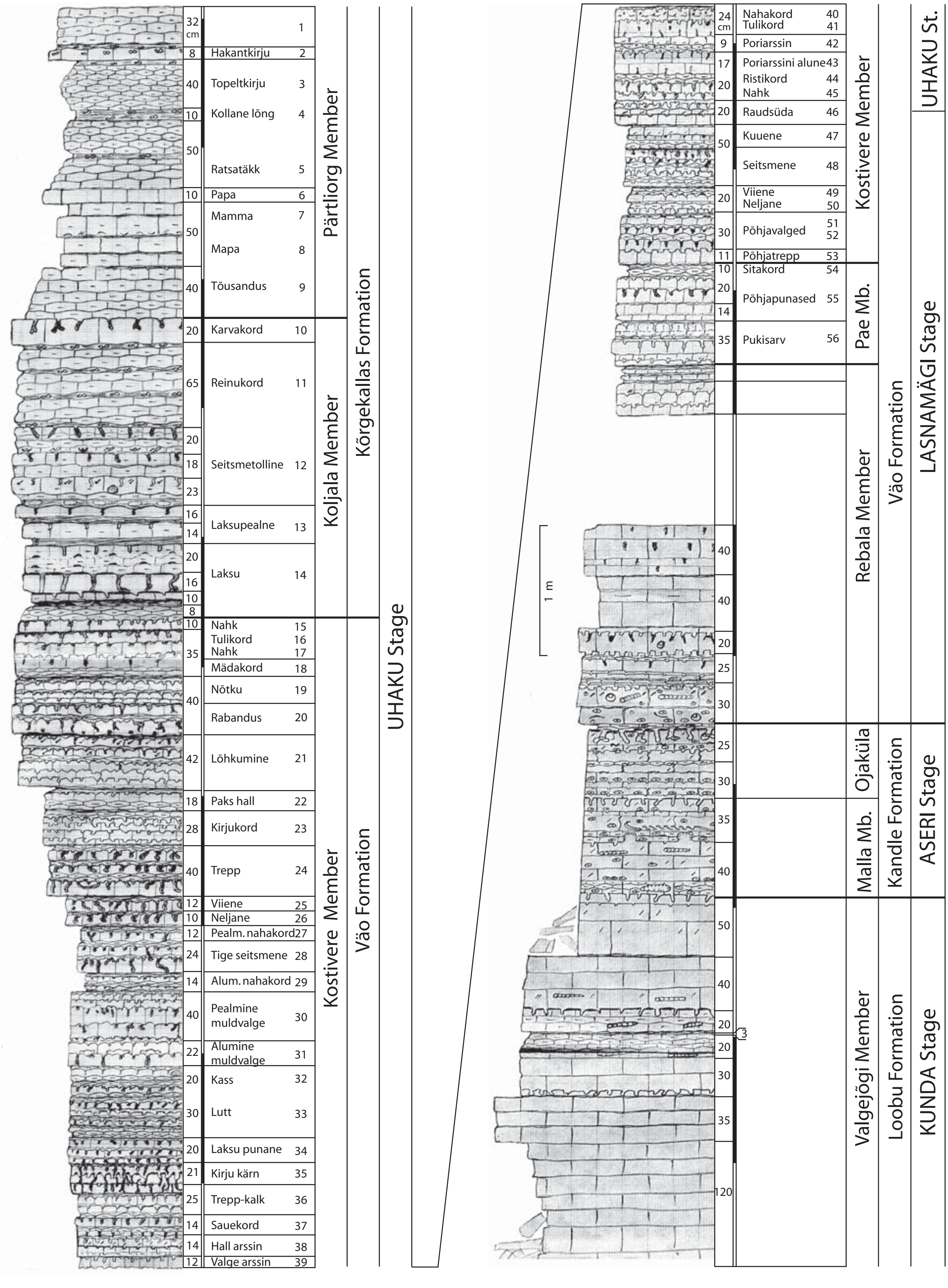

Fig. 2. Composite section of the Middle Ordovician in the Kunda-Aru quarry. The bed names (in Estonian) by quarry workers, thickness and number are given right of the log.

ondarily dolomitized and well visible as a more brownish belt in the sections. In Kunda the dolomitization is very weakly developed and the member does not differentiate very clearly in the quarry walls. The Pae Member consists of medium- to thick-bedded slightly argillaceous limestone with interbeds of calcareous marlstone (up to 10 cm thick in the upper part).

(c) The Kostivere Member (the upper part of the Väo Formation), 7.2 m in thickness, is represented by light grey thick- to medium-bedded pure skeletal packstone with numerous distinct discontinuity surfaces, interbeds of marlstone and some interbed of skeletal grainstone marks the basal part of microcyclites. The member represents the best quality building stone in northern Estonia.

