DANMARKS OG GRØNLANDS
GEOLOGISKE UNDERSØGELSE
RAPPORT 1996/98

WOGOGOB-94 Symposium

Working Group on Ordovician Geology of Baltoscandia, Bornholm - 94

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Preface

This is the volume of proceedings for the Working Group on Ordovician Geology of Baltoscandia (WOGOGOB). It contains 16 short papers which were presented orally or as posters in the symposium program. The volume is published by the Geological Survey of Denmark and Greenland (GEUS) and contains papers of many aspects on Ordovician stratigraphy and geology of Scandinavia, western Russia, the Baltic countries and Poland. The technical sessions of the meeting were held on Bornholm, Denmark, which is situated in the Baltic Sea and with pre and post meeting excursions on Bornholm, Sweden and Öland, Sweden. Approximately 50 persons from 14 different nations attended the symposium.

Many persons have helped in the process during the meeting and in the production of the volume. The organizer especially thanks Flemming Feilberg and Niels Schovsbo for their assistance during the meeting. Arne T. Nielsen guided the pre and post meeting excursions together with Svend Stouge. The editor especially thanks Hanne Sørensen and Heinke Andersen, Department of Stratigraphy, GEUS, for their help in retyping the manuscripts and assembling the final version of this volume.

Economical support has been given by the Danish Natural Science Research Council and by the Geological Survey of Denmark and the Geological Survey of Greenland, now the Geological Survey of Denmark and Greenland (GEUS).

Svend Stouge
Editor

Contents

REGRESSION IN CARADOC: EVIDENCES FROM SOUTHWESTERN ESTONIA (RISTIKÜLA CORE)

Leho Ainsaar, Kalle Kirsimäe and Tõnu Meidla
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Introduction

A stratigraphic gap in the Caradoc, at the boundary between the Keila and Oandu stages, has been recognized in East Baltic area, in North Estonia and in East Lithuania (Männil 1966, Laškovas & Paškevicius 1989, Männil & Meidla 1994), and in Sweden, both on Gotland and in Västergötland (Jaanusson 1973). This gap and related sediments of high water-energy origin are considered to be related to sea level fall in the shelf area (Vingisaar 1971, Kõrts *et al.* 1991). According to some suggestions (Bergström *et al.* 1992, Laškovas 1992) this regression may reflect a eustatic sea level change traced in different platforms.

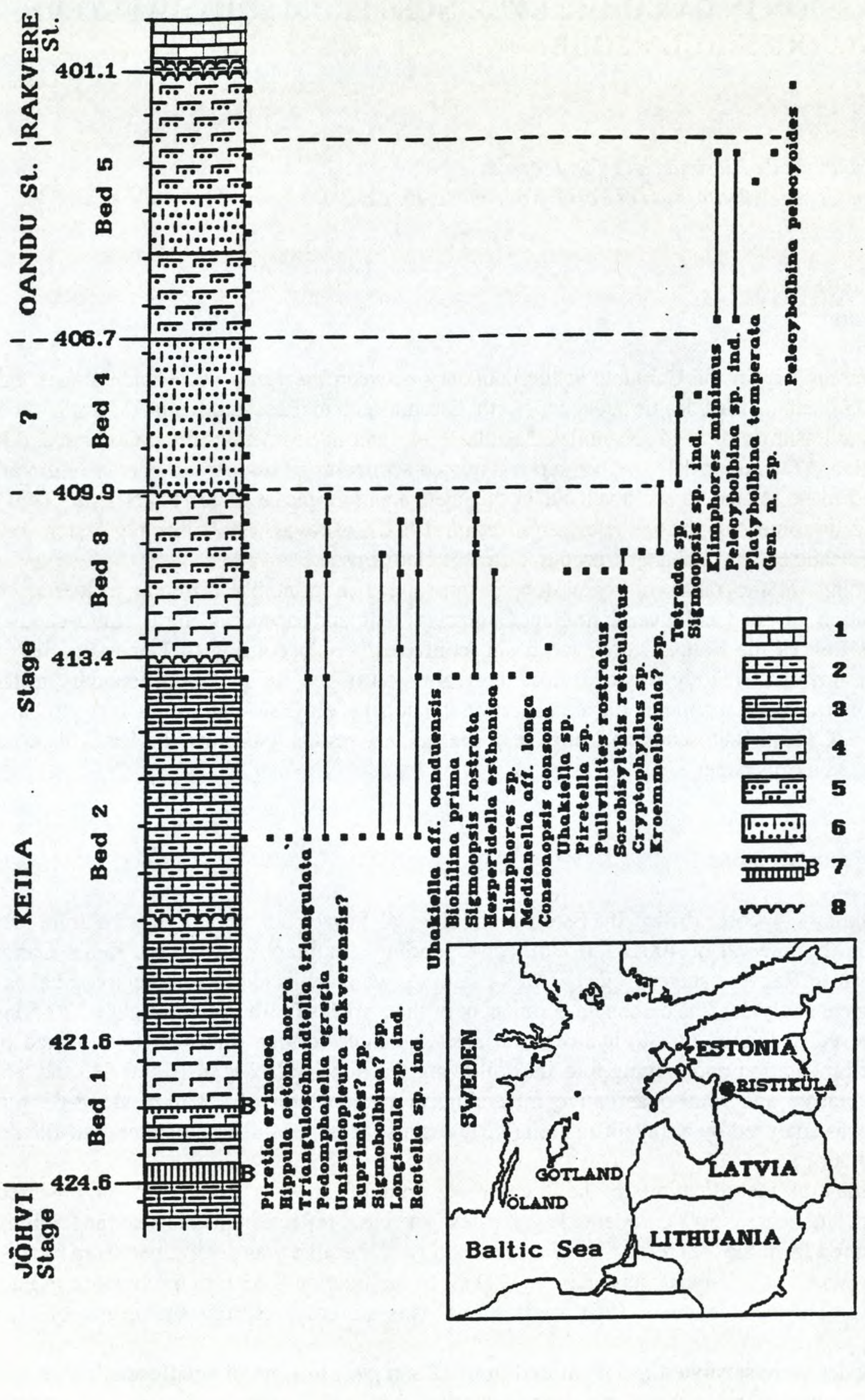
Depending on the range of regression, it must have in principle different influence on sedimentation in the shallower Estonian Confacies Belt and deeper Central Baltoscandian Confacies Belt of the Baltic Ordovician basin (confacies belts according to Jaanusson, 1976). The main aim of this study was to follow the regression of the Keila-Oandu boundary in the lithofacially transitional zone, in the area between the upper shelf (Estonian Confacies) and basin depression (Central Baltoscandian Confacies). Preliminary results, based on the Ristiküla core, are presented in this paper.

Sampling and Methods

The Ristiküla-174 core, drilled by Estonian Geological Survey 25 km SE from Pärnu, was sampled in the interval of 400.5-420.7 m, corresponding stratigraphically to the Keila, Oandu and lowermost Rakvere stages (Fig. 1). All 44 samples were subjected to sedimentological and mineralogical analyses. The carbonate component of the samples (with initial weight of 20-50 g) was dissolved in diluted (3.5%) hydrochloric acid. Insoluble residue (i.r.) was fractionated by gravity sedimentation and sieving into the following fractions: <2, 2-8, 8-16, 16-63, and >63 mm. Qualitative and semi-quantitative mineralogical composition of <2 mm and 8-63 mm fractions was analyzed by X-ray diffraction (XRD) using DRON-3M diffractometer (Mn-filtered Fe-*Ka* radiation).

Clay mineral identification in the fraction <2 mm of i.r. was performed on air-dried, glycolated, and heated (500°C) oriented aggregates. The basal reflection intensity of the kaolinite was measured from the 003 reflection. The crystallinity of the illite was determined from the 001 reflection (width at half-maximum; Kisch, 1991). For the fraction 8-63 mm the random mounts were used. The sand fraction (>63 mm) of i.r. was studied optically by means of light microscopy.

Ostracodes were separated and identified from 22 samples. In spite of small sample size (50-150 g in average) the material was representative for argillaceous limestones, but not for


JÕHVI Stage
KEILA Stage
?
OANDU St.
RAKVERE St.
424.6
421.6
413.4
409.9
406.7
401.1
Bed 1
Bed 2
Bed 3
Bed 4
Bed 5
B
B
Piretia erinacea
Hippula cetona norra
Trianguloschmidtella triangulata
Pedomphalella egregia
Unisulcopleura rakverensis?
Kuprimites? sp.
Sigmobolbina? sp.
Longiscula sp. ind.
Beotella sp. ind.
Uhakiella aff. oanduensis
Bichilina prima
Sigmoopsis rostrata
Hesperidella esthonica
Klimphores sp.
Medianella aff. longa
Consonopsis consona
Uhakiella sp.
Piretella sp.
Pullvillites rostratus
Scrobisylthis reticulatus
Cryptophyllus sp.
Kroemmelbeinia? sp.
Tetrada sp.
Sigmoopsis sp. ind.
Klimphores minimus
Pelecybolbina sp. ind.
Platybolbina temperata
Gen. n. sp.
Pelecybolbina pelecyoides
1
2
3
4
5
6
7
B
8
SWEDEN
ESTONIA
RISTIKÜLA
LATVIA
LITHUANIA
GOTLAND
ÖLAND
Baltic Sea

siltstones, that were poor in ostracodes. A traditional technique with sodium hyposulphite was used in sample preparation.

The Ristiküla-174 Core Section

Because of some difficulties in using traditional local lithostratigraphical units, the preliminary subdivision to beds 1 to 5 was used in the Ristiküla core description. The beds 1 to 5 correspond to the interval between the Big Bentonite Bed (bentonite "d") at the base of the Keila Stage and cryptocrystalline limestone of the Rakvere and Nabala stages (Fig. 1). The ostracode log of the section is presented in Fig. 1 and grain size distribution of i.r. in Fig. 2.

Description

Beds of the Jõhvi Stage underlying the Big Bentonite are mainly semi-nodular limestones (content of i.r. 15-25%).

Bed 1 (421.6-424.6 m). Semi-nodular argillaceous limestones (i.r. 24-54%, average 38.1% after 7 samples). The Big Bentonite Bed with thickness of 40 cm was taken for the base of the Bed 1. Another K-bentonite (bentonite "e") was recognized in the middle of the Bed 1. This interval belongs to the lowermost part of the Keila Stage.

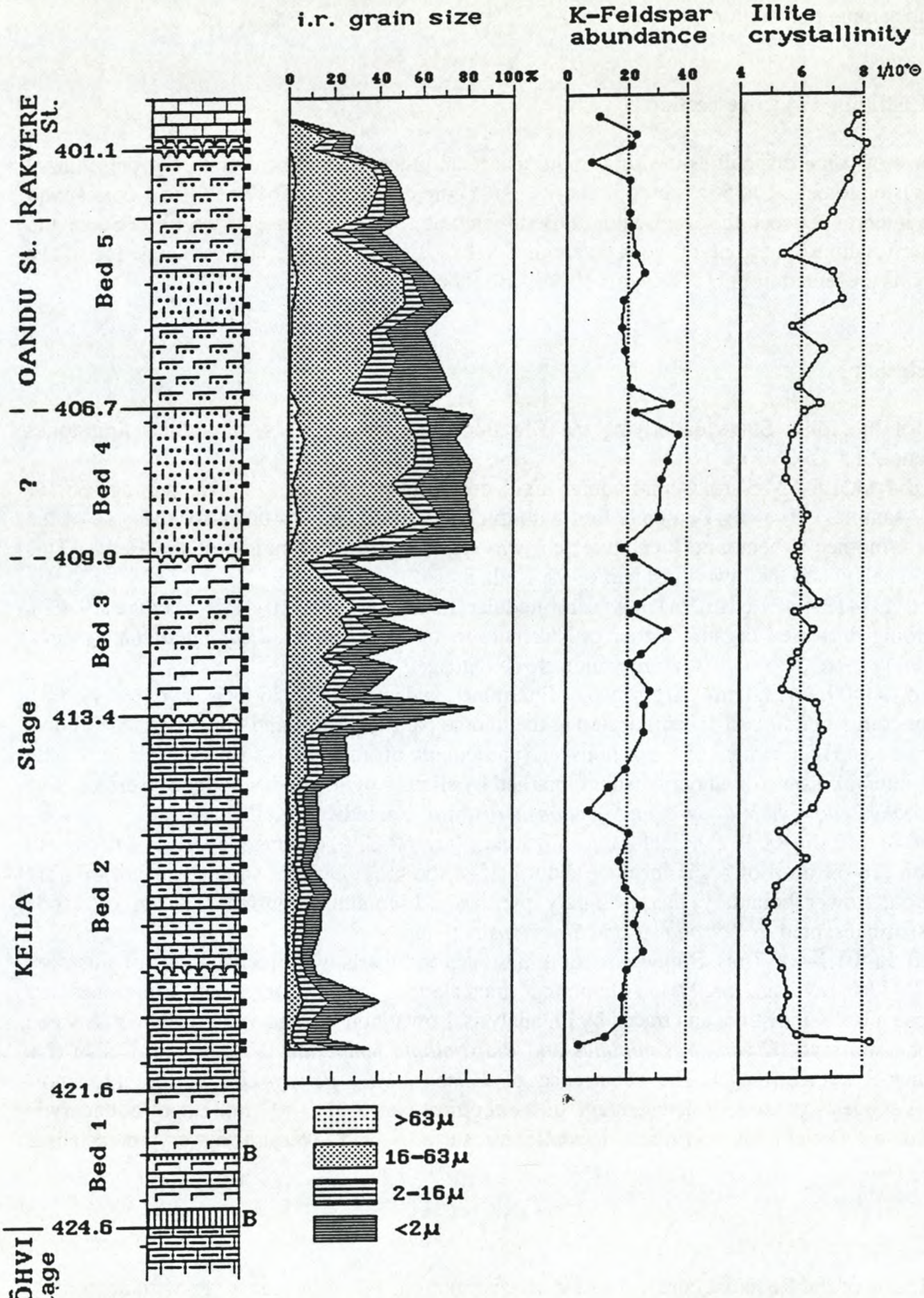
Bed 2 (413.4-421.6 m). Mainly semi-nodular limestones (i.r. 10-41%, average 19.4%), containing dispersed organic matter of kukersite in the middle part. *Pedomphalella egregia, Bichilina prima, Sigmoopsis rostrata* indicate Keilan age.

Bed 3 (409.9-413.4 m). Argillaceous limestones and marls (i.r. 26-82%, average 49.6%). Indistinct thin bedding (5-10 cm) occurs in the middle part, the lower and upper parts are almost monotonous. High content of terrigenous silt (particularly of fraction 16-63 mm) is characteristic to the interval. Lower boundary distinct, marked by slightly pyritized discontinuity surface. The ostracodes *Pedomphalella egregia, Sigmoopsis rostrata,* etc. indicate Keilan age.

Bed 4 (406.7-409.9 m). Dolomitic siltstones (i.r. 60-82%, average 76.9%). Coarse silt fraction (16-63 mm) of i.r. is forming about half of the rock. Distinct subvertical burrows are common. Lower boundary sharp, slightly pyritized (discontinuity surface?). Poor ostracode fauna is represented by *Tetrada* sp. and *Sigmoopsis* sp. ind.

Bed 5 (401.1-406.7 m). Silty dolomitic limestones and marls with intercalations of siltstones (i.r. 26-72%, average 53.9%). Alternating intercalations of siltstones and limestones are macroscopically indistinct and traced by i.r. analysis. Lower boundary is a sharp contact between siltstone and marl. *Klimphores minimus* and *Platybolbina temperata* indicate post-Keilan (i.e. Oanduan-Rakverean) age. The occurrence of *Pelecybolbina pelecyoides* in the uppermost sample suggests possible Rakverean age of the topmost part (401.1-402 m). Upper boundary is marked by two highly pyritized discontinuity surfaces and overlain by cryptocrystalline limestones.

Fig. 1. Log of the Ristiküla core and ostracode distribution. 1-3 - limestones (1 - with i.r. content <10%, 2 - i.r. 10-15%, 3 - i.r. 15-25%); 4 - argillaceous limestones (i.r. 25-50%) and marls (i.r. 50-75%); 5 - calcareous siltstones and silty limestones; 6 - siltstones; 7 - K-bentonites; 8 - discontinuity surfaces.



i.r. grain size
0 20 40 60 80 100%
K-Feldspar abundance
0 20 40
Illite crystallinity
4 6 8 1/10°Θ
RAKVERE St.
OANDU St.
?
KEILA Stage
JÕHVI Stage
401.1
406.7
409.9
413.4
421.6
424.6
Bed 5
Bed 4
Bed 3
Bed 2
Bed 1
B
B
>63μ
16-63μ
2-16μ
<2μ

Composition of Insoluble Residue

Põlma (1982) suggested that i.r. of Ordovician carbonate deposits in the lithofacially transitional zone has a coarser grain size (more rich in fraction 10-100 mm) than i.r. in Estonian and Central Baltoscandian confacies belts. Grain size distribution of i.r. in Bed 2 is close to that reported by Põlma (1982, p. 59) for the transitional zone. The average content of the fraction 16-63 mm in i.r. of Beds 3, 4 and 5 is 42%, 60% and 60%, respectively, and exceeds the transitional zone average 33% given by Põlma even for the i.r. fraction 10-100 mm.

In the upper, most terrigenous part of the section (401-410 m; Beds 4 and 5), it is possible to distinguish three sedimentary cycles according to i.r. grain size (Fig. 2): 401.1-402.8 m; 402.8-406.7 m; and 406.7-409.9 m. The grain size inside each of the cycles is increasing upward. Variation of this characteristic in Beds 1, 2 and 3 is irregular.

According to mineralogical composition of i.r. fraction >63 mm the Bed 2 clearly differs from Beds 3, 4, and 5. The fraction >63 mm of Bed 2 is mainly authigenic, consisting of silicified skeletal debris and pyrite, but in Bed 3 and particularly in Beds 4 and 5 it contains dominantly terrigenous angular quartz.

In the i.r. fraction 8-63 mm, quartz (80-95%), K-feldspar (5-20%) and pyrite (up to 15%) are the main minerals. The relative abundance of K-feldspar, calculated as the ratio of quartz 101 and K-feldspar 002 XRD peak intensities, shows enrichment of silty beds, particularly in Bed 4, by quartz (Fig. 2). The i.r. fraction 8-63 mm of Bed 2 is more rich in pyrite than of Beds 3 to 5 (analogous to the i.r. fraction >63 mm).

Clay mineral composition of the i. r. fraction <2 mm is very uniform in the sequence. Illite is prevailing clay mineral forming 75-90% of this fraction, chlorite content is 10-25%. Some samples, mainly in the upper part of Bed 5, contain kaolinite (up to 5% from the fraction <2 mm). Slight increase of relative abundance of chlorite and decrease of illite crystallinity in the upper part of Bed 5 can be observed (Fig. 2).

Discussion

The siltstones and silty marls described in the Ristiküla section in the interval 401.1-413.4 (Beds 3, 4 and 5) are rare rock types in the carbonate sequence of the Ordovician of Baltoscandia. The beds of siltstone or silty limestone are reported from Swedish islands Gotland and Öland approximately in the same stratigraphical position. Siltstone beds, 0.4-1.6 m thick, from the cores of Gotska Sandön and Gotland were referred to *Macrourus* Limestone (Thorslund 1958, Männil 1966) or Gräsgård Siltstone (Jaanusson & Mutvei 1982). Erratic boulders, occasionally with a siltstone succession up to 3 m, have been described in southern Öland by Andersson (1893), Martna (1955), Schallreuter (1977), Grahn (1981), and others. According to Jaanusson (Jaanusson & Mutvei 1982) the Gräsgård Siltstone beds in Gotland contain a fauna of Oanduan age, whereas boulders of the same lithostratigraphic unit on southern Öland are of Keilan and Oanduan age.

Fig. 2. The insoluble residue grain size distribution in the whole rock samples of Ristiküla core, relative abundance of K-feldspar in i.r. fraction 8-63 mm (calculated as the ratio of quartz 101 and K-feldspar 002 XRD peak intensities), and illite crystallinity in i.r. fraction <2 mm (determined from 001 XRD reflection as width at half-maximum). For lithological legend, see Fig. 1.

In southern Estonia, in the Keila and Oandu stages, beds of siltstone or silty limestone occur in several sections, e.g. in Kaagvere (Männil 1966) and Tartu borings (data of the authors) in SE Estonia and in Viljandi, Abja, and Häädemeeste borings in SW Estonia (Väärsi, Kajak *et al.* 1968, 1969, unpublished core documentation of Estonian Geological Survey). Like the Gräsgård Siltstone in Sweden, the silty beds in South Estonia include both the Keila and Oandu stages (Männil 1966). These beds seem to represent a distinctive lithofacies, extending from Swedish Baltic islands to South-East Estonia.

High content of terrigenous silt is also described in limestones of the Saku Member of the Vasalemma Formation in NW Estonia (Männil 1960, Põlma *et al.* 1988). Silt material (i.r. fraction 10-100 mm after Põlma) is concentrated in cross-bedded thin intercalations and lenses of siltstone or occurring as the main component of i.r. of limestones rich in skeletal debris and carbonate lithoclasts. The Saku Member with thickness up to 7 m represents a lithofacies of limited distribution in southern flank of bioclastic grainstone facies of the Vasalemma Formation and is of Oanduan age (Männil 1960, Männil & Meidla 1994).

The sections with available ostracode record closest to Ristiküla are Pärnu (25 km NW) and Abja (35 km E) core sections. The ostracode distribution in the Ristiküla section is generally in accordance with the Abja and Pärnu sections (Meidla in prep.). By ostracode correlation the Beds 2 and 3 of the Ristiküla section belong to the Keila Stage, the main part of the Bed 5 to the Oandu Stage and the uppermost part of the Bed 5 together with the overlying cryptocrystalline limestone to the Rakvere Stage. Bed 4 of the Ristiküla section has an unclear age and may correspond by its stratigraphic position to the gap between the Keila and Oandu stages in North Estonia, to a part of bioclastic grainstone bed (Vasalemma Formation) in NW Estonia or to the "shaly" member of the Mossen Formation in South Estonia. Thus, the silty beds in the Ristiküla section include the Keila, Oandu and Rakvere stages.

The boundary between the Keila and Oandu stages marks a major change in East Baltic region. According to Nestor (1990), the unification stage of basin development was changed to the differentiation stage with distinct facies zonation approximately at this time. The microcrystalline limestones rich in skeletal debris, formed in Kukruse to Keila time along the whole extent of the bathymetric profile, change to dominantly cryptocrystalline limestones with variable clay content, formed in Oandu to Porkuni time (Hints *et al.* 1989).