The Kõrgekallas Formation is represented by wavy-bedded to seminodular argillaceous limestone (unsorted skeletal wackestones) with interbeds (8–20 cm) of pure skeletal packstone and some discontinuity surfaces in the lower part. The total thickness of the formation is 4.7 m. The lowermost and middle members of the formation, the Koljala and Pärtlioru members, are exposed in Kunda.

(a) The Koljala Member is 2.3 m thick and consists of relatively pure skeletal packstone and argillaceous limestone (wackestone) with marlstone interlayers. On the upper surfaces of less argillaceous limestone there occur discontinuities with pockets, similar to those in the Lasnamägi Stage. They mark the boundaries of microcyclites. The kukersite kerogene grains appear in the marlstone interlayer in the upper part of the member (0.85 m). The upper boundary is marked by the last distinct discontinuity surface with burrows (pockets).

(b) The Pärtliorg Member, 2.4 m in thickness, consists of seminodular argillaceous wackestone with rare pure limestone (packstone) and kuckersinic marlstone interbeds.

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Stop 8. Kunda clay quarry

Kalle Kirsimäe and Kaisa Mens

The Cambrian terrigenous deposits are distributed in large areas on the East European Platform from the Baltic Sea to Moscow. Cambrian sediments outcrop as a narrow belt at the front of the Baltic–Ladoga Klint along the present northern margin of the Baltic Basin. In the Kunda quarry Lower Cambrian sand-to-siltstones and clays are exposed.

Based on lithological and palaeontological data, the Lower Cambrian sediments in the northern Baltic area are divided into eight formations (Mens *et al.* 1990). In northern Estonia the Lontova, Lükati and Tiskre formations form the Lower Cambrian succession (Mens & Pirrus 1997). These three formations are represented also in the vicinity of Kunda (Fig. 1).

The Lontova Formation consists mostly of up to 100 m thick pelagic marine greenish-grey and variegated clays with interbeds of coarse- to fine-grained sandstones in the lowermost and uppermost parts. The Lontova succession forms a nearly complete transgressive–regressive sedimentary cycle. The near-shore facies equivalents of the Lontova clays in NW Estonia are represented by the Voosi Formation.

The Lontova Formation is transgressively succeeded by the Lükati Formation consisting of lithologically similar rhythmically interbedded greenish-grey clays and very fine-grained sandstones or siltstones, up to 20 m thick. The Lükati sediments were probably deposited under alternating water-energy conditions in shallow gulf-like basins. The Lower Cambrian in northern Estonia is completed by shallow-water silty sandstones of the Tiskre Formation. The greenish-grey clayey sediments of the Lontova and Lükati formations are traditionally called as the Cambrian "Blue Clay".

The Lontova Formation corresponds biostratigraphically to the *Platysolenites antiquissimus* Zone and the Lükati Formation, unconformably overlying the Lontova Formation, corresponds to the lower part of the *Mobergella* and *Schmidiellus mickwitzi* Zone. In NW Estonia, 5–50 m thick sandstones interbedded with thin clay interbeds of the Sõru Formation, corresponding to the *Rusophycus parallelum* Zone, occur between the Lontova/Voosi and Lükati formations (Mens *et al.* 1990).

The "Blue Clay" deposits are traditionally ascribed to the Lower Cambrian Tommotian and Atabanian (534-524 million years ago). However, there is palaeontological evidence supporting an older age (545-530 Ma, Nemakit-Daldynian and Tommotian) for these sediments (Moczydlowska & Vidal 1988; Volkova *et al.* 1990).

The sequence of greenish-grey and reddish-violet spotted homogeneous clays in the old abandoned "upper" Kunda quarry was suggested by Armin Öpik in 1933 as the stratotype of the Lontova beds (Formation), named after a small village next to Kunda. The lower part of these beds is exposed today in the new "lower" Kunda quarry, where the fine-grained unlithified clays lithologically defined as the Kestla Member represent the Lontova Formation.

The Kestla Member is palaeontologically characterized by *Platysolenites antiquissimus* and *Platysolenites lontova*, *Sabellidites cambriensis*, numerous pyritized worm tracks and hyolithid casts, an early gastropod *Aldanella kunda*, horn-like chitinous (?) sclerites, and leiosphaerides and *Tasmanites* acritarchs.

The "Blue Clay" is unique sediment. Despite its old age the clay has retained natural plasticity and a high water content/porosity, and is used as raw material for cement and brick industry. The clays were never deeply buried and/or strongly heated. This conflicts with the high diagenetic grade of the clay minerals found in the sediment.

Today, the company Kunda Nordic Cement uses the Lontova clay from the Kunda quarry for cement production in mixture with Ordovician limestone from the Kunda-Aru quarry, a few kilometres to the south of Kunda.