The facies differentiation began at the end of Keilan age. It coincided with a regression, which caused formation of gaps in the shallower shelf areas. Significant faunal changes (Hints et al. 1989) and remarkable lithostratigraphic unconformity (Männil & Meidla 1994, etc.) in the East Baltic region are mostly considered as evidence of these gaps in the East Baltic region. The formation of bioclastic grainstones and cross-bedded siltstones with lithoclasts in NW Estonia (Männil 1960, Põlma et al. 1988) and lithoclastic limestones in eastern Lithuania (Laškovas & Paškevicius 1989) is related to this gap. The occurrence of these sediments and the siltstone facies, described above, reflects the increasing water energy and supports the concept of regression. In terms of Põlma (1982) the distribution area of the siltstone facies belongs to the lithofacially transitional zone. It may give some indirect evidence of the extent of this regression.

Three cycles of terrigenous input of coarser quartz silt in the Ristiküla section (Beds 4 and 5) may be interpreted as three phases of regression: the maximum regression at the boundary between the Keila and Oandu stages (coinciding with the most important faunal changes) and two following phases. The relatively higher ratio of quartz and K-feldspar in siltstones of the maximum regression supports the suggestion about more intensive reworking during this episode, although shape of quartz grains remained to be angular.

The Keila-Oandu regression approximately coincided with changes of climate in the Baltic area. By earlier studies climate is considered to be turned more warm (Spjeldnæs 1961, Jaanusson 1973; Webby 1984) or more arid (Hints *et al.* 1989) in this period. Some characteristic features of the clay minerals may be related with climatic conditions in the terrigenous provenance area (Chamley 1989). The increase of chlorite abundance and decrease of illite crystallinity in the Bed 5 of Ristiküla section may indicate possible turn of climate to more warm and arid.

Conclusions

The Keila-Oandu age regression is well traced both in the Estonian Confacies Belt and in the lithofacially transitional zone, but had clearly different influence to sedimentation in these areas. If the regression caused formation of stratigraphic gaps and related high water-energy deposits in North Estonia, a sedimentation of well-sorted quartz silt took place in the transitional zone. Possible reworking of terrigenous material was more intensive during the maximum regression.

Acknowledgments

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ORGANIC BUILDUPS IN THE LOWER ORDOVICIAN OF THE ST. PETERSBURG REGION

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Abstract

Seven large organic buildups were found in the Lower Ordovician (Arenig) deposits of the eastern part of the Baltic-Ladoga Glint. Their average dimensions are approximately 3-4 metres in height and about 40-50 metres in length. The buildups form spectacular conical humps surrounded by haloes of echinoderm debris.

All the humps are of a similar construction. Two main facies types can be recognized in each of them; 1) clay core facies and 2) micritic crust facies. The boundary between the humps and enclosing sediments shows a kind of heterolithic unconformity.

The buildups can be regarded as a very special, so-called "Hecker-type" mud mounds. Atoll model has been proposed to explain their main features and the mode of their development.

Introduction

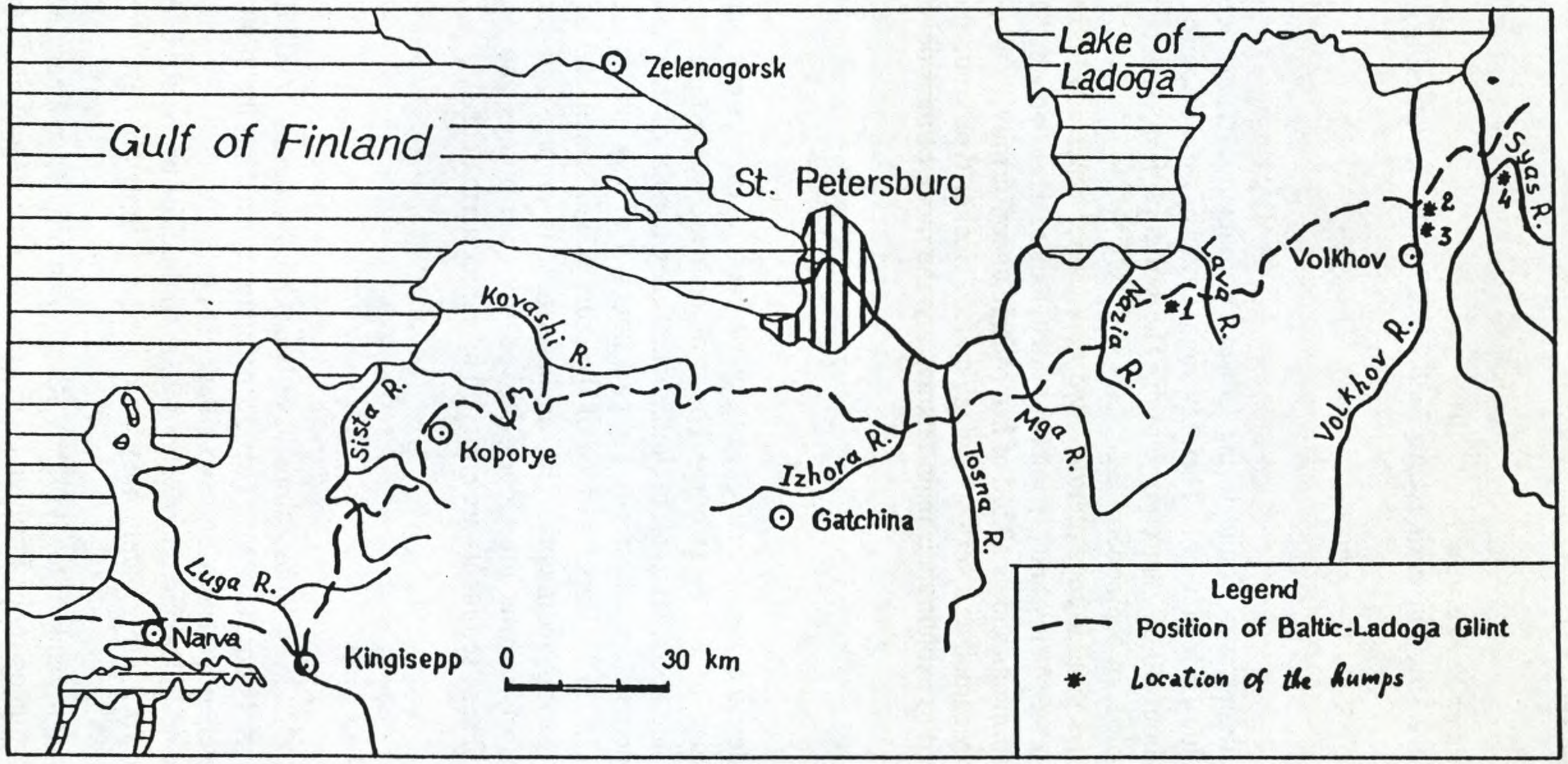
The present paper is devoted to unusual objects discovered in the eastern part of Baltic-Ladoga Glint to the East of St. Petersburg. These objects are synsedimentary uplifts or "humps" 3 to 4 m high and about 30 to 50 m in diameter. They consist of thin laminated clays with intercalations of bioclastic wackestones and packstones. The clay humps are covered by micrite crust 0.1 to 1.0 m thick. Upper surface of the crust makes some kind of heterolithic unconformity with enclosing sediments. The unconformity is emphasized by *Trypanites* borings and by pebbles of micritic limestones derived from the humps and redeposited in surrounding sediments. These sediments are enriched by echinoderm debris.

Geographical position

Since the summer of 1992 seven large humps have been discovered and studied (Dronov & Ivantsov 1994, Dronov & Fedorov 1994). Two of them are situated on the right bank of the Syas River between Kolchanovo and Montcevo villages (Fig. 1).

Two humps were found in Babino quarry about 2 km to the East of the River Volkhov not far from Old Ladoga town. Unfortunately, they were almost completely destroyed during the quarry exploitation.

One hump is on the right bank of the Volkhov River in an old quarry near the Simankovo Village. The others are situated in the Putilovo quarry 69 km to the East of St. Petersburg.


Lake of Ladoga
Zelenogorsk
Gulf of Finland
St. Petersburg
Kovashi R.
Sista R.
Koporye
Izhora R.
Gatchina
Tosna R.
Mga R.
Nazia R.
Lava R.
Volkhov
Volkhov R.
Syas R.
1
2
3
4
Luga R.
Narva
Kingisepp
0
30 km
Legend
Position of Baltic-Ladoga Glint
Location of the humps

Stratigraphy

All the humps are of the Volkhovian age. They began their development from the Steklo surface which coincides with the Latorp-Volkhov stages boundary. Tops of the Syas and Putilovo humps are enclosed to the sediments of B II subdivision. The top of the hump near Simankovo Village extends even into the sediments of the Kunda stage. In Putilovo hump the graptolites of the *Didymograptus hirundo* Zone (according to T.N. Koren) have been found in the core clays just about 40 cm above the Steklo surface.

Historical background

For the first time the existence of some disturbances in bedding structure in the very uniform Lower Ordovician succession was mentioned by V.V. Lamansky. He described rapid changes in lithology in one quarry along the Volkhov River and attributed them to the presence of a large synsedimentary fold (Lamansky 1905),

Later a special article was devoted to the intraformational breaks and disturbances in the Glauconite limestone of North-western Russia (Hecker & Vishniakov 1937). In this article the so-called "Syas hump" was first mentioned and described. The abrupt replacement of the BIIα limestones by the pure clays in the West part of the Putilovo quarry near the road to Antonovskoe Village was also mentioned. The "Syas hump" was interpreted by the authors as a large synsedimentary fold. That conclusion coincides with the previous interpretation of the disturbance on the Volkhov River (Lamansky 1905).

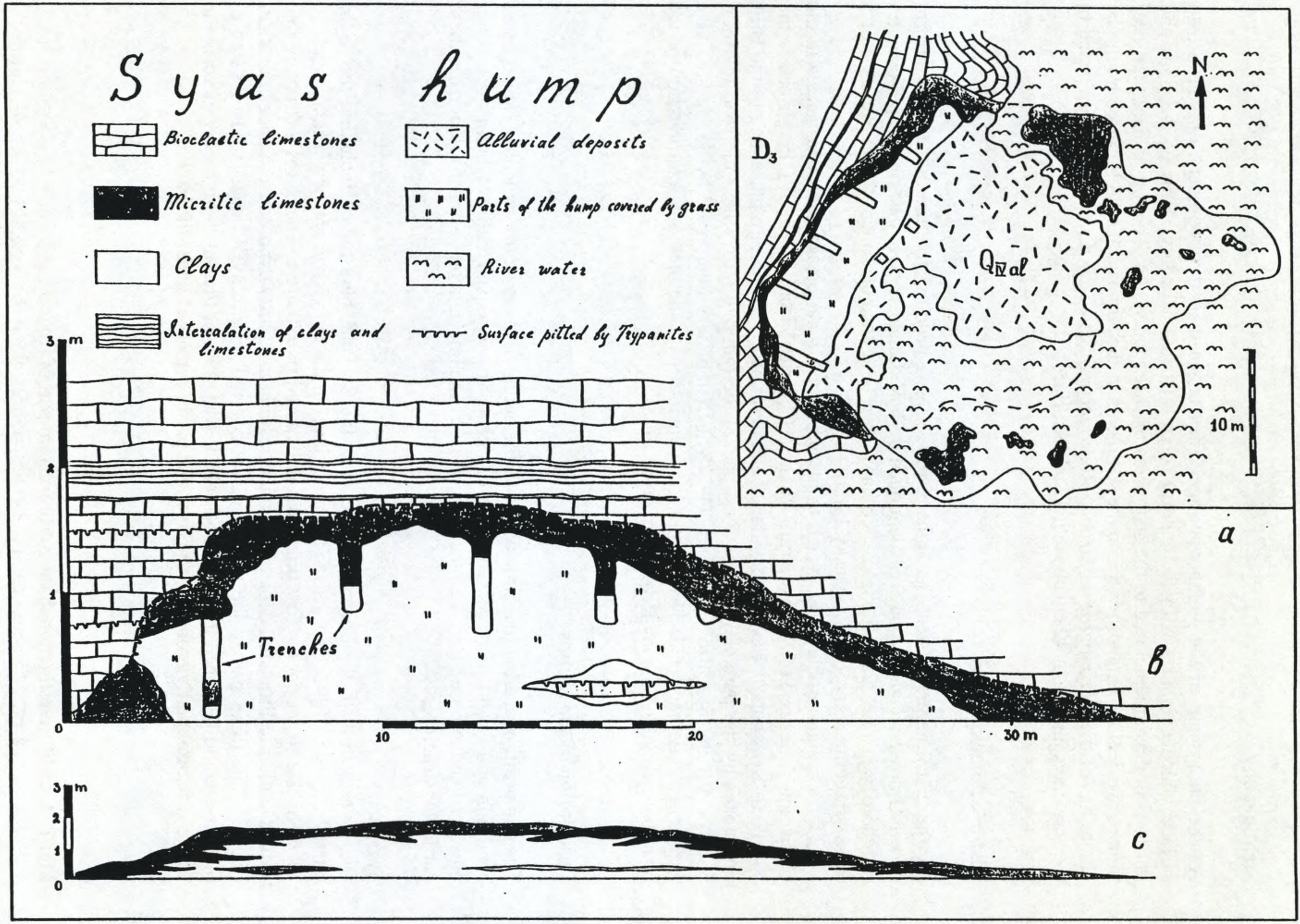
Description of the humps

The composition and internal structure of all the humps seem to be the same, but the exposure condition and extent of preservation are different. The Syas and Putilovo humps are better preserved and exposed than the others and most of the information has been taken from these particular structures.

Syas hump

Our investigation of the Lower Ordovician of the St. Petersburg region including the "Syas hump" started in 1992. This season the water level in the Syas River was lower than usual. So we were able to see for the first time parts of exposed river bottom. There was a structure that looked similar to little atoll depicted by micritic bioherms partly exposed from under the water (Fig. 2). The inner part of the stoll ring was covered with alluvial deposits but the line of bioherms could be traced to the hump itself. It was obvious that all these bioherms form the single structure, partly eroded by the river and partly preserved under the bank.

Fig. 1. Area of investigation in the north-western part of Russia.
1- Putilovo quarry, 2- Babino quarry, 3- Old quarries along the Volkhov River, 4- Humps in the Syas River Valley.


Syas hump
Bioclastic limestones
Micritic limestones
Clays
Intercalation of clays and limestones
Alluvial deposits
Parts of the hump covered by grass
River water
Surface pitted by Trypanites
N
D_3
$Q_{IV}al$
10 m
a
3 m
2
1
0
10
20
30 m
Trenches
в
3 m
2
1
0
c

The inner part of the structure was composed of red and yellow clays. There is no sign of folding or other disturbances in these clay deposits. In the core part of the Syas hump, even a thin laminated structure is preserved. Clays contain brachiopod shells, fragments of trilobite and bryozoans as well as echinoderm ossicles. There are also layers of bioclastic wackestones intercalated with clay deposits.

The hump is asymmetrical and its north-eastern slope is more gentle than the south-western. The outer surface of the hump is pitted by the *Trypanites* borings. Horizons of these borings can be traced on a distance of 20-50 m even in surrounding sediments.

Enclosing sediments are characterized by enrichment of echinoderm debris. Limestones close to the hump are biosparites and they contain less clay intercalations than average ones. Sometimes redeposited micritic pebbles can be found here.

Only the upper part of the Syas hump which is 1.60 m high is exposed in the river valley. The rest of the structure is hidden under the river bottom.

Putilovo humps

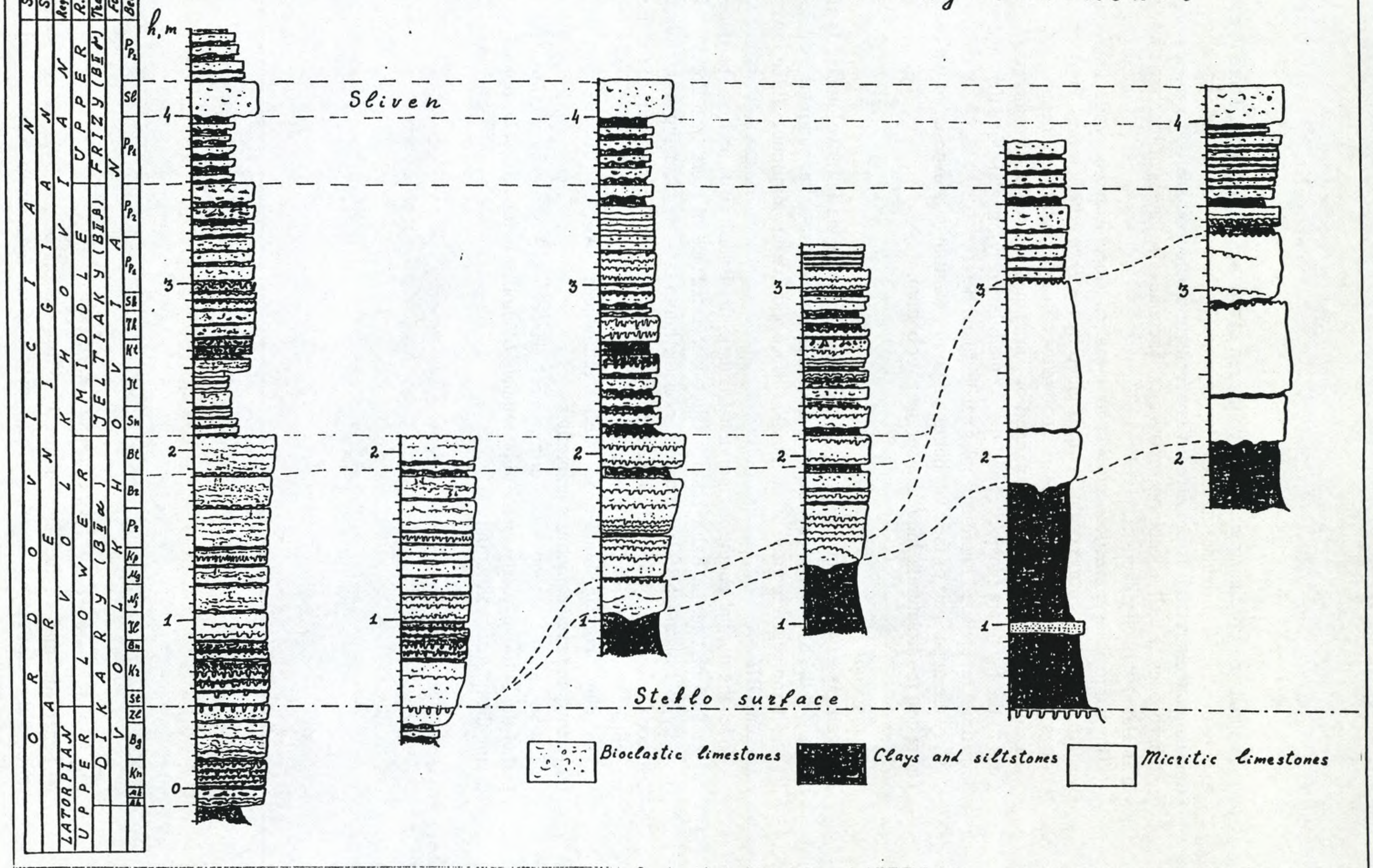
The well preserved hump is situated in the central part of the Putilovo quarry. In this case contrary to the Syas hump only the base, flanks and inner part of the hump are well seen (Fig. 3). The hump began to grow on the Steklo surface and extends minimum till the lower part of Frizy unit (BII).

Under the normal condition, individual beds in the Dikary (BII) unit can be traced along the Russian part of the Baltic-Ladoga Glint over a distance of more than 250 km (Dronov, Savitsky & Tsyganova 1993). But here, near the hump they are changing so rapidly that at a distance of about 20 or 30 m the beds are not recognizable. In the inner part of the hump they are replaced by the pure bluish-grey clays.

The clays contain brachiopod shells, echinoderm ossicles, trilobite and bryozoans fragments, and even sometimes graptolites.

The so-called "Antonovsckoe disturbance" (Hecker & Vishniakov 1937) in the western part of the Putilovo quarry seems to be remnant of another hump almost completely destroyed during the quarry exploitation.

Fig. 2. Preliminary results of the Syas hump investigation.
a- Schematic map of the Syas hump, b- Cross-section of the Syas hump. Vertical and horizontal scales are not equal, c- Cross-section of the Syas hump. Vertical and horizontal scales are equal.



Stratigraphic sections of the „Putilovo hump" and enclosing sediments

General Description

The main features of the humps can be summarized as follows:

1. The humps have average dimensions of about 3 to 4 m high and about 30-50 m in diameter.
2. The shape of the humps is isometric with flat base and conical surface. In a cross section they are often asymmetrical with some stairs-like intervals.
3. The boundary between micrite crust of the hump and enclosing sediments shows a kind of heterolithic unconformity. The surface of the humps is always pitted by *Trypanites* borings.
4. Two main facies can be recognized in each of the humps: 1) Clay core facies, and 2) Micritic crust facies.
5. There is some kind of sparitic halo or a sparitic flank facies around the humps. It seems possible that echinoderms lived on the hard surfaces of the micritic crust. The enrichment of the sediments by the echinoderm fragments can be regarded as an evidence of the existence of such a hump nearby.
6. The fauna distribution within the humps depends on facies. Only sparced fragments of brachiopod shells can be found in micritic crust facies. Core clay facies contain brachiopods, ostracods, bryozoans, trilobites and echinoderms. Even a graptolite remnants can be found here. Enclosing sediments are enriched by echioderm debris.

Genetic Interpretation

The humps are not synsedimentary folds. They are horizontal thin laminated structures in the core of the humps. They also are not remnants of the buried Latorp relief because they contain graptolites of the *Didymograptus hirundo* zone. This object can be classified as a sort of organic buildups.

There are no evidences of any kind of rigid framework in them. So one can conclude that these buildups belong to the mud mounds type. To distinguish them from other types a special name such as "Hecker-type mud mounds" is given to them. Professor R.F. Hecker was the first to discover and describe the Syas hump that can be considered as a model for all the buildups of this kind.

For explanation of all their features and mode of their development, atoll model has been proposed (Dronov & Ivantsov 1993). According to this model, mature Hecker-type mud mounds developed from the atoll-like structures appeared on a smooth surface of the sea bottom. The outer ring of the atoll was made from micritic bioherms. The inner lagoon was protected by this ring from the bottom currents and was filled with pure clays. From time to time strong storms brought bioclastic material to the lagoon forming wackestone lenses and layers.

Fig. 3. Stratigraphic sections of the Putilovo hump and enclosing sediments.

Mature hump was formed at the last stage of the atoll-like structure development by the backstepping of the outer micritic bioherms into the inner lagoon. The process takes place at the background of ongoing transgression. At the first stage a juvenile atoll structure appears on the smooth surface of the sea bottom. Then the outer bioherms grow up and the construction becomes higher. The inner lagoon discontinuity in the process when the hardground surfaces developed. At the last stage the inner lagoon completely disappeared and the structure became a real hump.