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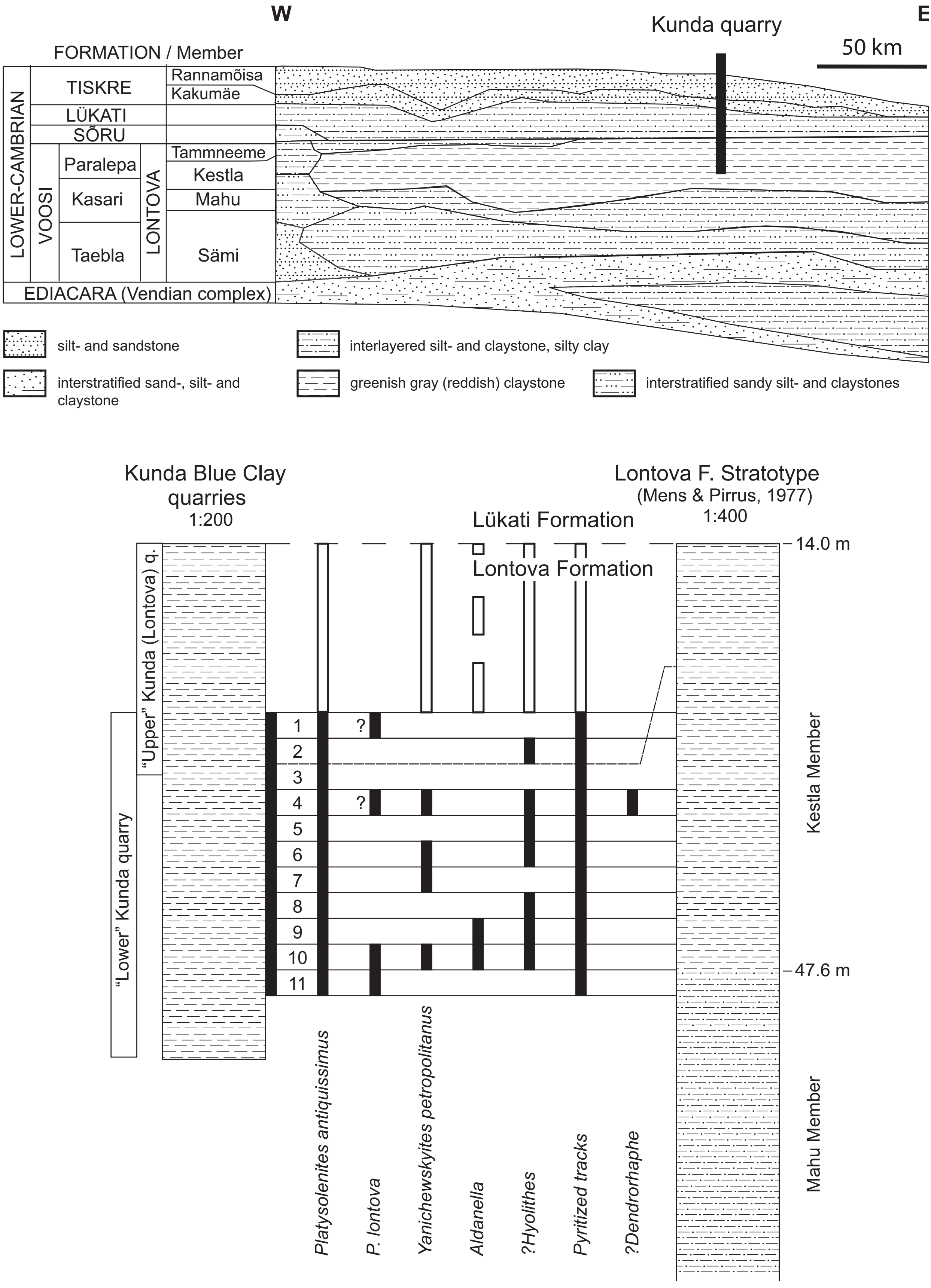

Fig. 1. Vendian - Lower Cambrian stratigraphy in northern mainland Estonia and sections in Kunda clay quarries and the type section of the Lontova Formation.

Stop 9. Kohtla quarry and Estonian oil shale

Tõnis Saadre

The Baltic Oil Shale Basin is situated in north-eastern Estonia, with a part of it extending eastward into Russia (Fig. 1). Three well-explored oil shale deposits – Estonia, Leningrad and Tapa – occur within this basin. The first two deposits are mineable, while the Tapa deposit in central Estonia is considered to be a prospective one. At present, the Estonia deposit, with an area of nearly 3000 km^2, is the largest commercially exploited oil shale deposit in the world. Total reserves of the deposit are approximately 5000 million tonnes. Nowadays oil shale is excavated in two underground mines – Estonia and Viru – and in three open cast pits – Narva, Aidu and Vanaküla (a small pit). In the year 2002 the total amount of excavated oil shale was 10.5 million tonnes. The Kohtla quarry works as part of the Aidu open cast pit.

Estonian oil shale – kukersite – is a unique mixed rock, widely distributed in the northern Estonian Ordovician sequence. It consists of the following main components: organic matter of algal and/or microbial origin (15–70%); terrigenous matter, mostly clay (10–75%); calcareous component, consisting mainly of calcitic matrix and skeletal detritus (10–75%; Bauert & Kattai 1997).

The oldest Ordovician occurrence of kukersite is in the sandy limestones of the Pakri Formation, Kunda Stage (see Stop. 2. Pakri Peninsula in this guide). Especially rich in kukersite is the sequence of the Middle–Upper Ordovician boundary beds of Llandeilo–Early Caradoc (Uhaku and Kukruse stages) age. Here up to 50 laterally continuous kukersite seams are registered. The seams can be traced laterally for 250 km in the east–west direction and 40–50 km in the north–south direction.

The Uhaku Stage is represented by the Kõrgekallas Formation in north-eastern Estonia. Though some of the individual beds are up to 40 cm thick, they have no commercial value. The Kukruse Stage is here represented by the Viivikonna Formation, comprising (from base to top) the Kiviõli, Maidla and Peetri members. In the Kohtla quarry the rocks of the Kiviõli Member, including the commercial oil shale seam, can be studied (Fig. 2; Kõrts & Einasto 1990). The Kiviõli Member is the richest in kukersite, especially its lower part. Here seven thick kukersite layers (A, A′, B, C, D, E and F_1) form the commercial seam of the Estonian Oil Shale Deposit (Fig. 2). The kukersite beds in the upper part of the member (F_2, F_3, F_4, F_5, G, H, J and K) have no commercial value, though some of them (G and H) are considerably thick (20–40 cm). The Maidla Member is comparatively low in kukersite – the seven mostly nodular to seminodular kukersite layers registered here have no commercial value. The Peetri Member has again a higher kukersite content – six kukersite layers and complexes are distributed here. Only seam III (0.6–2.3 m) at the base of the member may in future obtain commercial value in the Tapa region, in the western part of the oil shale basin (Bauert & Kattai 1997).

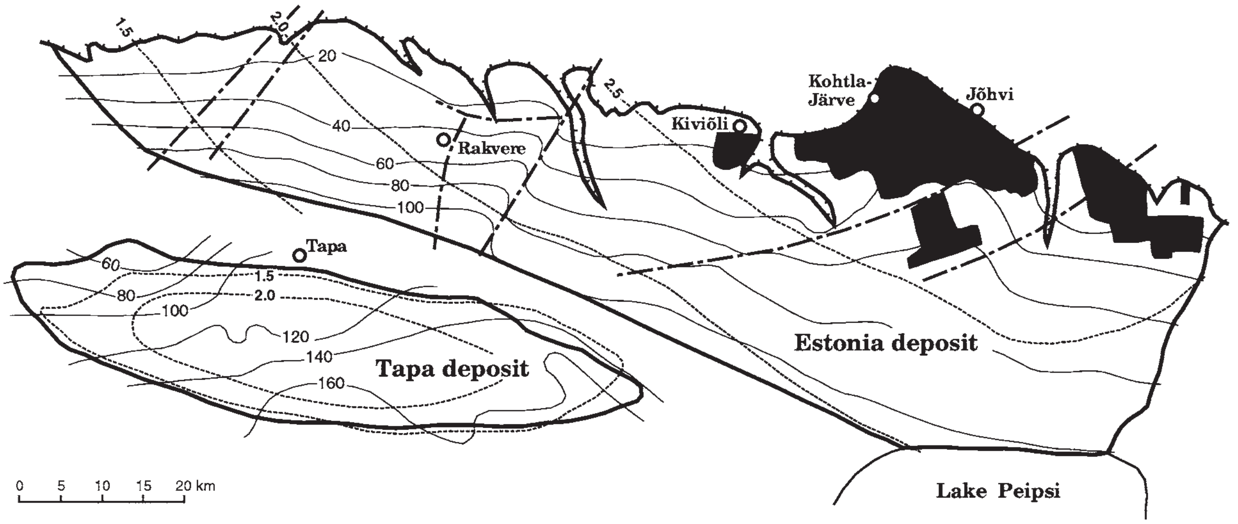

Fig. 1. Estonian Oil Shale Basin (after Bauert & Kattai 1997) showing depths from the surface and thickness isolines (dashed lines) for kukersite oil shale bed III (Tapa deposit) and for the commercial bed A-F_1 (Estonian deposit). Black area marks mined-out areas.

The Kõrgekallas and Viivikonna formations are characterized by frequent rhythmical alternation of different rock types: limestone with different contents of the argillaceous component and kerogene, kukersite and marl. The rocks are often dolomitized within tectonic disturbances and karstification zones. Both bedding structures and nodular textures of oil shale can be observed. The limestones are mainly medium-bedded (2–10 cm), marls – thin- to thick-bedded, of seminodular texture. Kukersite is thin- to thick-bedded, frequently also of nodular and semi-nodular (net-like) texture. The deposits have been bioturbated before lithification. Frequent burrows and discontinuity surfaces are registered in the sequence. The sequence is extremely rich in fossils: over 300 species have been identified in the Kukruse Stage, which is the most fossiliferous Ordovician stage of Estonia (Rõõmusoks 1970).