This mechanism seems to be the most probable explanation for the presence of the gratolite fragments and the thin laminated structure in the hump core clay facies. However, the reasons for the appearance of a large atoll-like structures on the smooth surface of the sea floor are still unclear.

It is interesting to note that the similar mechanism has been used for the explanation of the Middle Paleozoic Waulsortian type mud mounds development (Dronov 1993). Despite of differences in age, shape, lithology and sedimentary environment the same atoll model can be adopted for these two types of mud mounds. There is a possibility that it is not an accidental coincidence.

The nature of the organisms forming the buildups is also still unknown. First we thought that the reef-builders are calcareous sponges (Dronov & Ivantsov 1994). In thin sections of the micritic crust limestones some kind of short calcareous needles can be seen.

Another possible variant is an algae origin of the micritic bioherms. There are no identifiable remains of any algaes in micritic crust facies, but in June 1994 the micritic bioherm with stromatolitic-like thin laminae have been found in Volkhovian limestones of Western Estonia, near Paldiski. So now we can not exclude an algae origin of the micritic bioherms.

Embrionic buildups

The large mature Hecker-type mud mounds have been found only in the part of the Baltic-Ladoga Glint to the East of St. Petersburg. But in the western part of St. Petersburg region little structures that can be interpreted as juvenile forms of the Hecker-type mounds were discovered. One of these little humps is situated on the right bank of the Narva River just under Ivangorod fortress. It has a shape of a gentle fold 1 m long and 0.15 m high. The maximum thickness of clays in the core is about 0.07 m (compared to average 0.01-0.02 cm). Another little hump 3.5 m long and 0.25 m high was found in the Kingisepp quarry about 120 km to the West of St. Petersburg. Maximum thickness of the clays in the core reaches 0.20 m.

Both little humps developed in the Upper Latorpian deposits. There is a great similarity between these little humps and the sedimentary folds described by Lindstrom (1963) in the Latorpian sediments of Öland. But the Kingisepp and Narva structures seem to be different because of visible changes in lithology of the beds they consist of. The outer cover of the little humps reminds strongly of the micritic crust of the mature Hecker-type mud mounds.

There is a possibility that the little humps of the western part of St. Petersburg are juvenile forms of the Hecker-type mud mounds. They do not reach dimensions of mature mounds because they start to develop on a background of the progressing regression while the large humps of the eastern part of the Glint began to develop starting with the Volkhovian transgression.

Acknowledgments

We are grateful to Dr. T.N. Koren of All Russian Geological Institute (VSEGEI) for helpful criticisms and many language improvements to the manuscript. This investigation was funded by Grant 2-11-3-4 from the State Committee on Science and Higher Education of Russian Federation. We also thank Galina Gataulina for typing the drafts of the manuscript.

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SEQUENCE AND EVENT STRATIGRAPHY OF THE LOWER ORDOVICIAN OF THE ST. PETERSBURG - LADOGA REGION, RUSSIA

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The Lower Ordovician succession exposed in the St. Petersburg-Ladoga Region is subdivided into four Vail-type depositional sequences (Fig. 1).

Only the lower sequences have well developed sequence boundaries. The boundaries of the Kunda sequence that coincide with the unconformities at the base of the "Lower oolitic" and the "Upper oolitic" beds are easily identified only in the west part of the region. To the east their correlative conformities are not well recognizable. Lenses of the iron oolites occur both below and above boundary levels.

These four depositional sequences correspond to the four contemporaneous phases of relative sea-level changes that had been described in Canada (Barnes 1984). One can conclude that they are caused by the global eustasy.

Lower Ordovician succession near the Latorp-Volkhovian boundary allow more precise event-stratigraphic subdivision. For example, the lower Volkhov Formation (B IIα limestone) is traditionally subdivided in Putilov quarry into 15 individual beds (5-30 cm thick) (Fig. 2). Each of the beds has its own name and is characterized by individual diagnostic features. These beds are easily recognized in quarries and natural outcrops and can be traced over a distance of at least 250 km along the Russian part of the Baltic-Ladoga Escarpment.

The beds are interpreted as an effect from a short-term sea-level fluctuation possibly caused by the hypothetical Southern glaciation. They consist of elementary layers that seem to be proximal tempestites.

Thus, sections studied allow to establish the event-stratigraphic scale for the high-resolution correlation within the region.

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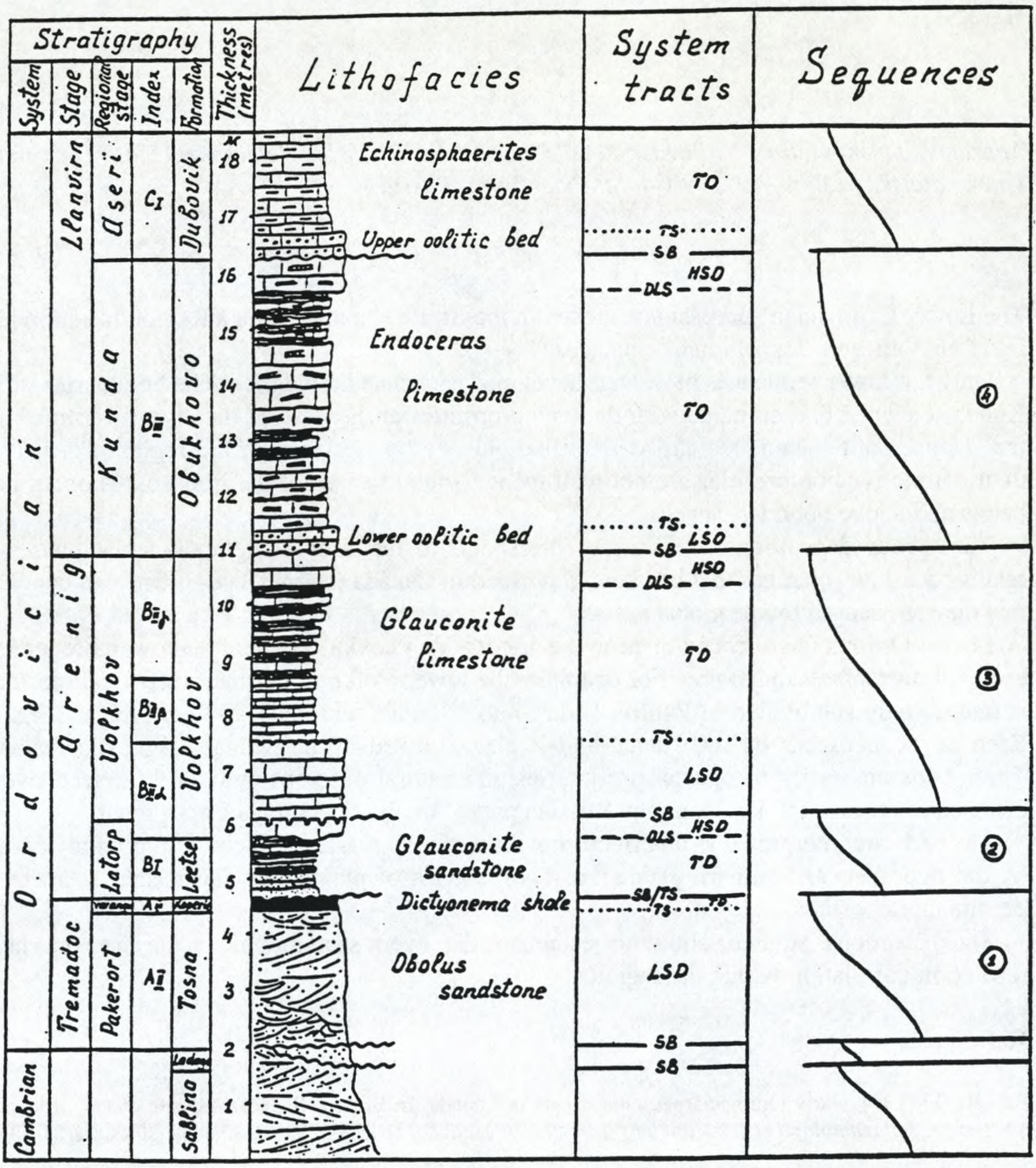

Fig. 1. Lithofacies, bio-, litho-, and sequence stratigraphy of the Lower Ordovician of the St.Petersburg Region, Russia. Symbols in the system tracts column are as follows: SB: Sequence boundary; TS: Transgressive surface; TD: Transgressive deposits; HSD: Highstand deposits; LSD: Lowstand deposits; DLS: Downlap surface; 1. The Pakerort-Varangu sequence; 2. The Latorp sequence; 3. The Volkhov sequence; 4. The Kunda sequence.

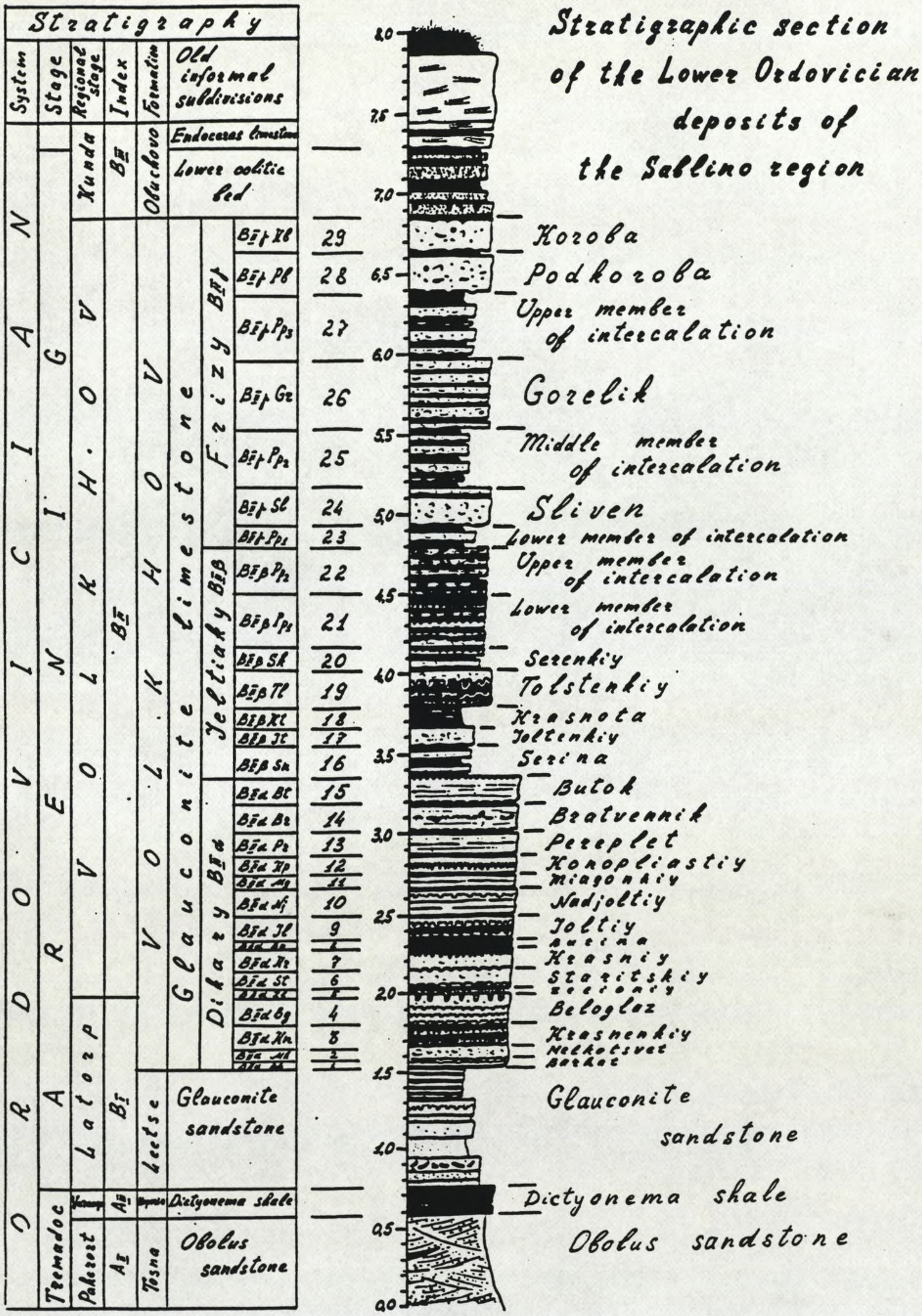

Fig. 2. Stratigraphic section of the Lower Ordovician deposits of the Sablino region showing the detailed event-stratigraphic subdivision.

BJØRKÅSHOLMEN FORMATION (UPPER TREMADOC) IN NORWAY: REGIONAL CORRELATION AND TRILOBITE DISTRIBUTION

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Introduction

The Bjørkåsholmen Formation (Upper Tremadoc), formerly the Ceratopyge limestone (3aγ), is present throughout Baltoscandia, and equivalents occur within the Parautochthon and Lower Allochthon of the Scandinavian Caledonides. It is characterized by its limited thickness, rich shelly fauna dominated by trilobites, and its broad regional distribution. Owen *et al.* (1990) defined the modern concepts of the Bjørkåsholmen Formation. The base is drawn at the abrupt change from the dark shale of the underlying Alum Shale Formation, to a light grey micritic limestone which consists of one or more planar or nodular limestone beds with a thickness of 5-30cm. Then follows an interval of alternating light and dark shale of varying thickness (5-15cm), with distinct dark limestone nodules (2-5cm) containing a fauna dominated by the olenid *Bienvillia angelini*. The remaining part comprises several beds (2-40cm), of micrite or fine- to medium-grained intrasparite with 1-5cm thick intercalations of shale. The topmost beds contain glauconite-like grains in the Vestfossen and Oslo-Asker districts, but this is not found further north in the Oslo Region (Bjørlykke 1974).

The object of this paper is to present the distribution of the Bjørkåsholmen Formation in Norway and to discuss briefly the trilobite distribution and some general depositional features.

Fossil preservation and the sediment

In rock thin sections fossils are seen to "float" in the matrix without preferred orientation. Cavities beneath curved tests have not been recognised. Trilobite exuvae are most common and are not much affected by compaction and diagenesis, although some intragranular compaction exits. Several relatively complete trilobite moults are known in association with the "free floating" disarticulated exuvae. Local accumulations of unsorted, mostly unfragmented, trilobite exuvae are also common. This mode of preservation usually indicates deep intrastratal bioturbation (Speyer 1987), where bioturbation is the main agent in homogenizing the sediment and with limited transportation by currents or waves.

In the basal beds of the Oslo-Asker district, fossiliferous septarian nodules are embedded in matrix containing a similar fauna. This suggests some early diagenesis and reworking of the sediment. In the upper limestone beds, brown-weathering dolomitic zones are a distinct feature. Some of these may represent discontinuity surfaces, caused either by pressure solution or locally developed bioclast accumulation. Stylolitic structures are common, but are not

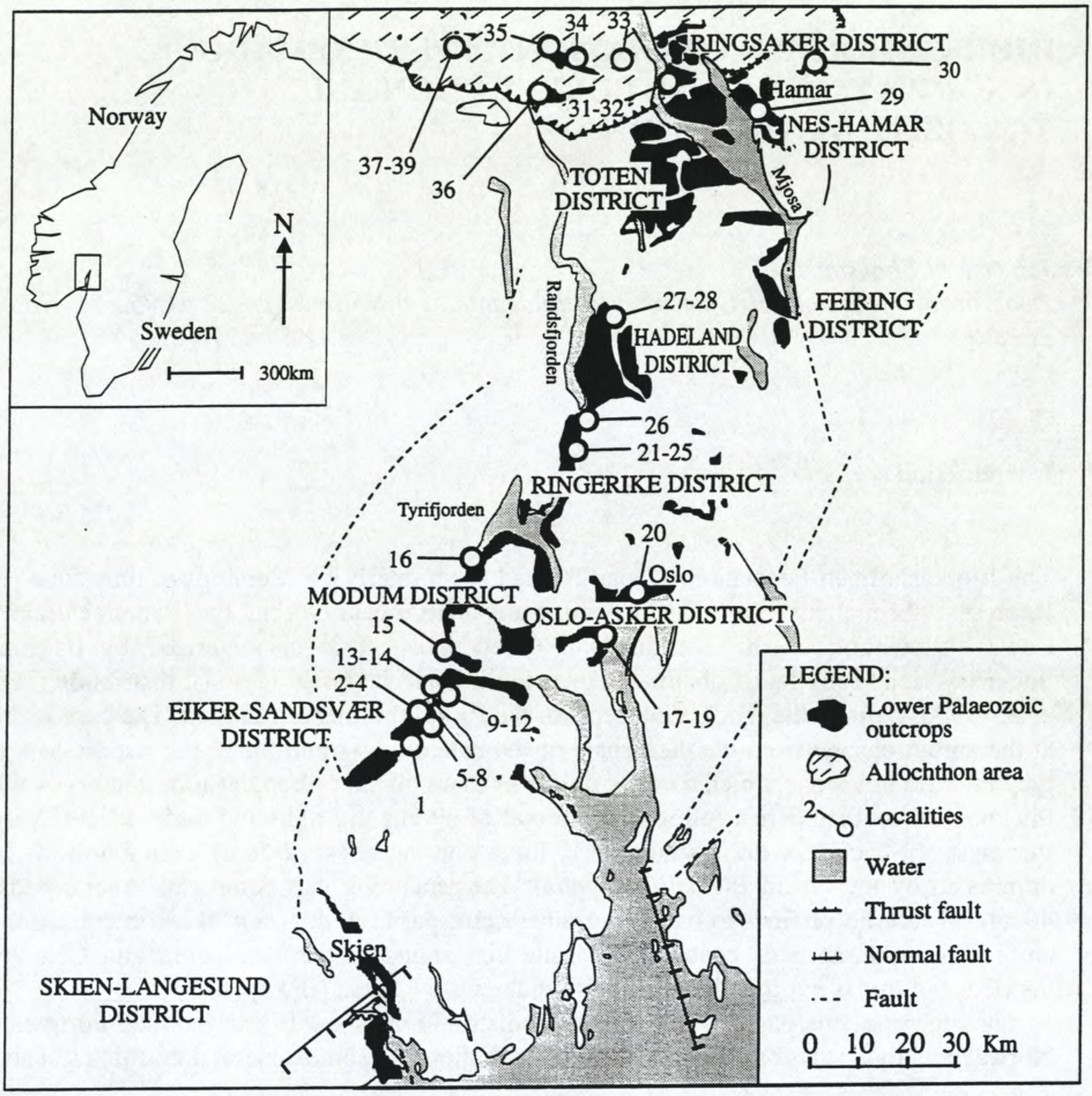

Figure 1. Map showing Lower Palaeozoic outcrops in the Oslo Region and part of the Lower Allochthon, Norway (after Størmer 1953). Numbers indicate localities of the Bjørkåsholmen Formation: **Eiker-Sandsvær district:** 1-Hvamsetrene, Krekling (NM 4530 1284); 2-Teigen, Krekling (NM 420 127); 3-Stavlum, Fiskum (NM 4425 1629); 4-Klunderud, Fiskum (NM 4476 1647); 5-8-Bakstevalås, Eikern (NM 4840 1622, NM 484 163, NM 483 164, NM 480 164, all metamorphosed); 9-Lunden, Vestfossen (NM 4842 2132); 10-Smellhaugen, Vestfossen (NM 4867 2196); 11-Vestfossen railway station., Vestfossen (NM 4872 2217); 12-Skara, Vestfossen (NM 4785 2238); 13-Skarahaugen, Vestfossen (NM 4717 2225); 14-Hals, Vestfossen, hypostratotype section (NM 4828 2300). **Modum district**: 15-Hornsrudveien, Dramdal (NM 5177 3417); 16-Øvre Øren, Vikersund, hypostratotype section (NM 5763 4442). **Oslo-Asker district:** 17-Prestenga bus stop (NM 8303 2724); 18-Bjørkåsholmen, Slemmestad, basal stratotype (NM 8442 2938); 19-Hagastrand, Asker, hypostratotype section (NM 8420 2980); 20-Væ kerø, Lysaker, hypostratotype section (NM 925 432); additional 21 localities, not listed. **Ringerike district:** 21-Hvalsveien , Klekken (NM 7336 7106); 22-Vang school, Haug, hypostratotype section (NM 7344 7166); 23-Færdenveien, Klekken (NM 7416 7143); 24-Andersløkka, Haugsbygd (NM 7450 7093); 25-Ringkollveien, Haugsbygd (NM 7490 7108, NM 7504 7108); 26-Sitpåveien, Ask (NM 7598 7601). **Hadeland district:** 27-Nedre Hvattum, Gran (NM 8791 9382); 28-Tuv, Gran (NM 863 946 or 863 948, based on Holtedahl & Schetelig 1923). **Nes-Hamar district:** 29-Stor Hamar strand, Hamar (PN 1090 4149); 30-Løiten Brenneri, Løten (PN 255 451). **Ringsaker district:** 31-Steinsodden, Ringsaker (NN 9185 5368); 32-Tømten, Ringsaker (NN 9365 5468). **Lower Allochthon:** 33-Skogheim, Snertingdalen (NN 8048 5075); 34-Lunden, Snertingdalen (NN 6829 5481); 35-Røste, Aust Torpa (NN 6145 4617); 36-Kolsrud, Austinni (NN 593 479); 37-Lødnøygarden, Bruflat (NN 38 53); 38-Nystøltjern, Bruflat (NN 374 549); 39-Granheim, Bruflat (NN 342 624); 40-Grøsli, Synnfjell (NN 173 703). Latter locality to the west of the map area showed above.

delimiting the dolomitic zones. Secondary growth of calcite perpendicular to the fossils is common, and pyrite framboids or larger crystals are found throughout.

Regional Correlations in the Oslo Region and Adjacent Areas

Some 60 outcrops of the Bjørkåsholmen Formation in Norway were registered during this study. Fig. 1 shows the distribution of localities in the different districts of the Oslo Region and the Lower Allochthon in Norway. In the Oslo-Asker district outcrops are common, whilst they are scarce in the remaining districts. The latter outcrops are therefore named in the following discussion. The UTM grid-references refer to topographical sheets 1:50 000 (M-711 series), and economic maps on the scale of 1:5000 or 1:10 000.

Eiker-Sandsvær district. 14 outcrops; Hvamsetrene, Krekling (NM 4530 1284); Teigen, Krekling (NM 420 127); Stavlum, Fiskum (NM 4425 1629); Klunderud, Fiskum (NM 4476 1647); Bakstevalås, Eikern (NM 484 162, NM 484 163, NM 483 164, NM 480 164); Lunden, Vestfossen (NM 4842 2132); Smellhaugen, Vestfossen (NM 4867 2196); Vestfossen railway station (NM 4872 2217); Skara, Vestfossen (NM 4785 2238); Skarahaugen, Vestfossen (NM 4717 2225); Hals, Vestfossen, (NM 4828 2300).