Several thin beds of kukersite and kerogenous marl occur also higher up, in the Haljala and Keila stages. In the eastern part of the Palaeobaltic basin thick kukersite layers are distributed in the Keila Stage, forming the commercial seam of the Tshudovo-Babinskoje Oil Shale Deposit. Thin interbeds of kerogenous marl are also distributed in the Rakvere and Nabala stages; a slight admixture of kukersite has been registered in the Oandu and Porkuni stages.

The genesis of kukersite and origin of organic matter are still problematic. All kukersite beds and seams form east–west directed sedimentary lenses with rather stable thickness. The organic matter of kukersite is alganite A (telalganite) derived from *Gloeocapsomorpha prisca*. Kukersite deposited in a basin with normal salinity. The northern part of the basin was embraced by the non-sedimentation area – rocky sea bottom, covered by algal mats. Kerogene was probably derived from there and deposited at some distance from the coastline (Männil *et al.* 1986; Bauert & Puura 1990). The kukersite deposition facies shifted gradually southward (basinward), showing a clear progradational pattern (Männil *et al.* 1986; Bauert & Kattai 1997).

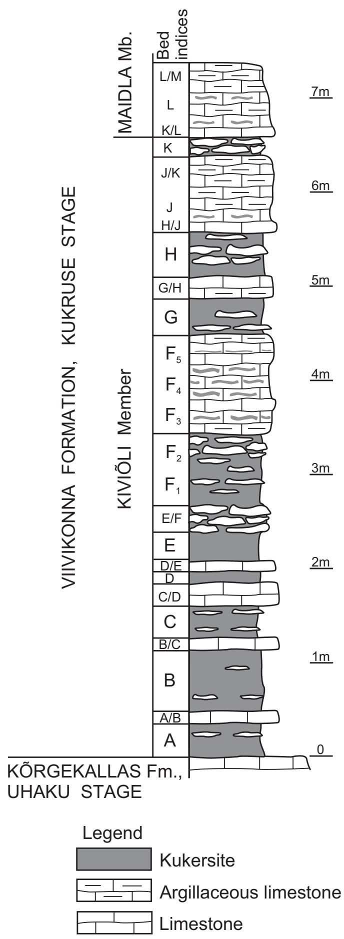

Fig. 2. Kohtla quarry section (Kõrts & Einasto 1990, updated in April 2004 by L. Ainsaar).

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Stop 10. Valaste waterfall

Oive Tinn

The artificial Valaste waterfall, falling from the 54 m high North Estonian Klint is the highest waterfall in Estonia. Its height is usually between 26 and 28 m, but after exceptionally heavy rainfalls the strong flow may erode a deep pit in the sandstone on the foot of the klint and the total height of the waterfall can be up to 30 m. Downwards, the waterfall continues as a 10–15 m high rapid, flowing into the sea.

Due to the slight southward dip (3–4 m per 1 km) of the limestone layers and the absence of water outlet, the fields in the klint area have been suffering from excessive water during rainy seasons. At the beginning of the 19th century, a 7 km long and up to 2 m deep drain was made to aid water to run off the manor's fields nearby. As a result, the water flow has cleaned and eroded the klint wall, exposing the wonderful Lower Cambrian to Middle Ordovician sedimentary section.

The section is stratigraphically similar to that described in the Ontika Klint, 3 km west of Valaste (Mägi 1990). It exposes the Lower Cambrian sandstone (Tiskre Formation), Furongian to Lower Ordovician sandstone (Kallavere Formation), black shale (Türisalu Formation) and glauconitic sandstone (Leetse Formation), and Middle Ordovician limestone and dolomite (Volkhov, Kunda and Aseri stages; Fig. 1). The banks of the stream below the klint expose locally the "blue clay" of the Lower Cambrian Lükati Formation.

In 1999 a platform was constructed in front of the waterfall to make the observation of this site safer and more attractive.