A hypostratotype section is at Hals, Vestfossen (Klemm 1982). One basal limestone bed is present. The upper part consists of 5-30cm thick limestone beds with 1-2cm shale partings. The top is defined by a 12-20cm thick glauconitic limestone. The total thickness is 55-85cm. At Vestfossen railway station, Vestfossen, the 78cm thick section yields fossils of the *Ceratopyge* community in the lower 50cm.

Modum district. 2 outcrops; Hornsrudveien, Dramdal (NM 5177 3417); Øvre Øren, Vikersund (NM 5763 4442).

A hypostratotype section is at Øvre Øren, Vikersund (Owen *et al.* 1990). The unit here is similar to that in the Eiker-Sandsvær district, but the glauconitic limestone is not developed. Wandås (1982) reported two basal nodular limestone beds, and a total thickness of 130cm at Øvre Øren. In this study the section was measured to be 118cm with one basal bed. The section yields fossils of the *Ceratopyge* community in the lower 61cm. The upper 10-15cm of this unit however, yield a trilobite fauna of the *armata/planilimbata* trilobite zone(s).

Oslo-Asker district. 24 outcrops

The basal stratotype section for this formation is at Bjørkåsholmen, Asker (NM 8442 2938), with hypostratotype sections at Hagastrand, Asker (NM 8420 2980), and Vækerø, Lysaker (NM 925 432)(Owen *et al.* 1990, Figs 3, 4, 5). One or two basal limestone beds are present. The upper part of the unit has a distinct 10-15cm shale unit separating the limestone above and below. The limestone beds are 5-20cm thick, and glauconite is present in the upper part. The total thickness is 90-120cm. In the 95cm thick unit at Prestenga bus stop, Slemmestad (NM 8303 2724), fossils of the *Ceratopyge* community are present in the lower 58cm.

Ringerike district; 6 outcrops; Hvalsveien, Klekken (NM 7336 7106); Vang school, Haug (NM 7344 7166); Færdenveien, Klekken (NM 7416 7143); Andersløkka, Haugsbygd (NM 7450 7093); Ringkollveien, Haugsbygd (NM 7490 7108, NM 7504 7108); Sitpåveien, Ask (NM 7598 7601). (2)

A hypostratotype section at Vang school, Haug (Owen *et al.* 1990), is almost covered. The basal limestone is developed as large elongated nodules, 5-10cm x 10-50cm in size, parted from the succeeding dark limestone nodules by some 10cm of shale. Above this is a 30-40cm thick compact limestone, followed by one or two 10cm thick limestone beds separated by 1-6cm of shale. Glauconite is not recognized. The total thickness is 85-95cm. At Færdenveien fossils of the *Ceratopyge* fauna are present in the lower 80cm of the 95cm thick section.

Hadeland district. 2 outcrops; Nedre Hvattum, Gran (NM 8791 9382); Tuv, Gran (NM 863 946 or 863 948). The latter data are based on Holtedahl & Schetelig (1923). (2)

The basal layer is 10-15cm thick. It is succeeded by 50cm of shale with light grey limestone nodules just below the horizon of dark limestone nodules. The succeeding compact unit is 20cm thick with a 2-3cm thick layer of limestone nodules at the base. It is followed by three 10cm thick limestone beds with 2-5cm shale partings. The total thickness is 75-115cm. At Nedre Hvattum, Gran, fossils of the *Ceratopyge* community are present in the lower 103cm of the 115cm thick section.

Nes-Hamar district. 2 outcrops; Storhamar Strand Hamar (PN 1090 4149); Løiten Brenneri, Løten (PN 255 451). (2)

The basal unit is composed of small limestone nodules, separated from the dark limestone nodules by 20cm of shale. Then follows a 50cm thick, compact unit made up of four distinct layers. The total thickness is at least 70cm.

Ringsaker district. 2 outcrops; Steinsodden, Ringsaker (NN 9185 5368); Tømten, Ringsaker (NN 9365 5468). (2)

In this district two or three basal beds are present, the upper two beds being nodular. The upper part of the formation consists of several 10cm thick limestone beds, followed by a 30-40cm thick, compact limestone unit with calcite at the top. The total thickness is 130cm. In the section at Steinsodden, Ringsaker, fossils of the *Ceratopyge* community are present in the lower 100-110cm (Fjelldal 1966).

The Lower Allochthon

Outcrops of the Lower Allochthon are important in that they contain rocks with facies representing the westernmost extension of the formation. The following areas are important:

a) Snertingdalen. 2 outcrops; Skogheim, Snertingdalen (NN 8048 5075); Lunden, Snertingdalen (NN 6829 5481). (3)

One or more basal limestone/nodular limestone beds are present, with the dark limestone nodules 18cm above the base. The upper limestone beds, with 2-5cm shale partings, are thickest at the top of the unit (30-35cm). The total thickness is 80-110cm.

b) Aust Torpa. 1 outcrop; Røste, Aust Torpa (NN 6145 4617). (3)

Three basal limestone beds, partly nodular, are present. The small, dark limestone nodules 50cm above the base are followed by two upper limestone beds, each about 18cm thick with 5cm shale inbetween. The section is 91cm thick, and fossils of the *Ceratopyge* community are present in the lower 80cm.

c) Dokka. 1 outcrop; Kolsrud, Austinni (NN 593 479). (3)

Bjørlykke (1979) found grey shale with limestone nodules thought to belong to the Bjørkåsholmen Formation.

d) Bruflat. 3 outcrops; Nystøltjern (NN 374 549); Granheim,(NN 342 624); Lødnøygarden (NN 38 53). (3)

Strand (1954) recorded limestone nodules with elements of the *Ceratopyge* fauna, thought to belong to the upper part of the Alum Shale Formation, formerly the Ceratopyge shale (3aß).

e) Synnfjell and Valdres thrust sheets. (3)

At Grosli in Synnfjell (NN 173 703), a limestone unit within the Solheim Slate Member of the Ørneberget Formation is probably equvivalent to the Bjørkåsholmen Formation in the Oslo Region (Bruton *et al.* 1984, Bruton *et al.* 1989). Twelve sections chosen from eight districts of the Oslo Region and adjacent areas discussed above, are correlated in figure 2.

Trilobite Distribution

The trilobite community of the Bjørkåsholmen Formation comprises 26 genera and 23 species. The most common, in order, are *Ceratopyge acicularis* (Boeck) (previously *C. forficula acicularis* (Boeck)), *Symphysurus angustatus* (Boeck), *Apatokephalus serratus* (Boeck), *Euloma ornatum* Angelin and *Niobe* ssp.. The fauna is remarkably consistent across the Baltic shield, in contrast to earlier held views suggesting a higher distribution frequency in the Oslo-Asker district. This must however, be regarded as an artifact of the high number of outcrops available in this district. The number of trilobites decreases steadily upwards in the unit.

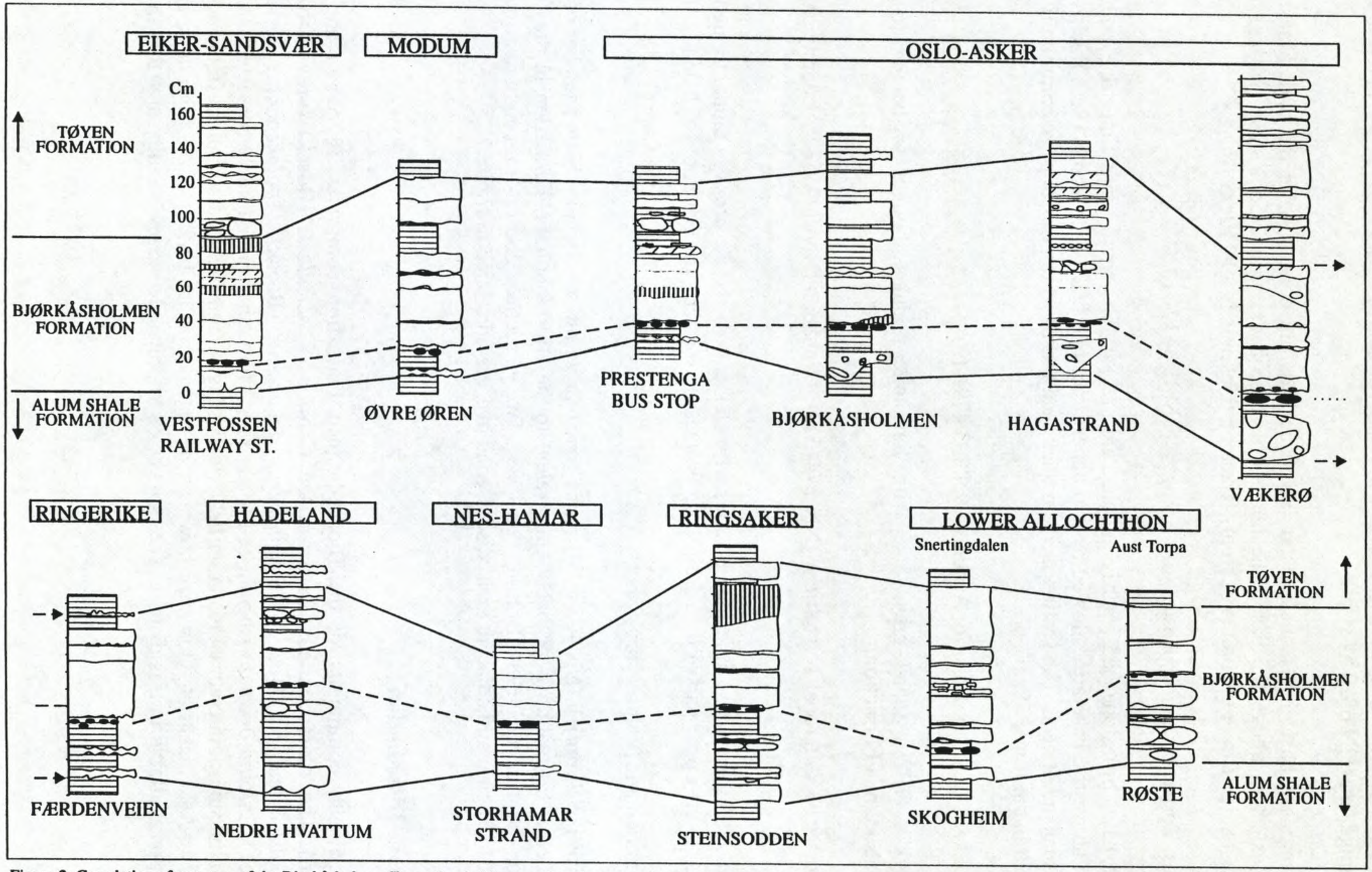

Figure 2: Correlation of outcrops of the Bjørkåsholmen Formation in Norway. See the text and figure 1 for further details.
Legend: ☐ Limestone. ● Dark limestone nodules. ☰ Shale. ▥ Calcite. ▨ Glauconite. —— Level of lower and upper boundary. – – Level of dark limestone nodules.

An outline of the vertical distribution is given here based on preliminary data from six outcrops chosen from different district: Eiker-Sandsvær: Vestfossen railway station (NM 4872 2217), Modum: Øvre Øren, Vikersund (NM 5763 4442), Oslo-Asker: Prestenga bus stop, Slemmestad (NM 8303 2724), Hadeland: Nedre Hvattum, Gran (NM 8791 9382), Ringsaker: Steinsodden, Ringsaker (NN 9185 5368), Lower Allochthon: Røste, Aust Torpa (NN 6145 4617).

The sample density is high, with sampled intervals of 10cm thickness or less which often corresponds to the thickness of individual beds. Information for the diverse, but sparse fauna of the dark limestone nodules is not given, and data from the lower limestone beds show no significant pattern in any of the investigated districts. Above this however, in the first succeeding light grey limestone beds, *Ceratopyge acicularis* dominates. Therafter its numbers decrease rapidly, accompanied by a small relative increase in the number of specimens of *Apatokephalus serratus*, followed by an acme of *Euloma ornatum*. There is a steady increase in the number of *Symphysurus angustatus*, which reaches an acme after that of *E. ornatum*. Species of *Niobe* are common throughout the unit, but show a relative increase in abundance near the top in most of the investigated outcrops.

Discussion

The reworked basal limestone beds in the Oslo-Asker district and the many basal limestone beds to the north of the region, all containing the same faunal elements, represent the initial widespread lime mud deposition of the Bjørkåsholmen Formation. However, based on the different development in different districts of these beds, the event is probably diachronous. The ensuing dark limestone nodule development represents an anaerobic event with a trilobite fauna different from the *Ceratopyge* community below and above. The upper light grey limestone beds vary in number, appearance and thickness, but the *Ceratopyge* trilobite fauna exhibits the same distribution pattern throughout, though its appearance is at different levels in different districts. This indicates that some of the limestone beds are condensed and diachronous. Short-lived episodes of shale incursions represent times of low oxygen environment, whilst longer periods of oxygenated lime mud sedimentation suggest fluctuations in depth, local palaeotopography or local environment conditions. These episodes are reflected in the trilobite faunal logs.

Acknowledgment

This paper is part of a larger study on the Bjørkåsholmen Formation and its trilobite fauna, prepared for a Cand. scient. thesis at the Palaeontological Museum in Oslo, Norway. I therefore wish to express my sincere gratitude to my supervisor Professor David. L. Bruton, for his encouragement and enthusiastic support during my study.

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Phosphatic microbrachiopod biofacies in the Lower Allochthon of the Scandinavian Caledonides

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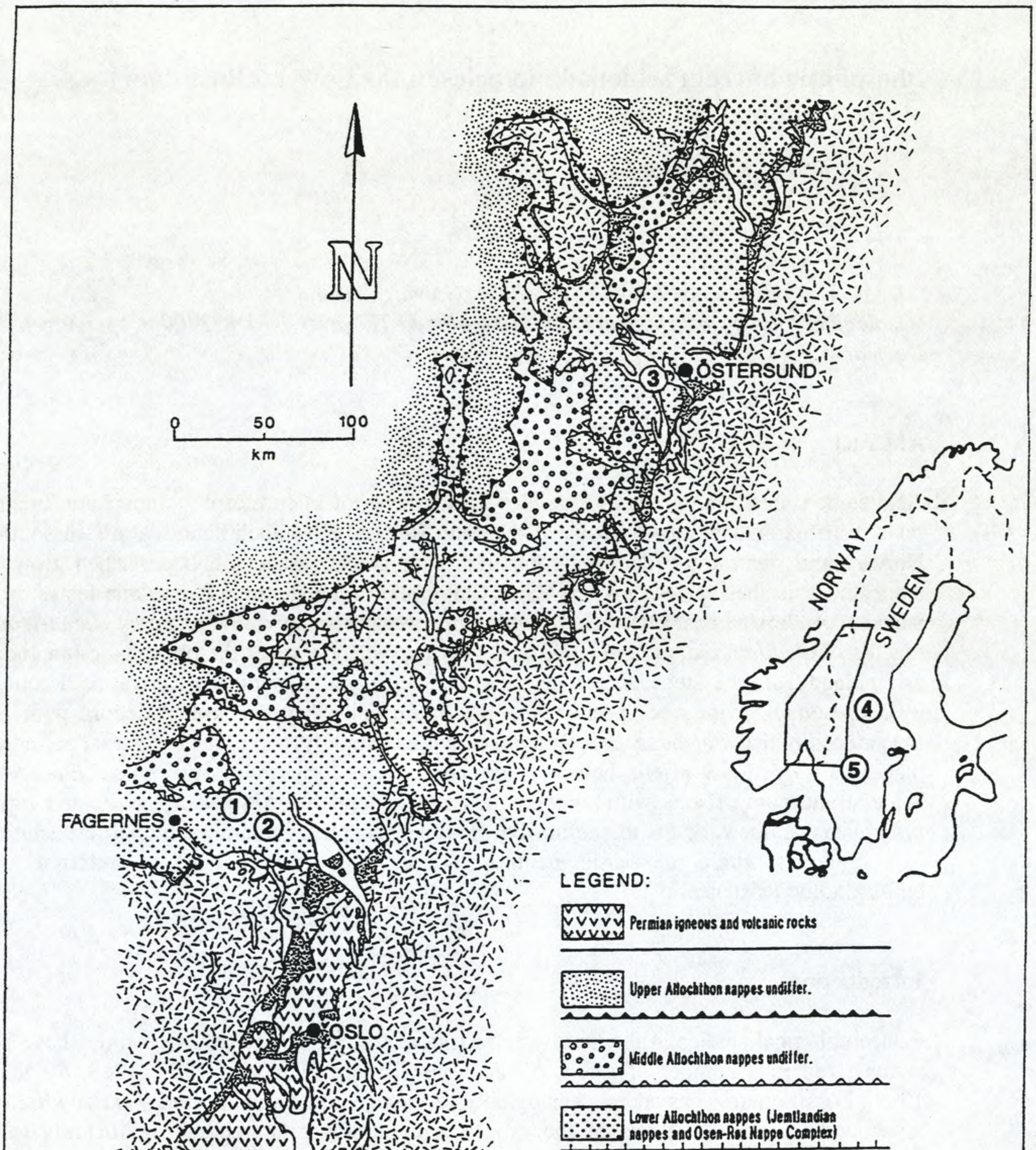
Abstract

Acid-etched, late Arenig - early Llanvirn assemblages of micromorphic phosphatic brachiopods from parautochthonous rocks in the Oslo Region and allochthonous units in southern Norway and central Sweden are compared with coeval associations described from the autochthonous shelf facies of the Swedish Baltic platform. The Stein Limestone facies on the island of Andersön, near Östersund, Jämtland is overwhelmingly dominated by *Numericoma*? together with *Biernatia*, *Scaphelasma* and *Torynelasma*. However the phosphatic brachiopod assemblage from the Stein Limestone facies in Snertingdalen, southern Norway is dominated by *Scaphelasma* with rare specimens of *Biernatia* and *Myotreta* and co-occurs with *Protopanderodus* and *Periodon* conodont associations. Slightly older, latest Tremadoc, assemblages at Gröslii, Synfjell, however are characterized by poorly-preserved *Eurytreta.* Multivariate comparisons with contemporary assemblages from Sweden and Poland form a preliminary framework for the definition of some phosphatic microbrachiopod associations. Life strategies within the sessile benthos as epi- or interstitial fauna is preferred to an epiplanktonic existence.

Introduction

Palaeontological studies in the Ordovician rocks of the Scandinavian Caledonides have been summarized in a number of recent publications (Bruton & Harper 1985, 1988, Spjeldnæs 1985). Fossil control has added considerably to the age constraints on many units within the lower and upper allochthons of the orogen and moreover has added significantly to the continuing debate concerning the provenance and movements of terranes defined within these major tectonic units now resting on cratonic Baltica. To date, conodont studies have dominated analyses of the microfauna providing the basis for biostratigraphical correlation in many parts of the Caledonian mountain belt; moreover the use of the group in biogeographical and thermal maturation studies is well advanced. More recently recognition of ecological associations within the conodont biofacies has further refined palaeogeographic reconstructions of palaeoenvironments and terranes, around the evolving Iapetus Ocean, now trapped within the Appalachian-Caledonian orogen. Nevertheless, there have been few records of phosphatic microbrachiopods from the Baltoscandian mountain belt although these acid-resistant fossils are common in conodont-rich residues (Bruton *et al.* 1989). Middle Ordovician phosphatic microbrachiopods have now been documented in extravagant detail from the platform rocks of the Baltic province (Holmer 1989). This contribution is a summary of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LEGEND:
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Middle Allochthon nappes undiffer.
Lower Allochthon nappes (Jemtlandian nappes and Osen-Røa Nappe Complex)
Cambro-Silurian sediments
Precambrian crystalline rocks
Autochthonous and Parautochthonous

research to date tracking microbrachiopod biofacies from the carbonate platform of Baltoscandia out into the lower allochthons of the Caledonide orogen.

New data from the Scandinavian Caledonides

Three sections within the Lower Allochthon of Norway and Sweden have been investigated and sampled in detail; in particular sequences across the Arenig - Llanvirn boundary at Andersön (Jämtland, Sweden) together with Snertingdalen and an uppermost Tremadoc section at Gröslii (East Jotunheimen, Norway) were targetted for conodont faunas (Fig. 1). The phosphatic microbrachiopods were retrieved from these acid-etched residues often rich in conodonts. The phosphatic brachiopods at the Arenig-Llanvirn boundary are from the Stein Formation (Rasmussen & Bruton 1994) and its lateral equivalents. At its type area in the Ringsaker belt of the northern Oslo Region, the formation is mainly composed of limestone (Størmer 1953); however westwards, in the Caledonides, the unit thins and becomes more argillaceous.

The Stein Formation at Andersön, Jämtland, Sweden occurs in one of the lowest limestone-dominated thrust sheets within the Jemtlandian Nappes (Gee *et al.* 1985). The Stein Limestone comprises about 40 m of lime mud and wackestone which correlates with the Volkov and Kunda stages (upper Arenig-lower Llanvirn equivalents). Some 30 km to the east autochthonous limestones crop out whereas the Stein Limestone is structurally overlain by the turbidites of the Föllinge Greywackes, farther west. The brachiopod fauna is dominated by the distinctive acrotretide *Numericoma*? although slightly rarer specimens of *Biernatia*, *Scaphelasma* and *Torynelasma* together with an indeterminate taxon also occur.

The Stein Formation at Snertingdalen, southern Norway occurs within the Osen nappe complex previously assigned to the Aurdal duplex (Bruton *et al.* 1989). At Haugnes the Stein Formation is developed as grey, impure, reticulate limestones; the section is tectonised and the formation probably exceeds 7 m in thickness. The conodont fauna is dominated by *Protopanderodus* and *Periodon*, the latter is uncommon in the Huk Formation of the Oslo Region and rare or absent from proximal localities on the Baltoscandian shelf. The *Protopanderodus* - *Periodon* Biofacies is typical of the outer shelf environment in Baltoscandia and the proximal slope in western Newfoundland (Rasmussen & Stouge 1995). The microbrachiopod fauna is overwhelmingly dominated by *Scaphelasma* with rare specimens of *Biernatia* and *Myotreta*.

The Solheim Slate is in part laterally equivalent to the Stein Formation and crops out at Gröslii, Valdres, southern Norway. The slate forms part of the Synfjell duplex or thrust sheet (Hossack *et al.* 1985, Nickelsen *et al.* 1985) and thus structurally overlies the thrust sheets of the Osen complex. This unit is probably condensed, ranging in age from Tremadoc to Llanvirn, and contains a variety of macrofossils and conodonts (Bruton *et al.* 1989); the unit is defined as the upper member of the Ørnberget Formation. The Solheim Slate is exposed at Grösliiseter where it comprises a variable sequence of light grey-weathering shales, calcareous slates and argillaceous, commonly nodular, limestones.