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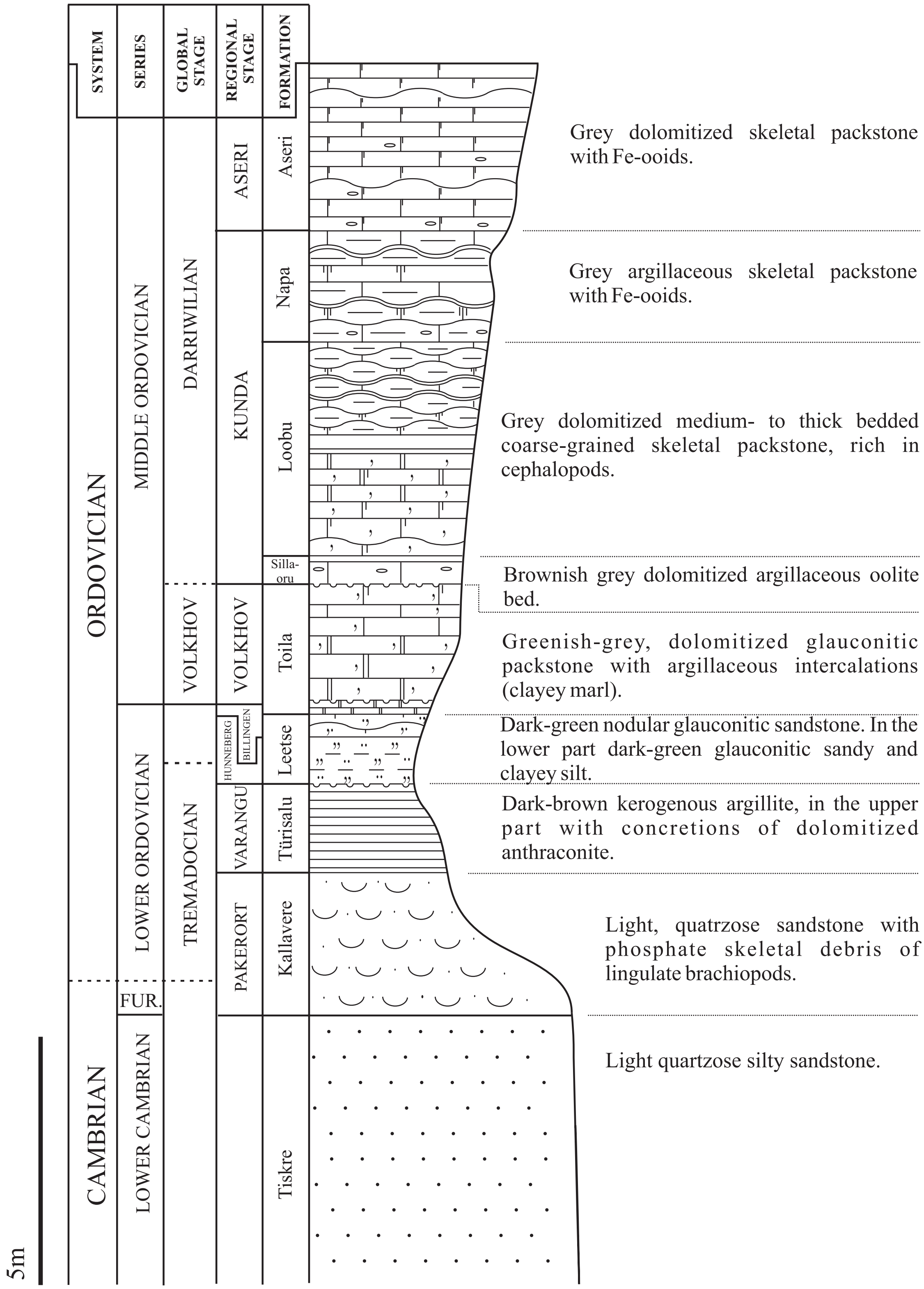

Fig. 1. Stratigraphy of the Valaste waterfall section.

Stop 11. Aluvere quarry

Leho Ainsaar and Tõnu Meidla

In the abandoned Aluvere quarry near Rakvere, south of the Tallinn–Narva road, the limestone succession of the Kahula Formation, Haljala Stage (Caradoc) is exposed. In the southern wall of the quarry 6 m of limestone can be studied, characterising the Aluvere and Pagari members of the Kahula Formation (Jõhvi Substage of the Haljala Stage; Fig. 1). The limestone can be classified as wackestone to packstone with a varying content (10–25%) of siliciclastic mud. This siliciclastics content varies rhythmically, as 10–20 cm thick cycles, clearly observable in the weathered walls of the quarry.

In the northern part of the quarry, in an old railway cut, the Vasavere and Aluvere members of the Kahula Formation are exposed. The section is well studied and repeatedly documented in publications (e.g., Rõõmusoks 1970; Põlma *et al.* 1988). In 1970, the limestone succession was documented in a thickness of 9.5 m by Rõõmusoks, but only about 7 m is still visible. Two thin (2–4 cm) K-bentonite beds outcropping in the middle part of this wall belong to the Grefsen K-bentonite complex (Bergström *et al.* 1995). The upper bed is considered as the base of the Jõhvi Substage (previously referred to as a stage) of the Haljala Stage. This part of the quarry wall is heavily weathered and dangerous to access.