Fig. 1. General map of Scandinavia with outline of main thrust units and sampled localities in the Scandinavian Lower Allochthon. Locality 1: Gröslii, 2: Haugnes, Snertingdalen, 3: Andersöen, 4: Dalarna, and 5: Västergötland.

Phosphatic brachiopods were rare in the acid-etched residues however the samples are dominated by both dorsal and ventral valves of *Eurytreta.* Despite a few indeterminate fragments the fauna is of very low diversity.

Conodont assemblages

The 33 investigated samples from the Andersön section contained in total 8536 conodont elements. The section correlates with the *B. navis*, *P. originalis*, *M. parva*, *L. variabilis* and *E. suecicus* biozones. The fauna is dominated by *Protopanderodus* (38 %) and *Periodon* (19 %), while *Baltoniodus* approximates to 13 %. Three samples were very rich in phosphatic brachiopods of which two samples are from the Arenig *B. navis* Zone (GCI99455 and GCI99457) whereas one sample correlates with the upper part of the *L. variabilis - M. parva* Subzone (GCI99474), and thus is of early Llanvirn age. A basal Llanvirn brachiopod-rich sample from the Haugnes section contained 381 specimens of which 43 % are *Protopanderodus*, 21 % *Periodon* and 10 % *Baltoniodus*. No conodonts were obtained from the Arenig-Llanvirn interval at Gröslii, but three uppermost Tremadoc samples collected from a thin layer of limestone nodules within the Solheim Slate Member was dominated by *Paltodus* cf. *deltifer* and *Paroistodus numarcuatus*.

A substantial database exists on the distribution of early Ordovician conodonts from deposits facing the Iapetus Ocean (e.g. Löfgren 1978, Stouge 1984, Rasmussen 1991, 1994) that has already featured in models for the evolution of the Iapetus Ocean and adjacent regions. Since the early 1960s it was clear that the occurrence and relative abundance of certain conodont taxa were related to their surrounding palaeoenvironments; but it was not until the early 1970s that the first proper palaeoecological models were proposed (see review by Pohler & Barnes 1990, table 1). Barnes & Fåhræus (1975) suggested that most conodonts were nektobenthic and a minority were pelagic. It is probable that the distribution of most Ordovician conodont taxa was related to depth and distance from shorelines, although some taxa apparently were pelagic (e.g. *Drepanodus* and *Scolopodus peselephantis*). Rasmussen & Stouge (1995) recognised five conodont biofacies within the Stein Formation: the *Scalpellodus*, *Microzarkodina*, *Baltoniodus*, *Drepanoistodus*, and *Protopanderodus-Periodon* Biofacies. These individual biofacies were believed to characterize distinct palaeoenvironments, suggesting that variations in the spatial distribution of biofacies reflect palaeoenvironmental changes through time. The *Scalpellodus* and *Microzarkodina* biofacies typified relatively shallow-water palaeoenvironments, whereas the *Baltoniodus*, *Drepanoistodus* and *Protopanderodus - Periodon* biofacies was interpreted as representing gradually deeper environments.

The brachiopod bearing samples from Andersön belong to the *Drepanoistodus* biofacies and *Protopanderodus - Periodon* biofacies, whereas the Haugnes sample may be assigned to the *Protopanderodus - Periodon* biofacies. As apparent in the numbers above, both *Protopanderodus* and *Periodon* are relatively more common at Haugnes than at Andersön, suggesting that Haugnes possibly was positioned relatively more distal on the shelf margin than Andersön.

In conclusion, both the conodont biofacies distribution and the sedimentological characteristics of the Andersön, Haugnes and Gröslii sections indicate an increasing depositional depth from Andersön, through Haugnes to Gröslii.

Comparison with platform faunas

The detailed study of the platform microbrachiopod faunas (Holmer 1986, 1989) forms a standard reference for studies of the rarer, less well-preserved faunas of the adjacent mountain belt. Holmer (1989) investigated two sections on the Swedish Baltoscandian platform in detail. Sections from the Holen Limestone and its equivalents (upper Arenig-lower Llanvirn) to the Skagen Limestone and its equivalents (upper Caradoc) at Kårgärde, Dalarna and Gullhögen, Västergötland were intensively sampled (Holmer 1989, figs 11 and 12). The red-bedded limestones of the Holen Formation are broadly coeval with the Stein Limestone and its equivalents; this facies of mainly lime mud and skeletal sand, is widespread across the Central Baltoscandian Confacies belt. In the Holen Formation of both Dalarna and Västergötland, *Numericoma*? *spinosa* is most dominant accounting for over 90% of the specimens retrieved; *Hisingerella*? and *Scaphelasma* were recorded from both areas however *Conotreta*? was only found at Gullhögen whereas *Myotreta* occurred only at Kåregärde. Faunas of the same age from the peri-Baltic area of northern Poland are similarly dominated by *Numericoma*? (Bednarczyk 1992).

Phosphatic microbrachiopod associations

Acrotretide brachiopods have recently been assigned to the class Lingulata on the basis of their phosphatic shells and similarity in morphology to other members of group (Popov *et al.* 1993). Based on a simple body plan with an active pedicle and a variety of dorsal internal structures the group generated considerable morphological diversity. The order first appeared during the early Cambrian and finally disappeared in the late Devonian (Harper *et al.* 1993); however the group reached its acme during the Ordovician. Biernat and Bednarcyzk (1991) have charted the diversity of the acrotretides through the Ordovician sequences of Poland; the highest diversities of both genera and species occurred through the Tremadoc to Llandeilo epochs and was followed by a decline in the late Ordovician probably related to global cooling. Wright and McClean (1991) however have emphazised the high diversities of latest Ordovician microbrachiopod faunas in carbonate mudmound environments in eastern Ireland suggesting in some areas at least diversity was maintained to near the end of the period. Nevertheless the group evolved a diversity of form during the Ordovician occurring a wide range of environments.

A few studies have attempted describe phosphatic microbrachiopods in terms of assemblages or associations that may be environmental controlled. Krause and Rowell (1975) described the inarticulate brachiopod fauna from the lower Llanvirn carbonate mudmound at Meiklejohn Peak, Nevada. Detailed maps of species occurrence indicated their distribution was not random; each had a preferred habitat in and around the mudmound. Moreover Krause and Rowell (*op. cit.*) considered a number of acrotretides may have been attached directly to the lithified substrate.

Biernat and Bednarczyk (1991) discussed changes in the Polish acrotretide fauna in terms of bathymetric and substrate changes with the clear implication that associations were related to ambient environmental conditions. Tremadoc and Arenig assemblages are diverse associated with nearshore facies; thick-shelled forms such as *Conotreta* and *Spondylotreta* may have been suited to sandy substrates. However deeper-water conditions during the Ashgill supported low diversity assemblages; *Acrothyra*, *Hisingerella* and *Scaphelasma* occurred in outer shelf and slope environments.

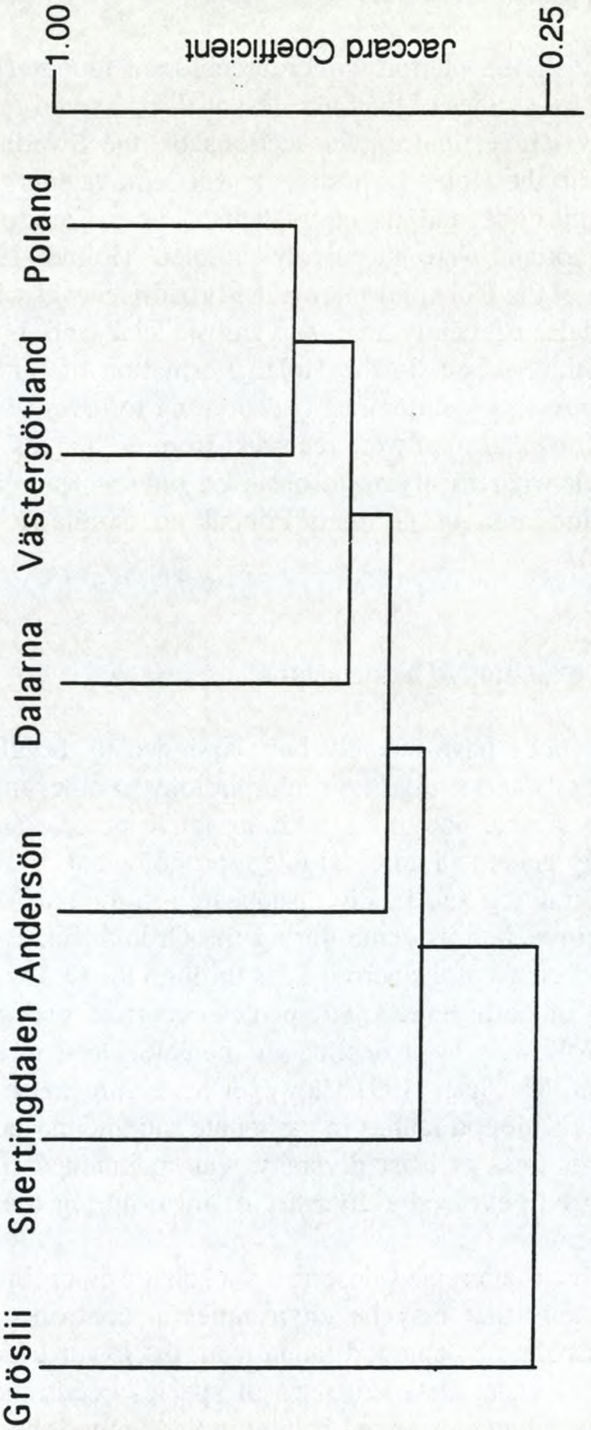

Fig. 2. Dendrogram generated from cluster analysis of 6 principal faunas from the Lower Allochthon and the Baltoscandian platform based on the presence and absence of 12 lingulate genera. The analysis interrogated a similarity matrix of Jaccard coefficients and the weighted pair group method was used to construct the dendrogram.

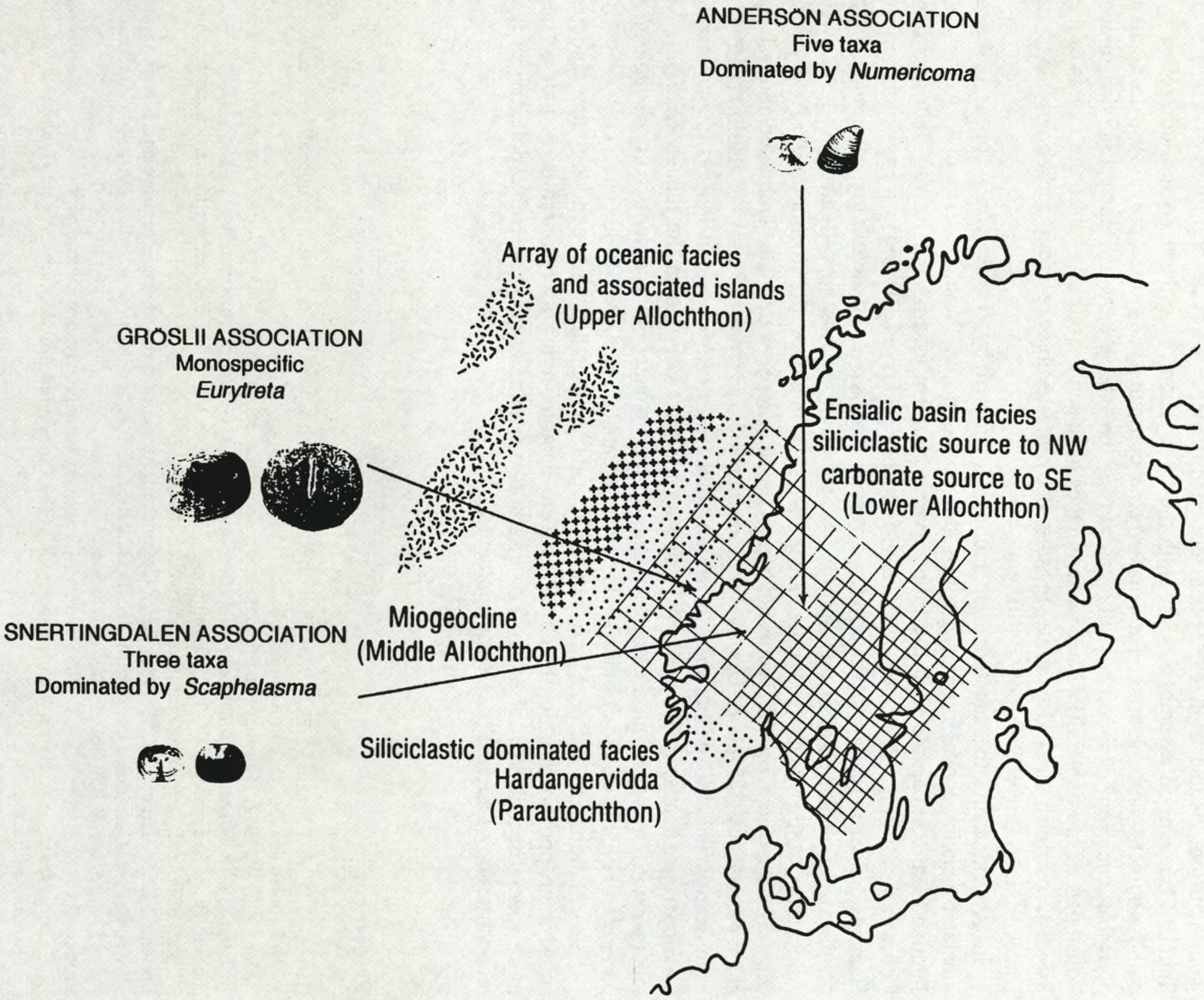

Fig. 3. Postulated palaeoenvironmental settings of the three phosphatic microbrachiopod faunas documented from the Lower Allochthon of the Scandinavian Caledonides.

In addition, Bednarczyk (1992) has refined this theme with the definition of a series of lingulate brachiopod assemblages, through the Ordovician of northern Poland, controlled by both bathymetry and substrate. The *Numericoma* community, as noted previously, dominated phosphatic microbrachiopod assemblages during the late Arenig and early Llanvirn in the Polish part of the peri-Baltic. These assemblages have some similarities with the faunas of the Swedish platform and also some broadly coeval assemblages from carbonate facies in Kazakhstan (Nazarov & Popov 1980).

In general terms the most diverse faunas, dominated by *Numericoma*, appear to have inhabited shallower-water environments; the fauna in the Stein Limestone at Andersön, Jämtland are comparable with the more diverse platform assemblages of the Baltoscandian Confacies belt (Fig. 2). Outer shelf environments preserved in the Lower Allochthon at Haugnes, Snertingdalen have lower diversity faunas overwhelmingly dominated by *Scaphelasma*. The most distal and deep-water environments investigated in the Lower Allochthon, at Gröslii in the Valdres area, have rare, monospecific assemblages of *Eurytreta* (Fig. 3). Elsewhere on the Baltoscandian platform deeper-water late Tremadoc faunas are characterized by lingulate faunas with species of *Eurytreta* (Popov & Holmer 1994). Significantly the slope rather than the mudmound facies of Meiklejohn Peak is characterized by *Eurytreta* (Krause & Rowell 1975). During the later Ordovician the related *Hisingerella* together with *Scaphelasma* are characteristic of deeper-water environments with *Hisingerella* particularly characteristic of black shale environments (Harper *et al.* 1985).

The relationships between acrotretides and facies together with conodont associations suggest the phosphatic microbrachiopods were benthic, living as epifauna or as part of the more specialised interstitial fauna. The group as a whole has a considerable potential in the reconstruction of ancient palaeoenvironments.

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THE LATE ORDOVICIAN (ASHGILLIAN) BRACHIOPOD ASSOCIATIONS IN THE EAST BALTIC

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Abstract

The distribution and composition of the Ashgillian *Holorhynchus* and *Hirnantia* faunas and the *Elsaella Streptis* associations in the East Baltic are presented. By the distribution of the chitinozoans these associations have a stratigraphic position between the first mentioned faunas. The youngest Ashgillian strata with graptolites of the *persculptus* Zone are known only from the southern Baltic.

Keywords: Brachiopods, associations, Ordovician, East Baltic.

Introduction

Brachiopods as predominating benthic organisms on clastic and carbonate shelves have an essential role in the study of mass extinction close to the Ordovician-Silurian boundary. The terminal world-wide Ordovician brachiopod fauna, the *Hirnantia* fauna has repeatedly been a subject of research during the last 20 years and more. Part of these studies are based on data from Baltoscandia (Bergström 1968, Brenchley & Newall 1975, Cocks 1982, Brenchley & Cocks 1982, Brenchley *et al.* 1994). Another widely distributed Ashgillian brachiopod fauna is characterized by the presence of a large pentamerid *Holorhynchus* (Rong Jia-yu and Harper 1988). Its co-occurrence with *Proconchidium, Tcherskidium* and *Eoconchidium* may indicate a distinctive biogeographic province in Eurasia (Rong Jia-yu *et al.* 1989). In Baltoscandia *Holorhynchus* has a patchy distribution and represents the topmost Ordovician brachiopod association in many sections of Norway, Latvia and Lithuania (Kulling 1933, Jaanusson 1982, Brenchley & Cocks 1982, Paskevicius 1968, Hints 1993).

The aim of this paper is to summarize the new and earlier obtained data on the composition and distribution of the *Hirnantia* and *Holorhynchus* faunas in the East Baltic and to establish two new onshore brachiopod associations. An attempt is made to specify the ecological position of the studied faunas in the Baltic Basin.

Some general aspects of the Ashgillian shelly fauna in the Baltic Basin were discussed by R. Männil (1966, Fig. 44). He identified in the East Baltic three roughly contemporaneous late Ashgillian biofacies - The *Holorhynchus, Ilmarinia* and *Dalmanitina* biofacies. These biofacies can be considered a basis for the analysis of shelly faunas, although their correlation, essence and index taxa of faunas were revised during later studies (Paskevicius 1968, Hints 1968, 1993, Nõlvak *et al.* 1989).

This study is mainly based on the data from core sections in different parts of the East Baltic. The limited amount of fossils and preliminary identifications of taxa make it impossible to distinguish unambiguously various brachiopod associations or to typify quantitatively their composition. Still, in the East Baltic four distinct late Ashgillian brachiopod faunas can be identified. They occur in facially different deposits and are partly or completely separated in space and time. For these faunas the following terms are used: the *Streptis*, *Elsaella* and *Holorhynchus* Association and the *Hirnantia* fauna. The latter is used here in its primary, wider-meaning (Temple 1965, Rong Jia-yu & Harper 1988) in comparison with the term "*Hirnantia* Association" by Brenchley and Cocks (1982). The data available do not enable to establish the ecological associations of the *Hirnantia* fauna in the East Baltic.

The above-mentioned three associations represent discrete onshore faunas and the term "association" by Brenchley and Cocks (1982) seems to be appropriate. The age of these faunas is estimated by the distribution of chitinozoans (Nõlvak 1986, Nõlvak & Grahn 1993). The correlation of stratigraphic units used in this paper is given in Fig. 1.

The *Holorhynchus* association

The large pentamerid brachiopod *Holorhynchus* was first established in Lithuania (Paskevicius 1963) in the Taucionys Formation, subsequently recorded in the same formation in eastermost Latvia (Ulst *et al.* 1982) and Pskov district (Nõlvak *et al.* 1989), and in the upper part of the Kabala Formation in Estonia (Hints 1993). The East Baltic specimens of *Holorhynchus* are referred to *H. gigenteus* Kiaer (Paskevicius 1963, 1968, Männil 1966, Männil *et al.* 1968). The complete shells are rare in the core sections and the species level identification of the fragments of thick shelled brachiopods, which presumably belong to *Holorhynchus*, is complicated. On some levels within the Taucionys Formation the valves of *Holorhynchus* form a coquina.

In the East Baltic the *Holorhynchus* Association is a low diversity association with rare *Eospirigerina* sp., *Thaerodonta* sp., *Ptychopleurella* sp., strophomenids and undescribed orthids (*Platystrophia dalinkevichusi* Paskevicius (nom. nud.), Laskovas *et al.* 1993). In the Taucionys Formation the algae *Vermiporella*, rugose and tabulate corals are the most common fossils, which accompany *Holorhynchus* (Paskevicius 1963, 1968, Nõlvak *et al.* 1989).

There is no firm evidence yet about the continuity of the *Holorhynchus*-bearing deposits in westernmost Estonia and other areas of the East Baltic (Fig. 2), but everywhere they are overlain by strata with the chitinozoan *Spinachitina taugourdeaui* (Nõlvak 1985, Nõlvak *et al.* 1989). This suggests that the *Holorhynchus* Association in the East Baltic is of late Pirguan age (Hints 1993).

In the late Ordovician regressive sequence the strata with the *Holorhynchus* Association constitute a separate part within the complex of algal (*Vermiporella, Paleoporella?*) limestones and reefs of the Pirgu and Porkuni stages. These rocks are well-developed in Estonia (Oraspõld 1975a, b). The algal and reef-like limestones underlie the *Holorhynchus*-bearing Taucionys Formation also in the south-eastern East Baltic (Ulst *et al.* 1982). The up to 10 m thick Taucionys Formation (Laskovas *et al.* 1984) stands out in this complex as a cryptocrystalline limestone with marly layers. The alternation of limestones with *Holorhynchus* and corals with bioturbated marls expresses the cyclicity of the sedimentation (Laskovas *et al.* 1993), which may show the oscillation of sea level or quite rapid changes in the income of the terrigeneous material. Notable is the beige or brown colour of the limestones and the dark brown colour of some of the marly layers, which presumably is caused by a high content of the organic matter.

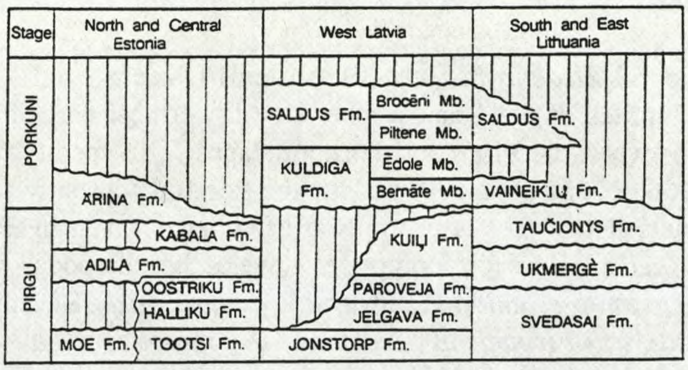

Fig 1. Correlation of the formations and members of two uppermost Ordovician stages in theEast Baltic (by Männil 1986, Laskovas & Paskevicius 1991, Roomusoks 1991, Nolvak & Grahn 1993).

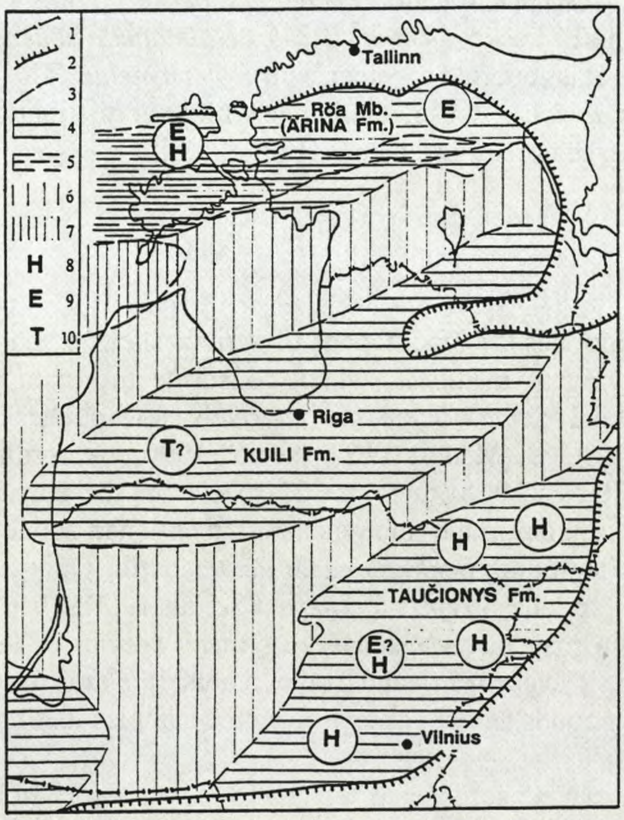

Fig. 2. The distribution of the Pirguan rocks and late Pirguan faunal associations. 1) administrative boundaries; 2) outer margin of the distribution of the Pirguan rocks (by L. Polma, unpublished data); 3) boundary of the distribution of formations; 4) area of the distribution of the late Pirguan rocks (the topmost unit is shown); 5) area of the distribution of the Kabala Formation; 6) area characterized by an essential gap in the upper part of the Pirgy Stage; 7) area, where the Pirgu Stage is missing; 8) *Holorhynchus* Association; 9) *Elsaella* Association; 10) *Tretaspis* Association?

The *Elsaella* Association

An enteleracean brachiopod *Elsaella bekkeri* (Rosenstein) (Rosenstein 1943, Hints 1975) is characteristic of the dolomites of the Röa Member (the lowermost part of the Ärina Formation) on the transition between the Pirgu and Porkuni stages. This member is included either to the Pirgu (Männil 1966, Rõõmusoks 1991) or the the Porkuni stages (Martna 1957, Männil 1987). The most complete list of fossils in the Röa Member is given by Rõõmusoks (1991, Tab. 1). Alongside *Elsaella bekkeri* it comprises several brachiopods, such as *Ilmarinia sinuata* (Pahlen) *Eoplectodonta schmidti* (Lindström), *Luhaia vardi* Rõõmusoks, *Geniculina pseudoalternata* (Schmidt), *Pirgumena martnai* Roomusoks, *Thaerodonta nubila* (Rõõmusoks), which are already common within the underlying Adila Formation. *Elsaella bekkeri* is presumably one of the few brachiopods, which appears in the Röa Member and occurs frequently in several core sections throughout northern Estonia. *T. nubila* is the most frequent species, that is associated with *E. bekkeri* in many cores. Among other fossils that are predominant in the Röa dolomites are the disarticulated columnales of crinoids. The stem segments are square in outline and probably belong to the genus *Xenocrinus*.

The genus *Elsaella* occurs also in Lithuania (*Elsaella* cf. *bekkeri*, Hints 1975, *Elsaella vokensis* Paskevicius (nom. nud.) Laskovas *et al.* 1984; = *Dalmanella vokensis* Paskevicius in Laskovas *et al.* 1993), where it is reported together with *Holorhynchus*. The co-occurence of species of *Elsaella* and *Holorhynchus* does not coincide with their distribution in Estonia and presumably also in the Ukmerge core in Lithuania (Hints 1993).

The *Streptis* association

The *Streptis* Association comprises the brachiopods of various skeletal and reef limestones of the Vohilaid, Siuge and Tõrevere members, which constitute the main part of the Ärina Formation and of the Porkuni Stage. There are relatively few published data about the brachiopods in the Porkuni Stage (Männil 1962, Nestor 1990). According to Rõõmusoks (unpublished data, doctoral thesis 1966) there are 22 species of the articulated brachiopods and about half of them are common in the three members mentioned above. They are, first of all, different strophomenids including *Schmidtomena acuteplicata* (Öpik), *Platystrophia* cf. *humilis* Oraspold, *Reushella* sp. and *Streptis undifera* (Schmidt). The list of brachiopods of the Porkuni Stage in Estonia may increase notably as many species of several genera (i.e. *Laticrura, Pthychopleurella, "Zygospira"* and some others) are not described yet. These species are mostly tiny brachiopods and are known from the samples that have been dissolved in acid.

The *Streptis* Association can be considered tentatively as an association characteristic of the reef complex, where the index genus is represented by *S. undifera* (Schmidt). It occurs in the reefs and in the grainstones of the flank facies. Based on data from core sections this species is quite frequent in northern Estonia in comparison with other brachiopods (excluding strophomenids). The specific shell ornament of *S. undifera* (Hints 1986) allows to identify the genus also in the cases when the shells are badly preserved.

Streptis undifera is also mentioned in the southern East Baltic in the Ukmerge Formation (Laskovas *et al.* 1993), i.e. below the *Holorhynchus*-bearing Taucionys Formation. This stratigraphical position is not consistent with the range of that species in Estonia.

The *Hirnantia* fauna

In the East Baltic the latest Ordovician brachiopod fauna is represented by the taxa of the typical *Hirnantia* fauna (Rong-Jia-yu & Harper 1988). Some key members of this fauna i.e. *Dalmanella testudinaria* (Dalman), *Eostropheodonta hirnantensis* (M'Coy) and *Hindella* sp. are referred to in several publications (Männil 1963, 1966, Männil *et al.* 1968, Ulst *et al.* 1982, Laskovas & Paskevicius 1993). The species are known from the Kuldiga Formation of the lowermost Porkuni Stage in the westernmost Latvia, Lithuania and in southern Estonia (Fig. 3). The revision of the Hirnantian brachiopods and their distribution in the East Baltic is in progress and here some preliminary data are given.

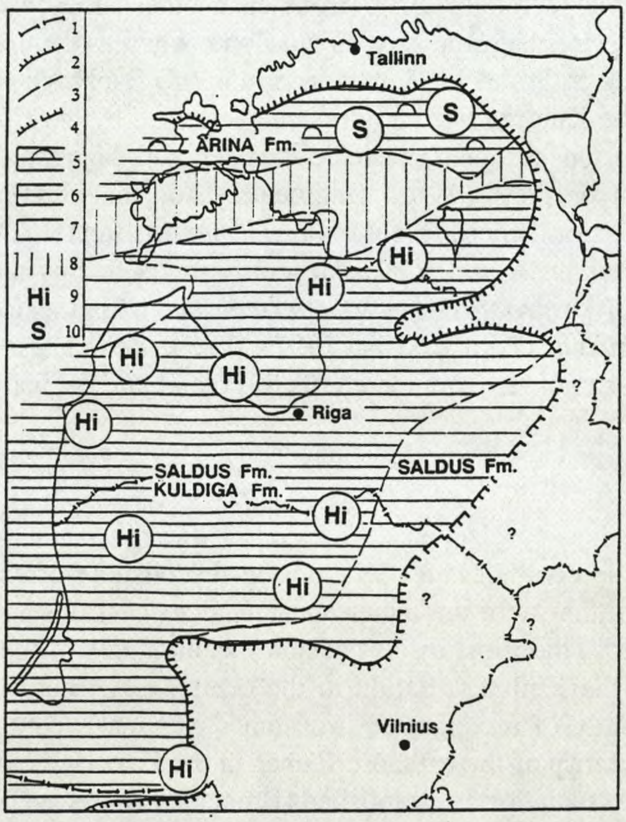

Fig. 3. The distribution of the Porkunian formations and brachiopod associations. 1-3) see Fig. 2; 4) probable boundary of the distribution of the Porkunian rocks (eastward there may occur rocks of the Porkuni age); 5) area of the distribution of the Röa Formation; 6) Area, where the Porkuni Stage is represented by the Kuldiga and Saldus formations; 7) the stage is represented by the Saldus Formation; 8) area, where the Porkuni Stage is missing or is represented by transitional rocks; 9) *Hirnantia* Association; 10) *Streptis* Association.

The lower part of the Kuldiga Formation of the Bernate Member, differs from its middle and upper parts i.e. the Edole Member by the scarcity of the brachiopods, and only *Cliftonia* sp., *Onniella* sp., *Kinnelle?* sp. and some strophomenids have been found. The whole fauna assemblage is diverse and cephalopods, microgastropods, tiny crinoid columnals of a square outline and algae occur.

The upper Edole Member of the Kuldiga Formation comprises, especially in westernmost Latvia, a diverse brachiopod fauna. In addition to the species mentioned above *Trematis norvegica* Cocks, *Hirnantia saggitifera* (M'Coy), species of the genera *Plectothyrella, Cliftonia, Aphnomena* and *Leptaenopoma* have been noted. The density of brachiopods increases upwards in the section, but several species (*Dalmanella testudinaria, Plectothyrella* sp.) are represented on some bedding planes by numerous shells, apparently in life position. Laterally, in the onshore direction, brachiopods become less diverse and rare. In the periphery of the Kuldiga Formation only *D. testudinaria* is found.

The possible existence of a separate early Porkunian association of the *Hirnantia* fauna in the lower Bernate Member of the Kuldiga Formation offers much interest for the correlation. The Bernate and Edole members differ from each other also by chitinozoans. The lower Bernate Member comprises the *S. taugourdeaui* Zone, whereas the upper Edole Member the *C ? scabra* Zone (Nõlvak & Grahn 1993, person. comm of J. Nolvak).

In the study area the Kuldiga Formation is characterized by the *Hirnantia* fauna is overlain by the Saldus Formation comprising shallow water and lagoonal deposits, which do not contain shelly fossils or they are too fragmentary for the identification. In the western, oceanward direction from these areas, the *Hirnantia* fauna (*Dalmanella testudinaria, Eostropheodonta hirnantensis* and *Hirnantia* sp.) is succeeded by the graptolite fauna of the *persculptus* Zone (*Glyptograptus? exgr. persculptus* (Salt.), *Climacograptus normalis* (Lapw.) and *Cl. angustus* (Pern.)) (Ulst 1992). These completest Ashgillian sections are located in the Baltic Sea not far from the coasts of Poland and the Kaliningrad region.

Conclusions

The establishing of the ecological associations in the onshore-offshore direction is complicated due to the discontinuity in the facies and faunal successions. According to the accepted stratigraphic framework, the most distant offshore equivalents of the *Holorhynchus* Association should occur in the Kuili Formation in the central East Baltic (Fig. 2). This formation consisting predominantly of redcoloured limestones and marls contains *Tretaspis latilimbus* which shows the similarity of the offshore faunas in the East Baltic and Norway. In the latter region the *Tretaspis* Association is established (Brenchley & Cocks 1982).

The onshore *Holorhynchus* Association was separated from the trilobite-dominated fauna in the central East Baltic by areas, where the late Pirguan deposits are missing. It is supposed (Ulst *et al.* 1982) that the gaps in the sequence are caused by the late Pirguan and/or early Porkunian denudation.

The distribution of the late Pirguan *Elsaella* Association falls into the period of possible sea-level rise to some extent, in comparison with the time, when the *Holorhynchus* Association dominated. This is marked also by the transgressive nature of the dolomites of the Röa Member, where the first association occur.

The relationships of the *Elsaella* and *Streptis* associations with the *Hirnantia* fauna are difficult to explain without using microfossils. The chitinozoans seem to be the best markers confirming that these two associations may be older than the typical and diverse *Hirnantia* fauna. The chitinozoan *taugourdeaui* Zone is established above the Taucionys Formation comprising the *Holorhynchus* association and in the lower part of the Kuldiga Formation (in the Bernate Member), where the typical elements of the *Hirnantia* fauna are rare. In Estonia the highest find of the *S. Taugourdeaui* comes from the kerogeneous limestones of the Siuge Member in the middle part of the reef complex of the Porkuni Stage (Nõlvak 1986).

The most essential environmental changes caused by the glacio-eustatic sea-level fall took place at the begining of the Porkuni time, when the *Streptis* Association became extinct and the first elements of the *Hirnantia* fauna appeared in the central part of the basin. It is clearly confirmed by the shift in th oxygen and carbon isotope values (Brenchley *et al.* 1994).

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DISTRIBUTION OF LINGULATE BRACHIOPODS IN THE EARLY ORDOVICIAN OF THE EAST BALTIC

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Introduction

In the East Baltic, the late Tremadoc and early Arenig interval is associated with a major change in the composition of the benthic faunas; during this time, the low diversity assemblages were replaced by medium to high diversity bentic assemblages including brachiopods, bryozoans, trilobites, ostracodes, and echinoderms. The brachiopods constitute the most important components of this new fauna. The most comprehensive study of the lingulate brachiopods from these beds was published by Gorjansky (1969), but the detailed distribution of lingulates in the early Ordovician of the East Baltic is still poorly known.

This brief review of the distribution of lingulate brachiopods in the Hunneberg, Billingen and Volkhov stages is based mainly on field work in Estonia and Ingria, during 1989-1993, the main results of which are still unpublished.

Stratigraphy

In Ingria (St. Petersburg district), the main lithostratigraphic units in the interval studied are the following: Nazya Formation (upper Tremadoc); Lakity (Hunneberg), Mäeküla, Vassilkovo and Päite members (Billingen) of the Leetse Formation; and the Volkhov Formation. The bases of the Nazya Formation and the Lakity Member represent erosional surfaces, and these units are preserved only in the eastern part of the area (Fig 1). In northern Estonia, the correlative sequence is represented by Varangu Formation (upper Tremodoc), and the Klooga, Joa (Hunneberg) and Mäeküla (Billingen) members of the Leetse Formation, the lower boundary of which is marked by a discontinuity surface. In Estonia, the Päite Member (uppermost Billingen) and overlying Volkhov beds are usually included in the Toila Formation (Fig. 1).

It was recently proposed (Williams et al. 1994) to place the lower boundary of the Arenig Series at the base of the *Tetragraptus approximatus* Biozone, somewhat below the first appearence of *Prioniodus elegans*; however, this level cannot be recognized clearly in the examined sections, but possibly lie somewhere within the Joa Member. In the most complete sections in Ingria, only the uppermost part of the *Paroistodus proteus* Biozone is present, and the Lakity Member and overlying beds may be referred to the Arenig.

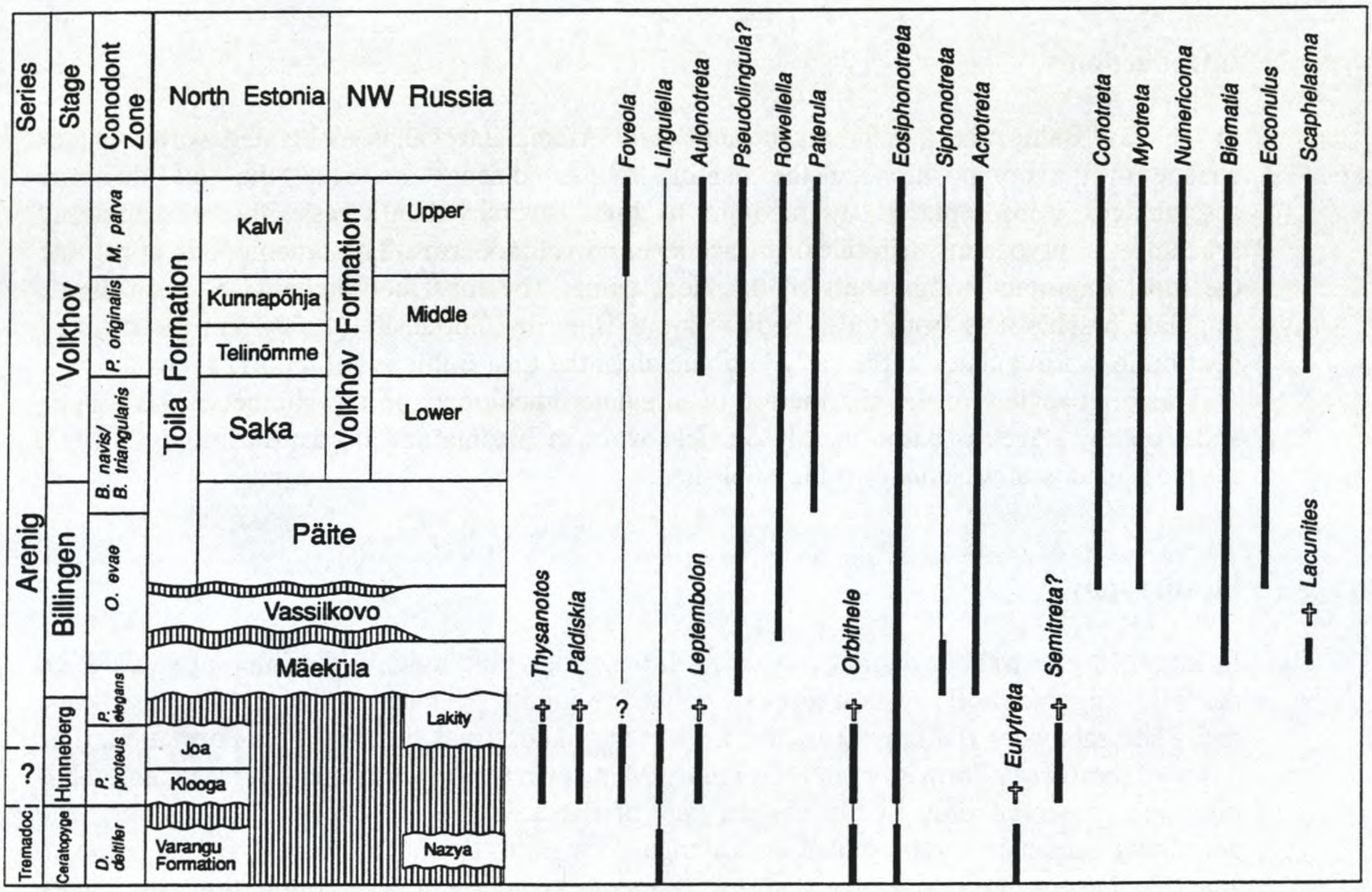

Fig. 1. Stratigraphical ranges of selected genera of lingulate brachiopods in the East Baltic

Distribution of Lingulate Brachiopods

The change (above the base of the *Drepanoistodus deltifer* Biozone) from deposition of mainly black mud to the formation of glauconitic sands and clays within the Varangu and Nazya formations indicate a sea level change as well as a decline in the organic productivity of the basin. These units contain only a few genera of lingulate brachiopods, including *Lingulella, Orbithele, Eurytreta*, and *Eosiphonotreta*. In the overlying transgressive sequence, comprising the Klooga and Joa members (within the *P. proteus* Biozone), the *Leptembolon/Thysanotos* assemblage first appears in northern Estonia. The assemblage also includes *Acrotreta*, usually associated with genera like *Orbithele, Eosiphonotreta*, and *Eurytreta*. In Estonia and the South Urals, the assemblage invariably occurs together with conodonts of the *P. proteus* Biozone. *Thysanotos* and *Leptembolon* are not known from correlative beds in Ingria, where only *Eosiphonotreta* has been found.

The Billingen Stage (upper part of *P. elegans* Biozone), the base of which is marked by an erosional surface, is associated with the onset of carbonate sedimentation. The *Leptembolon/Thysanotos* assemblage disappears and only *Acrotreta* remains abundant. The associated lingulate brachiopod genera in the lower Billingenian (the upper part of *P. elegans* and the lower part of *O. evae* biozones) includes *Lacunites, Siphonotreta, Eosiphonotreta*?, and *Biernatia*.

A diverse assemblage of micromorphic lingulates including *Paterula, Rowellella, Conotreta, Myotreta, Numericoma, Eoconulus*, and *Biernatia*, appears in the upper part of the *O. evae* Biozone. *Conotreta* and *Eoconulus* are also recorded for the first time from approximately contemporaneous beds in the South Urals and north Kazakhstan (Popov & Holmer 1994). In Scandinavia, *Rowellella*, and *Biernatia* are known already from the Tremadoc Bjørkåsholmen Limestone (Popov & Holmer 1994).

The base of the Saka Member is an erosional surface representing the sequence boundary between the Billingen and Volkhov stages. It is associated with the earliest occurrences of *Aulonotreta, Scaphelasma* and some other lingulates, but there is no significant change in the generic composition of the assemblages of micromorphic lingulates.

Discussion and Conclusion

During the early Arenig of the East Baltic, changes in the diversity of lingulate brachiopods seem to be due mainly to immigration of new taxa associated with the main sequence boundaries at the base of the Hunneberg, Billingen, and Volkhov stages. Above the base of the Volkhov stage most changes in the lingulate faunas seem to be related to faunal shifts and evolution within local lineages, but the these patterns have yet to be studied in detail. Some micromorphic acrotretid lineages appear to have evolved fairly rapidly during the Early Ordovician; there is a succession of assemblages with diagnostic taxa that may be of use for biostratigraphic control. However, the published record of the majority of these is extremely limited, and thus it is difficult to make wider comparisons outside of the study areas.

One important problem in understanding Lower Ordovician lingulate biostratigraphy is related to the age and origin of the *Leptembolon/Thysanotos* assemblage. The assemblage is now known from a wide geographic area surrounding the East European platform, as well as from the South Urals (Popov & Holmer 1994). The assemblages in all these districts are not associated usually with good zonal fossils; the presence of *Thysanotos*, in itself, has usually been regarded as evidence for a Tremadoc age. However, this is mainly a result of the previous interpretation of the glauconitic sands of northern Estonia as Tremadoc in age. Our observation suggest, that the range of this assemblage is restricted mainly to the Hunneberg Stage.

The affinities between the Early Ordovician brachiopod assemblages from Baltoscandia, the South Urals and north central Kazakhstan, may lend support to the palaeotectonic reconstruction recently proposed by Sengör et al. (1993).

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FACIAL ZONES, STRATIGRAPHY AND CORRELATION OF ARENIGIAN AND LLANVIRNIAN DEPOSITS IN THE SOUTH-EASTERN PART OF THE ORDOVICIAN BALTIC BASIN.

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The Arenigian and Llanvirnian deposits are penetrated by a large number of prospecting, structural, stratigraphic, mapping and other kinds of boreholes. Within the Baltic syneclise the Arenigian-Llanvirnian strata occur on depth from -2500-2600 to -1100-1200 metres, on the north-western slope of the Byelorussian-Mazurian anteclice to +70 m. On the highest part of anteclise the deposits after sedimentation are partly or entirely washed away and occur under the deposits of Devonian, Permian and Cretaceous age (Fig. 1).