Rhythmically bedded argillaceous wackestones–packstones of the Haljala Stage have been deposited in the temperate climate open marine conditions on the upper shelf or ramp (Nestor & Einasto 1997). The abundance and high diversity of benthic shelly fauna refers to the depositional environment at a moderate depth, probably within the photic zone. The Aluvere quarry has been a well-known fossil site since the beginning of the 20th century (Rõõmusoks 1970). More than 150 species of macrofossils are reported from the locality, brachiopods, bryozoans and various microfossils being abundant and diverse. Some examples of the most common macrofossils are *Porambonites baueri* Noetling, *Platystrophia lynx lynx* (Eichwald), *Clinambon anomalus* (Schlotheim), *Estlandia pyron silicificata* Öpik, *"Chasmops" wenjukovi* (Schmidt), *Pyritonema subulare* (Roemer), *Diplotrypa petropolitana* (Nicholson), *Ischadites murchisoni* (Eichwald), *Conichnus conicus* Männil (for an extended lists see Rõõmusoks 1970 and Põlma *et al.* 1988).

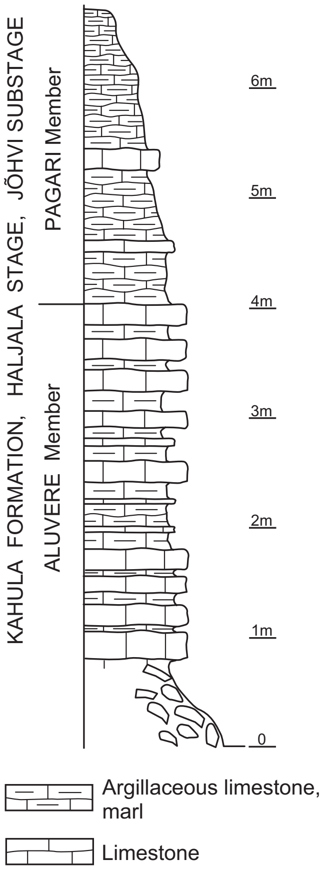

Fig. 1. Aluvere quarry section, southern wall.

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Stop 12. The Kärdla Impact Event in the Soovälja K–1 drill core (Arbavere field station of the Geological Survey of Estonia)

Kalle Suuroja and Anne Põldvere

The Soovälja (K–1) drill hole (58°58.530′N, 22°46.333′E), located inside the circular Kärdla impact structure, was made in the Kärdla meteorite crater in the course of geological deep mapping of Hiiumaa Island in 1990 (Põldvere 2002). It is the deepest drill hole (815.2 m) in Estonia, penetrating 24 m thick Quaternary deposits, Ordovician post-impact sedimentary rocks (to a depth of 301.2 m) and the 221.6 m thick polymict clastic impact breccia complex (at 301.2–522.8 m). The following, 65.7 m interval is represented by the strongly brecciated crystalline basement (crater floor from monomictic breccia to a depth of 588.5 m). The lowermost part of the Palaeoproterozoic subcrater basement is fractured (down to 815.2 m; Põldvere 2002).

The crater structure in the north-eastern part of Hiiumaa Island (close to the town of Kärdla) is a unique geological object discovered in 1974. Its impact (meteoritic) origin was established in 1981 (Masajtis *et al*. 1980). The age of the Late Ordovician Kärdla impact event is precisely dated by micropalaeontological data and biostratigraphical correlations. The crater was formed in a shallow epicontinental sea not far from the non-sedimentation area (Puura & Suuroja 1992; Ainsaar *et al*. 2002) 455 million years ago (Haljala Stage, Idavere Substage; Puura & Suuroja 1992; Grahn *et al*. 1996; Nõlvak 2002). At that time the crystalline basement was covered by poorly compacted siliciclastic and clayey Cambrian–Lower Ordovician sediments and an about 20 m thick layer of Middle and Upper Ordovician carbonate rocks.

Nowadays the circular crater structure is 4 km wide and more than 500 m deep, with a central uplift (diameter about 800 m) rising about 130 m from the crater floor (Fig. 1). The rim wall (the highest point 110 m above the target level) is cut by at least two resurge-excavated gullies (lowlands between the town of Kärdla and Paluküla village, and Ala and Tubala villages). The crater is surrounded by an elliptical ring fault (diameter 12–15 km), with deformed sedimentary rocks below the target level. Marine geophysical investigations in 1996 revealed a presumptive ring fault zone (Suuroja *et al*. 2002) at two sites in Kärdla Bay, marked by a 5–10 m high terrace in the bedrock and by a submarine beach ridge. The blocks of the Cambrian sedimentary cover between the rim wall and the ring fault are strongly disturbed, fractured and folded, but dislocations do not reach the crystalline basement. In physical properties the vertical influence of the impact is observable until the bottom of the Soovälja (K–1) core, but it extends to a depth of about 950 m, that is, about 430 m below the crater floor (Plado *et al*. 1996). The basement of the surroundings of the crater is overlain by sedimentary bedrock consisting of siliciclastic and carbonaceous rocks, which have formed partly during the Vendian, but mainly during the Cambrian, Ordovician and Silurian.