The studies of litho- and biostratigraphy of Arenigian-Llanvirnian sequence were based on the cores of boreholes and when the boreholes were drilled without cores - on the geophysical logs (as a rule resistivity and gamma ray curves). Lithological studies were made by J. Laskovas, researching of brachiopods, graptolites, trilobites - by J. Paskevicius, conodonts and cephalopods - by V. Saladzius (conodonts in Butkunai borehole by S. Mägi), ostracodes by N. Sidaraviciene and S. Sauleniene.

On the south-eastern part of the Ordovician Baltic Basin a wide shelf monocline complicated with intrashelf Jelgava and Natangai depressions were established during Arenigian and Llanvirnian (Fig. 1). Within the shelf monocline Lithuanian - Byelorussian zone of grey carbonate facies (thickness 1,9-14 m) and Lithuanian - Latvian zone of variagated carbonate facies (thickness up to 40-45 m) was divided. Intrashelf depressions also are original facial zone but with variegated carbonate-argillaceous sedimentation (thickness up to 100-110 m). The original lithostratigraphical scheme was made for each of these zones.

The stratigraphy and facies distribution in the Lithuanian - Byelorussian zone have been described in articles by Laskov *et al.* (1981, 1983) and for the Jelgava and Natangai depression by Ulst *et al.* (1982) and Paskevicius & Laskovas (1983).

The stratotype sequence for the Arenigian-Llanvirnian of Lithuanian - Latvian zone is the Burkunai - 241 borehole (Fig. 2). The Leetse, Armonys, Medeikiai, Gindviliai, Mikonys, Obeliai, Juodupe, Vaidlenai (Segerstad) and Vyzunai formations have been established within Arenigian-Llanvirnian.

The Leetse Formation is cyclic in composition. 2-3 to 5-6 rarely 1 cycles are present. The lower parts (bases) of the cycles are composed of quartzitic - glauconitic sandstones (sand) and the upper parts vary from grey to rare variegated or red-brown limestones which often are dolomitized. In the uppermost part of the cycles and on the bedding surface the deep (3-5 cm) and wide (0,5-2 cm) pits are observed which are filled as a rule by glauconitic sand. Some of the surfaces, particularly in the upper part of formation are impregnated with Fe hydroxides. The thickness of cycles are up to 0,8 m. The total thickness of the formation is up to 1,35 m.

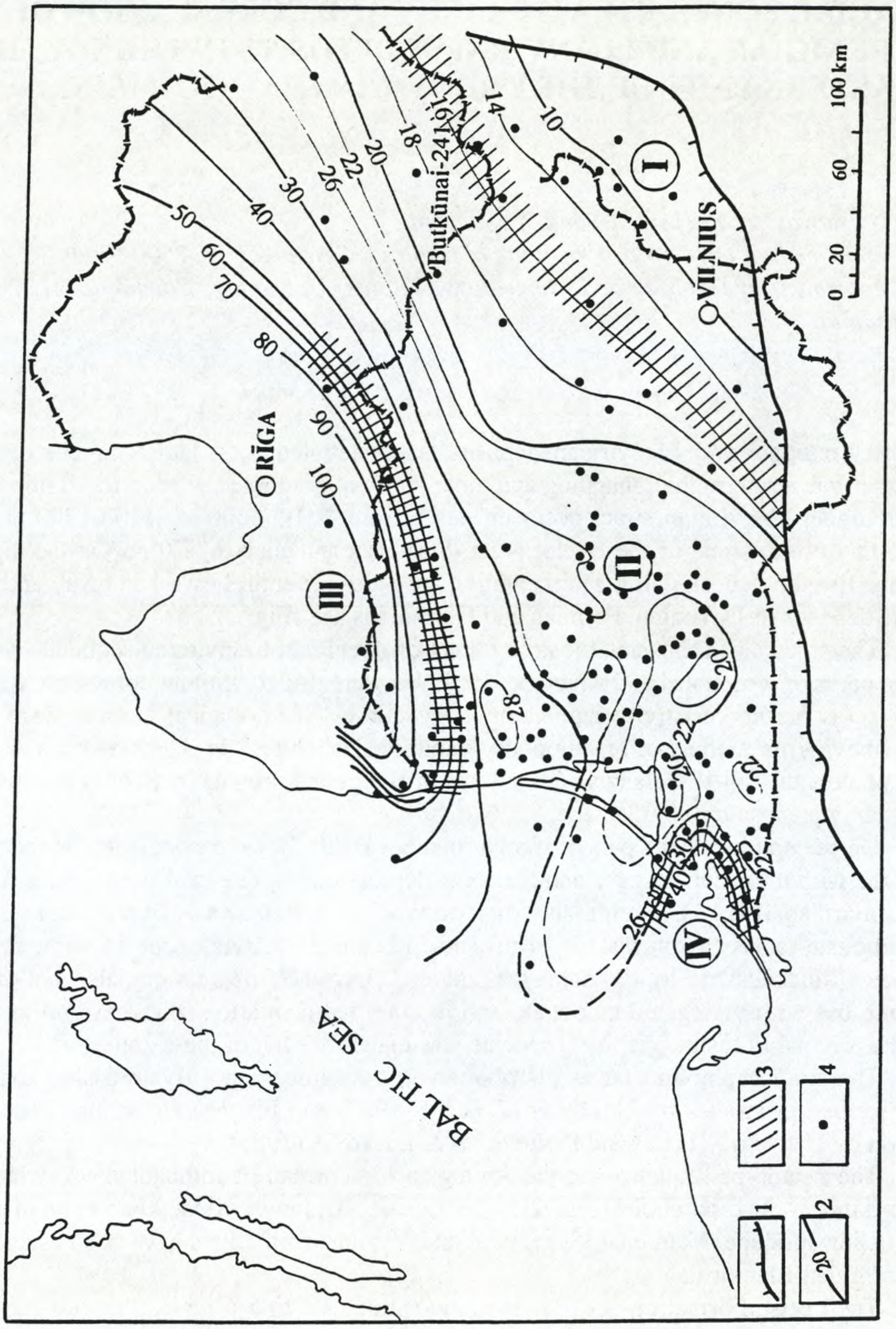

Fig. 1. The map of thickness and facial zonation of Arenigian-Llanvirnian deposits in the South-Eastern part of the Ordovician Baltic basin.

1 - border of modern distribution of deposits, 2 - isopachites of deposits (m), 3 - borders of facial zones, 4 - important boreholes penetrated deposits. Facial zones: I - Lithuanian-Byelorussian, II - Lithuanian-Latvian, III - Jelgava (depression), IV - Natangai (depression).

The fauna comprises conodonts and brachiopods. The conodont fauna includes the following conodont species listed in order of abundance: *Cordylodus angulatus* Pander, *C. rotundatus* Pander, *Oistodus lanceolatus* Pander, *Paroistodus proteus* (Lindström), *P. parallelus* (Pander), *Acodus deltatus* Lindström, *A. triangulatus* Fåhraeus, *Drepanodus conulatus* Lindström, *D. homocurvatus* Lindström, *D. suberectus* (Branson and Mehl), *D. forceps* (Lindström), *Oepikodus evae* Lindström, *Scandodus pipa* Lindström, *Scolopodus rex* Lindström, *S. pselephantis* Lindström. The brachiopod *Schizambon esthonae* Walcott is present.

The Armonys Formation consists of red-brown often dolomitized limestone. The detrital matter (30-70%) consists of trilobite remains. The typical feature of the formation is the large number (12-22) disconformity surfaces with yellow (goethitic) impregnation. Eastwards the quantity of disconformities diminishes. Westwards the goethitic impregnation changes into hematitic impregnation. The thickness of deposits is up to 1,3 m.

The fauna includes the following conodonts: *Acontiodus arcuatus* Lindström, *Oistodus basiovalis* Sergeeva, *O. forceps* Lindström, *Trichonodella flabellum* Lindström, *Protopanderodus rectus* (Lindström), *Cornuodus longibasis* (Lindström) and others.

The Medeikiai Formation is characterized by red-brown, rarely green argillites intercalated with greyish-green, rarely red-brown or variegated detrital micritic limestone, often dolomitized. In the eastern part of zone the thickness of the argillite is 0,5-3 cm, the thickness of the limestone or dolomite beds is 2-5 cm, whereas in the western part the thickness of argillitic beds increases to 10-12 cm. The texture of limestone is nodular. In the eastern part of the zone the surface of the carbonate is often impregnated with goethite (up to 16 surfaces). In the western part of the zone the goethitic surfaces are absent. The thickness of the formation is up to 4-5 m.

The conodont fauna comprises: *Drepanodus originalis* Sergeeva, *Cordylodus perlongatus* Lindström, *Falodus parvidentatus* Sergeeva, and *Microzarkodina flabellum* (Lindström), *Baltoniodus triangularis* (Lindström), *B. navis* (Lindström), *Paroistodus originalis* (Sergeeva), *Triangulodus brevibasis* (Sergeeva), *Drepanoistodus basiovalis* (Sergeeva) and others occur in the upper part of the formation.

The Leeste, Armonys and the lower part of the Medeikiai formations belong to the Latorp Stage of the Oeland Series.

The Spalviskiai Formation is composed of three packages. The lower and upper packages are carbonaceous and the middle package is argillaceous limestone. In the lower part of the formation one layer of glauconitic limestone is found. Up to 52 disconformity surfaces with goethitic impregnation are present. The thickness of the deposits is up to 5 m.

The fauna includes conodonts, brachiopods, ostracods, cephalopods and trilobites. The conodont fauna consists of *Baltoniodus alatus* (Hadding), *B. norrlandicus* Löfgren, *Acontiodus arcuatus* Lindström, *Scolopodus cornuformis* Sergeeva, *Oistodus complanatus* Lindström, *Roundya minor* (Lindström), *Microzarkodina parva* (Lindström), *Scandodus* sp. Lindström, *Tetraprioniodus minax* Sergeeva. Other fauna elements are brachiopods *Panderina tetragona lata* (Pander), *Ranorthis carinata* Rubel, *Plectella uncinata* Lamansky, the cephalopod *Dideroceras popovkense* Bal., the ostracodes *Ogmoopsis bocki* Öpik, *Tallinnella primaria* (Öpik) and the trilobite *Megistaspis* cf. *limbata* (Boeck) and others. The formation belongs together with the upper part of the Medeikiai Formation to the Volkhov Stage.

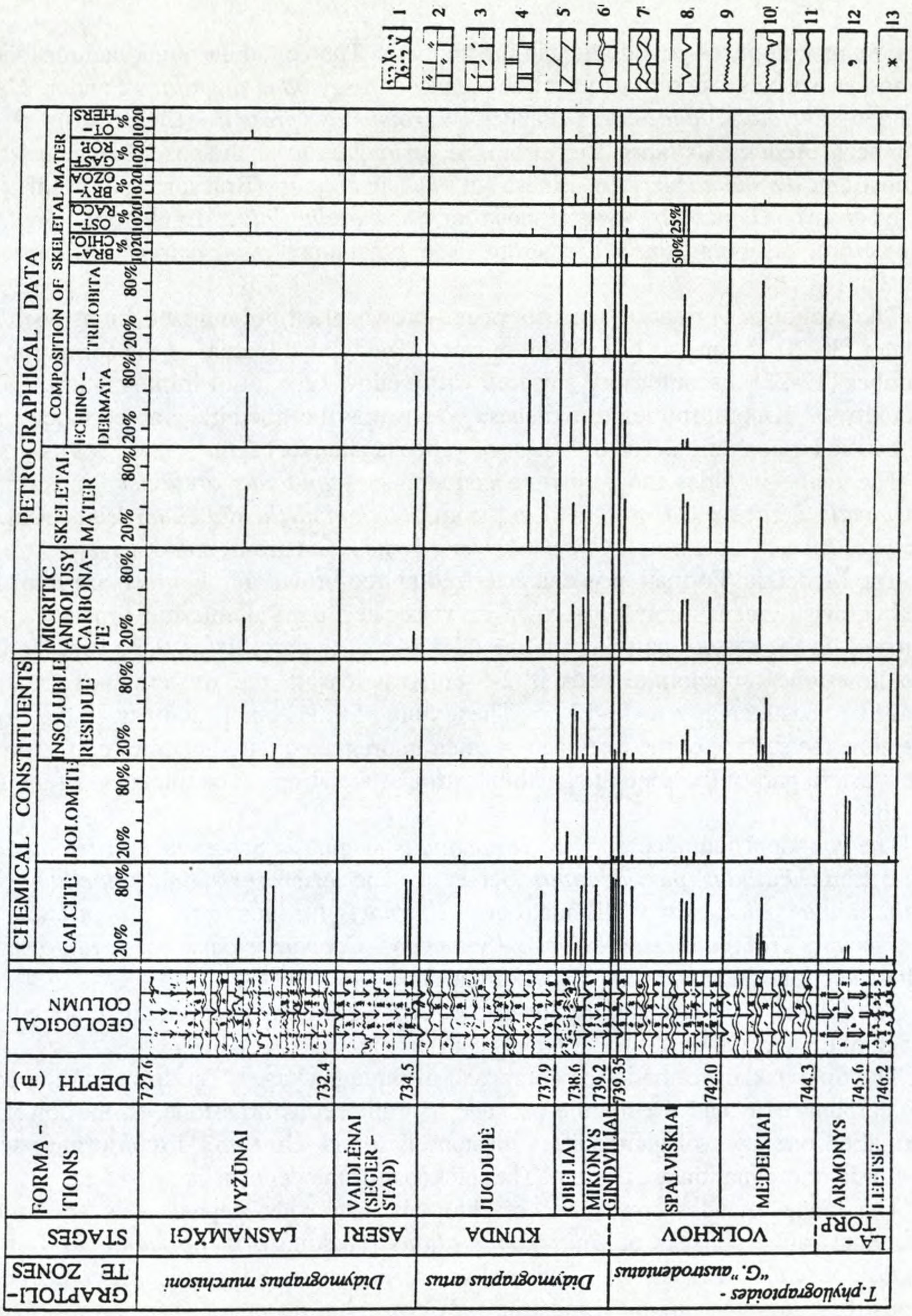

Fig. 2. Stratigraphy and lithology of Arenigian-Llanvirnian sequence of Butkūnai-241 borehole).

1 - quartz glauconitic sandstone, 2 - detrital limestone (skeletal matter more than 50%), 3 - detrital micritic limestone (skeletal matter 10-50%), 4 - dolomite, 5 - marl, 6 - nodular structure, 7 - wave band bedding, 8 - discontinuity surface with yellow (goethitic) impregnation, 9 - discontinuity surface with red (hematite) impregnation, 10 - discontinuity surface stylolitized, 11 - stratigraphical break, 12 - glauconite, 13 - reddish-brown rock.

The Gindviliai Formation is characterised by grey and greenish grey micritic, detrital micritic glauconite bearing limestones, often dolomitized and intercalated with marls. Westwards the formation changes into yellow, red-brown detrital micritic limestones similar to the underlying and overlying variegated rocks. The thickness of the formation is up to 1 m.

The fauna includes the conodonts: *Scandodus tortus* Viira, *Scolopodus varicostatus* Sweet & Bergström, *Acontiodus triangulatus* Fåhraeus, *Prioniodus* sp. 1 Lindström and the brachiopod *Antigonambonites schmidti* Rubel. The formation is tentatively referred to the Volkhov Stage.

The Mikonys Formation is composed of red-brown detrital, detrital micritic limestones, which sometimes are dolomitized. The detrital quantity is 40-55%, dominated by trilobite (40-70%) and echinoderm fragments (25-55%). Up to 32 goethitic impregnated surfaces are distinguished. The thickness of the formation is up to 2,6 m.

The fauna includes conodonts, but new conodont species are not distinguished, but some species are found which are transitional with those from lower formations; the brachiopod *Orthambonites orbicularis* (Pander) and the trilobite *Asaphus raniceps* Dalman are present.

The Obeliai Formation consists of green and dark green detrital micritic and detrital limestones and marls. The latter are dark grey and dark. The texture are horizontal bedding or wavy banded bedding. The thickness of the limestone beds is 2-5 cm, and the marls are 0,5-1,5 cm thick, in the western part the pyritized surface is present.

The thickness of the formation is up to 1,2 m.

The fauna includes the conodonts: *Amorphognatus variabilis* Sergeeva, *Ambalodus planus* Sergeeva, *Falodus parvidentatus* Sergeeva, graptolite: *Didymograptus* cf. *acutus* Ekström, trilobites: *Megistaspis heros* Dalman, cephalopods: *Suecoceras recurvum* (Holm), ostracods: *Pinnatulites procera* (Kumn.), *Uhakiella cicatriata* Sarv and Sphaeronites and others.

The Juodupe Formation is represented by red brown yellow grey detrital, rarely detrital micritic limestones with thin lenticular laminations of marls. Up to 128 discontinuity surfaces with goethitic impregation are found.

The detrital matter accounts for 55-60%. The predominante detrital component is echinoderm fragments (70-90%), in smaller quantities the trilobite (10-35%), brachiopods, ostracode and gastropod (1-4% each).

The fauna includes the conodonts *Microzarkodina ozarkodella* Lindström, *M. hagetiana* Stouge, *Paracordylodus* sp. 1 Lindström, *Baltoniodus* sp. 1 (Lindström) *Spathognathodus* sp., *Prioniodus alatus* Hadding, *P.* sp. 1 Lindström, the graptolite *Glyptograptus dentatus dentatus* (Brongnin), the brachiopod *Orthambonites majuscula* Rubel, the cephalopod *Paracyclendoceras concelatum* (Eichwald), the trilobite *Ielaenus dalmani* Wolb., the ostracodes *Aulacopsis simplex* Krause and *Pinnatulites procera* (Kumm.) and others.

The Mikonys, Obeliai and Juodupe formations belong to the Kundan Stage.

The Vaidlenai (Segerstad) Formation is composed of red brown detrital and detrital micritic limestones. Detrital matter account for 80%, dominated by echinoderms (70-80%) and to a lesser extent by trilobites (20-25%). Due to calcite overgrowth on crinoid fragments, the rock has the form of crystalline limestone. Up to 105 discontinuity surfaces with goethitic impregation are found. The surface, and sometimes the conchs of cephalopods, are covered by stromatolites and pigmented by hematite. In the western part of the zone a hematite impregnation of the discontinuity surfaces are found. The thickness of the formation is up to 4,1 m.

The fauna includes the conodont species *Walliserodus ethingthoni* (Fåhraeus), *Panderodus gracilis* (Branson & Mehl), the trilobites *Asaphus platyurus* Angelin, *A. punctatus* Lesn., *Paraceraurus ecul* Beyr., the cephalopods *Orthoceras centrale* His., *Nanno belemnitiforme* (Holm) and *Cochlioceras avus* Eichwald and the ostracode *Piritella tridactyla* Jaanusson. The formation belongs to Aseri Stage of the Middle Ordovician.

The Vyzunai Formation consists of grey, grey-green, yellow, red-brown detrital and detrital micritic limestones and marls. Detrital matter is represented by echinoderms (50-65%) and trilobites (20-45%). Three units (may be members or formations) are recognized in the sequences: the lower-variegated coloured unit, the middle-grey unit and the upper-variegated coloured unit. The texture of the rocks is often nodular. Up to 77 discontinuity surfaces with yellow goethitic and phosphatic impregation have been recognized. Stylolites also occur. The thickness of formation is up to 6,8 m.

The fauna includes the conodonts *Scandodus unistriatus* Sweet & Bergström, *Roundya inclinata* (Rhodes) *Oistodus abundans* Branson & Mehl, *Oistodus exelsus* Stauffer, *Tetraprioniodus minax* Sergeeva, *Scolopodus cornuformis* Sergeeva, *Paracordylodus lindströmi* Bergström, *Prioniodus variabilis* Bergström, *Falodus robustus* (Bergström), *Falodus prodentatus* (Gravers & Ellison), *Eoplacognathus foliaceus* (Fåhraeus), *Paltodus iniquus* Viira, *Panderodus compressus* (Branson & Mehl) and *Tetraprionidus asymmetricus* Bergström, the graptolite *Glyptograptus dentatus dainavensis* Paskevicius, the brachiopods *Christiania oblonga* (Pander), *Ladogiella transversa* (Pander), the ostracodes *Tallinnela sebyensis* Jaanusson, *Steusloffia linnarsoni* (Krause) and *Euprimites effusus* Jaanusson, the trilobites *Pseudoasaphus tecticaudatus* (Steinhardt), *P. acculatus* (Angelin), and the cephalopod *Orthoceras regulare* Schl. and others. The formation belongs to the Lasnamägi Stage.

Comparison of the Arenig-Llanvirn sequence of the Lithuanian-Latvian facial zone with contemporaneous deposits of Öland, which are described by Tjernvik (1956), Jaanusson (1960, 1972) and Van Wamel (1974) shows many common features which allow a correlation of the formations. It is possible to correlate the Leetse Formation with the Köpingsklint Formation, the Armonys and Medeikiai formations with the lower part of the Bruddesta Formation, and Spalviskiai Formation with the upper part of the Bruddesta Formation and Horns Udde Formation. In the Lithuanian - Latvian facial zone the Gindviliai, Mikonys, Obeliai and Juodupe formations are distinguished in the upper part of the lower Ordovician sequence. Information about this part of the Lower Ordovician on Öland has not been available. In our opinion, the Obeliai Fm, which consists of glauconitic limestones with *Sphaeronites*, is the best lithological marker. We found it in Poland (boreholes Stadniki, Bartosice and others) and in Southern Estonia (boreholes Elva, Karula). In middle Sweden the glauconite limestone (some layers) were established in Lyckhen II, III, Borensberg IV, Ekön boreholes (Wikman *et al.* 1980). The analogue glauconitic limestones are present on Öland (WOGOGOB field trip 1994[1]). The Middle Ordovician Veidlenei Formation may be compared with the Segersted Formation and the Vyzunai Formation with Skärlöv and Seby limestones.

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[1] The beds are included in the Gillberga Formation see. Stouge WOGOGOB fieldtrip 1994

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LATEST ORDOVICIAN OSTRACODS OF BALTOSCANDIA

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The late Ordovician ostracode fauna of Baltoscandian type is known from the East Baltic, northwestern Russia, Byelorussia, Podolia, northeastern Poland, Sweden and with some specific features, in Norway (Abushik& Sarv 1983, Gailite in Ulst *et al.* 1982, Henningsmoen 1954, Meidla & Sarv 1990, Neckaja 1973, Sidaraviciene 1992, Sztejn 1985 and references therein) and from galcial erratics of northern Europe (e.g. Schallreuter 1987). In the North and Central East Baltic latterally succeeding associations have been recorded in the ostracode distribution (Ulst *et al.* 1992, Meidla 1986). Their layout nearly follows the distribution of main confacies belts of the paleobasin (Jaanusson 1976) and apparently displays its ecological zonation. The aim of the present paper is to analyze the distribution and composition of ostracode associations in the uppermost Ordovician in various regions of Baltoscandia, integrating the data from the above publications and new data from the Upper Ordovician of Estonia.