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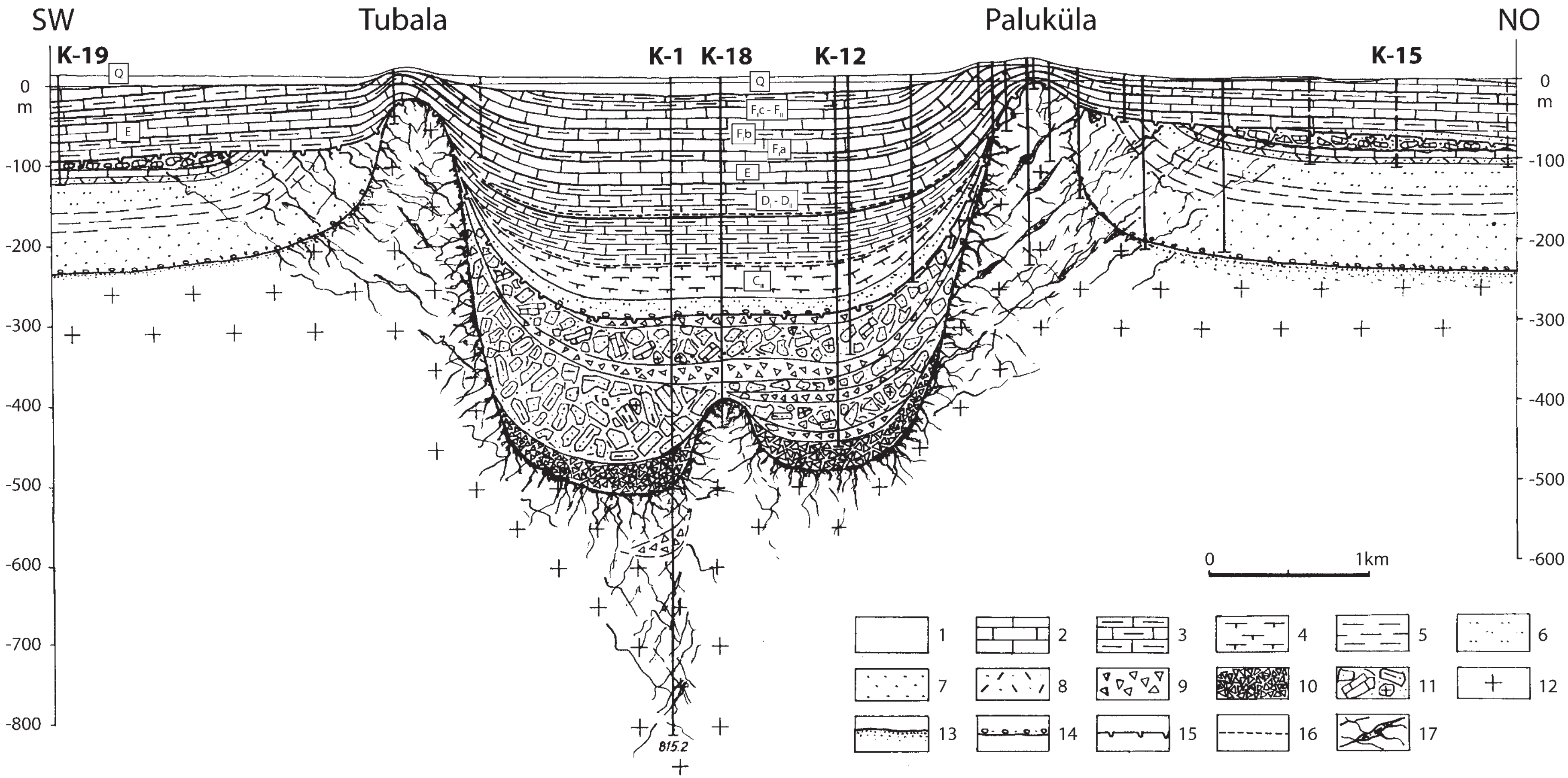

Fig. 1. Schematic geological cross-section of the Kärdla crater with location of some boreholes (after Puura & Suuroja 1992). **1** – Quaternary deposits; **2** – limestone; **3** – clayey limestone; **4** – marl; **5** – clay; **6** – silt and siltstone; **7** – sand and sandstone; **8** – sandstone with phosphate debris; **9** – **11** breccias, 9 – containing mostly crystalline rocks, 10 – crater bottom breccia containing crystalline rocks, 11 – consisting mostly of sedimentary rocks with admixture of crystalline rocks; **12** – crystalline rocks; **13** – weathered crystalline rocks; **14** – basal conglomerate; **15** – main discontinuity surface; **16** – metabentonite layer; **17** – fractured rocks with breccia dikes.