General features of the late Ordovician ostracode fauna

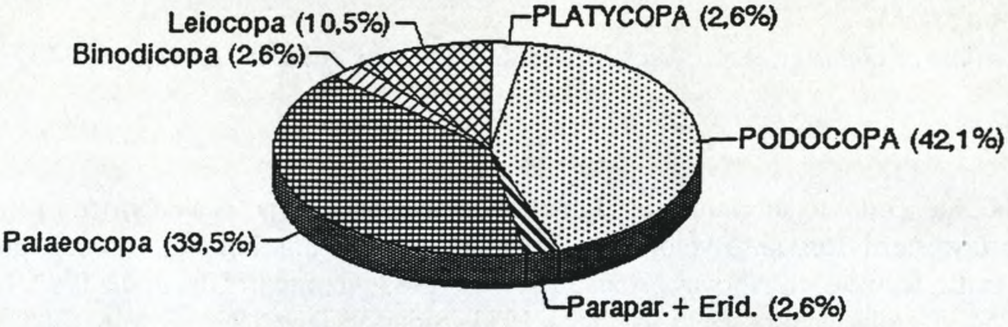
During the Oandu-Porkuni time, remarkable changes in the diversity of the East Baltic ostracode fauna can be followed. The diversity increase during the Oandu time reaching its maximum at early Pirgu time is followed by the decline and turnover at the Ordovician - Silurian boundary (Hints *et al.* 1989). The prevailing tendency in latest Ordovician seems to be the disappearance of taxa, resulting in the ostracode associations impoverished to a diferent degree.

The ostracodes of the uppermost Ordovician of the East Baltic are represented by the orders Beyrichiocopa, Podocopa and, in lesser extent, by Platycopa. Among beyrichiocopes, two major suborders - Palaeocopa and Binodicopa - are the most diverse ones in the Ordovician. The palaeocope/binodicope ratio as a measure of relative abundance has been considered to be useful for studying biogeographical and environmental aspects (Vannier *et al.* 1989). In most Ordovician sequences of Baltoscandia, the palaeocope/binodicope ratio exceeds 1. The average ratio for the post-Tremadoc Ordovician, 3:1 at the generic and 5:1 at the species level, is considered to be typical of high stability carbonate environments. In contrast, the ratio below 1 is characteristic of Ibero-Armorica, positioned on high latitude Gondwana, where the Ordovician succession with the ostracode record is represented mainly by clastic sediments (Vannier *et al.* 1989).

A trend of change in palaeocope/binodicope ratio occurs in the confacies belts of the Baltic palaeobasin. In the Pirgu Stage the ratio is higher within the Estonian-Lithuanian Confacies Belt up to 12) decreasing gradually towards the Central Baltoscandian Confacies Belt (>2), where the carbonates contain more clastic material. A similar trend occurs in other stages. Only some clasitc mudstone formations in Baltoscandia are binodicope-dominated, e.g. Sularp

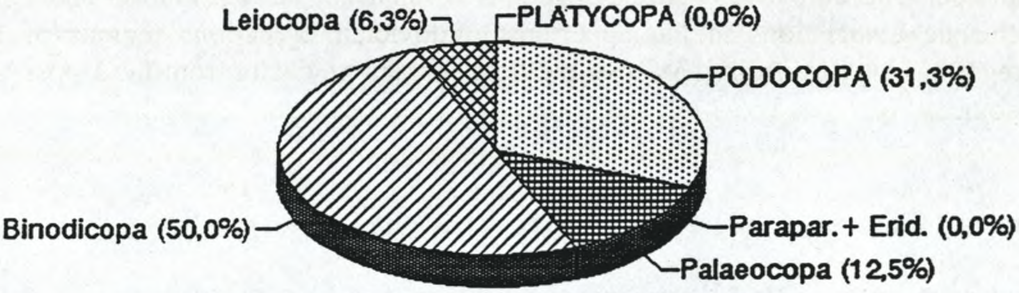
North Estonia, Ärina Formation.

Total number of genera 38
Palaeocope/binodicope ratio 15



South Estonia and West Latvia, Kuldiga Formation.

Total number of genera 16
Palaeocope/binodicope ratio 0.25



Östergötland, Borenshult locality.

Total number of genera 22
Palaeocope/binodicope ratio 0.67

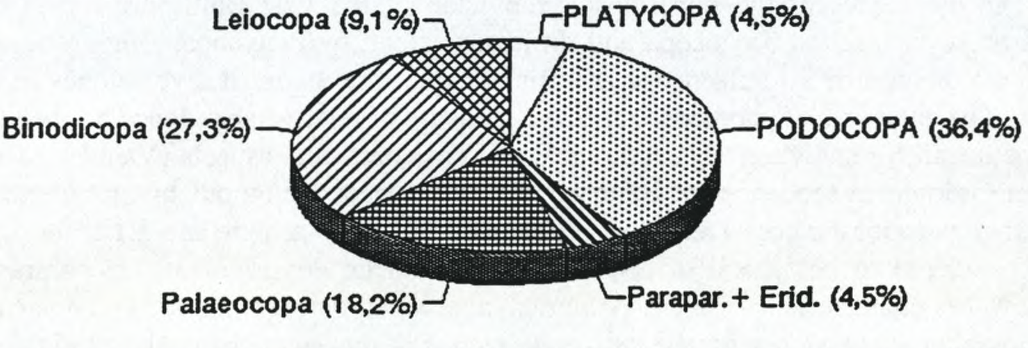


Fig. 1. Generic structure of the latest Ordovician associations in Baltoscandia

Shale of Scania (Schallreuter 1980, Vannier *et al.* 1989) and the "shaly" member of the Mossen Formation in West Latvia (Gailite in Ulst *et al.* 1982).

Late Ordovician ostracode associations in North Estonia

Two ostracode associations are known in the uppermost Ordovician of Baltoscandia. One of them occurs in northern Estonia, in calcarenitic and arenaceous limestones and dolomites of Ärina Formation containing small bioherms (Strubur 1956, Sarv 1962, Sarv, Meidla 1986) and another has originally been described by G. Troedsson (1918) in the uppermost Ordovician of Scania.

Beyrichiocope-dominated association of northern Estonia

The association occurring in the northern Estonia has been described by several authors (Stumbur 1956, Sarv 1962, Sarv, Meidla 1986). The association has high genric diversity and is dominated by beyrichiocopes, with 40% of podocopes and less than 10% of platycopes (Fig. 1). More than 50% of the beyrichiocopes are palaeocopes and the content of leiocopes is also remarkable. Less than 10% of all beyrichiocopes belong to other three suborder: Binodicopa, Paraparchiticopa and Eridostraca.The palaeocope/binodicope ratio is high 15:1.

The association shows the percentage of genera close to that of the Pirgu association in North Estonia, although the total number of genera is smaller: 38 against 55. Only three appearing genera are recorded in the Ärina Formation, the distribution of two of them being poorly documented. Both the taxonomical relationships and general association structure suggest that the association is mostly of local origin. No elements of the Hirnantian fauna have been discovered in the Ärina Formation, although *Holorhynchus* has lately been recognized from the underlying strata (Hints 1993).

Harpabollia association from Sweden and East Baltic

The *harpabollia* association has originally been described from Scania (Troedsson 1918). Revision of Troedsson's collection shows the close similarity of this association to the association described by Gailite (1970) from the Kuldiga Formation in western Latvia. A similar association has been discovered also in Lithuania and Poland (Sidaraviciene 1992, Sztejn 1985) and South Estonia. The Kuldiga Formation consists of limestones and marls, locally silty or sandy insome degree. In the East Baltic the formation has been tentatively correlated to the Ärina Formation in northern Estonia and assumed to correspond (together with the overlying Saldus Formation) to the Porkuni Stage within the Central Baltoscandian Confacies Belt (Männil, Meidla 1994). According to our preliminary studies, the samples from Borenshult locality in Sweden yield a similar assemblage.

Table 1.

Distribution of ostracode genera
in the Kuldiga-Borenshult (*Harpabollia*) association
in Baltoscandia and Poland.

	Estonia	Latvia	Lithuania	Sweden	
				Scania	Borenshult
BEYRICHIOCOPIDA					
Palaeocopa					
Platybolbina	*				
Tetradella					*
Sigmobolbina		*			
Laevanotella					*
Tvaerenella					*
Quadritia				*	
Neoprimitiella					*
Binodicopa					
Drepanella ?	*	*	*		*
Kinnekullea		*	*		
Ardenita			*		*
Scanipisthia	*	*	*	*	
Harpabollia	*	*		*	**
Pseudoancora	*	*	*		**
Aechmina	*	*	*	*	**
Pseudulrichia	*	*	*		
Circulina	*	*		*	**
Satiellina				*	
Eocytherella				*	
Leiocopa					
Baltonotella					*
Tschingizella		*			*
Paraparchiticopa					
none					
Eridostraca					
Cryptophyllus					*
PLATYCOPA					
Gotlandina ?					*
PODOCOPA					
Reversocypris ?					*
Pseudorayella					*
Microcheilinella	*	?			*
Daleiella					**
Longiscula				?	*
Pullvillites	*				**
Rectella	*			?	**
Medianella	*	?			*
Duplicristatia	*				

** = abundant   * = present   ? = questionably present

The composition of the ostracode association in the Kuldiga Formation and its equivalents (here called the *Harpabollia* association) is significantly different from the North Estionian association. It could also be considered as a beyrichiocope-dominated association, with one third of podocope genera and insignificant platycope component (Fig. 1), but among Beyrichiocopida, the suborder Binodicopa is prevailing which is unusual for Baltoscandia. In the generic percentage the palaeocope/binodicope ratio is 1:4 for the North and Cenral East Baltic, 1:5 for Borenshult locality, and 1:6 for Scania while in Lithuania the palaeocopes are missing and only binodicopes are recorded (Tab. 1).

The prevalence of binodicopes in Baltoscandia and elsewhere is mostly characteristic of clastic formations (see above). In contrast, *Harpabollia* association occurs in the limestones and marls, locally silty or sandy in some degree. In similar formations of the East Baltic a predominance of palaeocopes has been recorded throughout the Ordovician. By these conditions, the characteristic composition of the *Harpabollia* association is apparently not directly caused by substrate characteristics.

A distinct binodicope nucleus can be recognized in the *Harpabollia* associations in South Estonia, West Latvia, Lithuania, Scania and Östergötland (see Tab. 1). Several of these binodicope genera (*Drepanella, Scanipisthia, Harpabollia, Ardenita*, and possibly *Aechmina*) are recorded for the first time in the Ordovician of Baltoscandia. The composition of podocopes seems to be rather stable, while the recorded palaeocope composition varies in different localities.

Harpabollia association in the East Baltic is usually characterized by high number of specimens, but the taxonomic diversity and variability are low. The composition of the association in Scania cannot be estimated according to the published data. In the Borenshult locality the composition of the association is more variable (Tab. 1), but the genera *Aechmina, Circulina, Harpabollia* (binodicopes) and *Rectella* (podocope) constitute over 80% of specimens, whereas most of palaeocopes and some podocopes are represented by a single specimen in a sample of 800 specimens.

The appearance of the *Harpabollia* association is well-documented in several South Estonian and western Latvian sections in the lower part of the Kuldiga Formation, and occurs together with the normal type *Hirnantia* brachiopod fauna (Gailite 1970). The latest evidence shows that the appearance of this assemblage in the East Baltic succession coincides with the carbon and oxygen isotope shifts, indicating the beginning of late Ordovician glaciation event (Brenchley *et al.* 1994).

Brood (1978), who described the bryozoans of the Borenshult locality, noted that the material comes from a high energy environment and contains the trilobite *Mucronaspis mucronata* and the brachiopods *Hirnantia sagittifera, Dalmanella testudinaria* and *Leptaena rugosa* indicating Hirnantian age. These characteristics and the ostracode composition suggest the close relationships of the Borenhult material with that of the East Baltic Kuldiga Formation.

Discussion and Conclusions

The biostratigraphic correlation of the uppermost Ordovician in Baltoscandia is not unequivocally solved. One of the most discussed questions has been the relation of the Porkuni Stage and the Hirnantian, sometimes considered to be rough equivalents. The difficulties of correlation arise, first of all, from the biofacies differences traditionally explained by confacies zonation.

Taking into account the unique structure of the *Harpabollia* association, its link to the Hirnantian Fauna and isotope evidence in the same beds, it can be suggested, that the appearance of the *Harpabollia* association reflects a climatically induced immigration of ostracode faunas from higher latitudes and related changes (mostly decline) in the local assemblages. The rare representatives of the order Palaeocopa in the *Harpabollia* association are considered to be a recurrent element, mixed with immigrating binodicopes.

Considering the palaeogeographic position of Baltoscandia in Late Ordovician, the direct influx of cold water masses to a comparatively shallow shelf basin seems unlikely. It is more likely that the immigrating new fauna suppressed many palaeocope taxa and changed the taxonomical composition of binodicopes, e.g. by the disappearance of the genera *Klimphores, Ulrichia* and *Bullaeferum* in the East Baltic. Podocope ostracodes appeared to be the most stable element in the Baltoscandian ostracode fauna, as they almost maintained their role in the structure of the assemblage throughout the Late Ordovician, although their diversity decreased.

The Immigration concept is supported by the first bedrock finds of some elements of this fauna outside Baltoscandia. Schallreuter (1989) reported *Scanipisthia rectangularis* and *Harpabollia harparum* from the Cellon profile of the Carnic Alps, known as a locality of the *Hirnantia* fauna.

Vannier et al. (1989) have pointed out the increasing faunal links over the Rheic Ocean throughout the Ordovician, but usually they are expressed on the genric level. Therefore the finds in the Cellon profile could be considered as an evidence of more extensive faunal unification in very late Ordovician. Presumably, the *Harpabollia* association may prove to be a part of the Hirnantian fauna in a wider sense.

If we accept the assumption of immigration, it is difficult to explain the existence of a biohermal complex with high diversity fauna (Ärina Formation) comtemporaneously with a cold water *Harpabollia* association with no transitional assemblage. In Estonian example, if the Ärina Formation is equivalent ot the Kuldiga Formation, the presence of a geographical barrier in Central Estonia should be assumed. As there is no evidence of it, the main part of the Ärina Formation (and lowermost Porkuni Stage, accordingly) could be considered as pre-Hirnantian in age, but this suggestion could be best tested by the detailed isotope study. As the galciation apparently developed quite rapidly in the Late Ordovician, the faunal changes should be almost instant and the appearance level of the *Harpabollia* fauna could be considered a perspective correlation level within Baltoscandia. It can be traced effectively within the East Baltic and the presence of the assemblage is also verified in several regions of Sweden. The published materials about ostracods of Poland and Norway do not discuss this part of the sequence in detail, but in both cases, according to the preliminary data, amore or less distinct time separation of the palaeocope-dominated fauna and *Harpabollia* association can be assumed.

Acknowledgments

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ORDOVICIAN IN THE SOUTHERN LUBLIN UPLAND (SE POLAND)

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Abstract

The area of the southern Lublin Upland during the Ordovician was a part of a vast marine basin the northeastern part of which represented the southernmost periphery of the epicontinental Baltic Basin and the remaining part - a passage toward deeper sea.

The Ordovician succession starts with Lower Tremadocian. The sediments are clayey-silty-arenaceous and with graptolites. A shortlasting sedimentary gap is noted at the Tremadoc - Arenig Series boundary. From the lower Arenig clayey-carbonate sediments accumulated and started a new sedimentary cycle.

The most complete Ordovician sequence is preserved in the central part of the area. The succession begins with a glauconitite layer overlain by a clayey complex with graptolites in abundance.

Introduction

The southern region of the Lublin Upland was during the Ordovician times a part of the vast marine basin. The northeastern part of the Lublin Upland represented the southernmost portion of the epicontinental Baltic Basin (Mannil 1966). The remaining part was a passage toward the deeper sea.

The Ordovician sediments have been penetrated in several boreholes (Fig. 1) executed by the State Geological Institute and the Oil Industry. The profiles of these boreholes are diversified both in stratigraphy and lithology. This diversity forms the basis for distinguishing three basic areas: northeastern area (Krasnystaw - Hrubieszów), central one (Bilgoraj - Narol) and the southwestern one (Lubaczów) (Fig. 2).

Stratigraphy

Tremadoc

The Ordovician succession starts with Lower Tremadocian sediments which consist of a characteristic lithological complex. In the northeastern area these sediments rest directly on the Middle Cambrian rocks. In the central and southwestern areas the sediments overlie the Upper Cambrian deposits. The central and southwestern sediments are clayey - silty - arenaceous deposits. Only in the northeastern area coarse-crystalline limestone inliers and lenses appear in the uppermost part of the profile (Lendzion, Modlinski & Szymanski 1979).

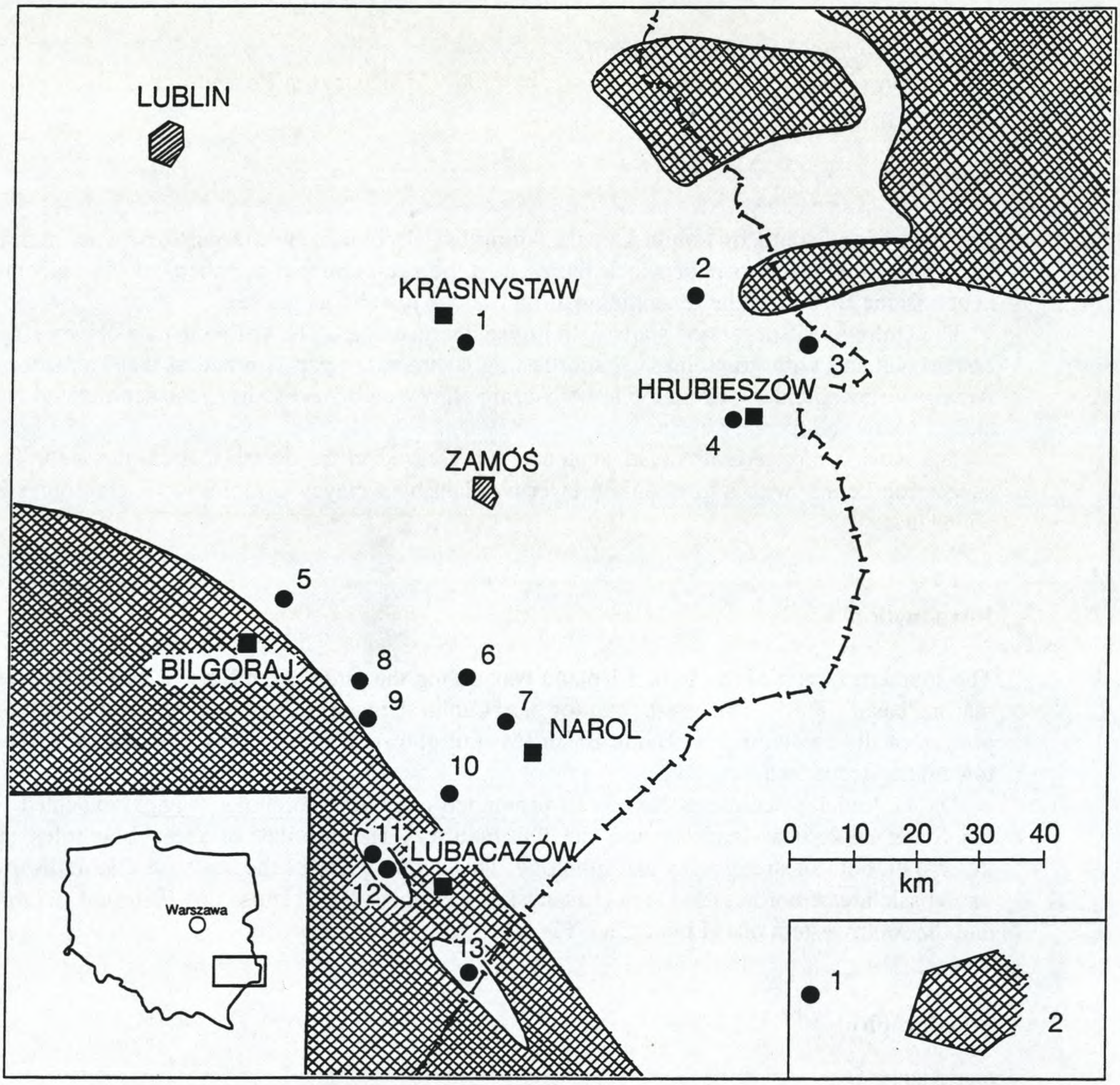

Fig. 1. Location sketch-map of the investigated area: 1. - boreholes that reached the Ordovician sequence; 2. - area devoid of Ordovician deposits due to epigenetic erosion; the boreholes: 1 - Lopiennik IG; 2 - Bialopole IG 1; 3 - Horodlo 1; 4 - Terebin IG 5; 6 - Narol PIG 2; 7 - Narol IG 1; 8 - Kozaki 1; 9 - Osuchy 1; 10 - Doliny 1; 11 - Uszkowce 4; 12 - Uszkowce 1; 13 - Lubaczów 14.

The age is documented by brachiopods and graptolites among which the following taxa have been found: *Rhabdinopora flabelliforme* s.l., *Rh. f. polonicum* (Tomczyk), *Anisograptus* sp., *A.* cf. *grandis* Bulman, *Clonograptus* sp., *Broggeria* sp., *Obolus* sp., O. cf. *apollinis* Eichwald. The above fossil assemblage is referred to Lower Tremadoc which corresponds to the Baltic Pakerort Stage. The presence of Upper Tremadocian sediments coeval with the Ceratopyge Stage has not been satisfactorily documented. It is possible however that the highest part of the sequence in the central area containing graptolites of the Bryograptus family (S. Pawlowski 1992) as well as the carbonate interbeds found in the northwestern area correspond to the Ceratopyge Stage.

A short gap in deposition marks the Tremadoc-Arenig series boundary in all three areas. A new depositional cyclus started in the Early Arenig and persisted throughout the whole of the Ordovician period. Clayey - carbonate sediments are dominating lithologies in the cyclus.

Arenig

The most complete Arenig succession has been recorded from the central area. The sequence begins with glauconitic beds which are overlain by a thick unit of clayey sediments with a very rich graptolite fauna (e.g. the Kozaki 1 borehole). The following graptolite zones have been established: *Tetragraptus phyllograptoides, Didymograptus balticus, Phyllographus densus, Phyllograptus angustifolius elongatus* and *Didymograptus hirundo*. The maximum thickness of the Arenig clayey sediments is 90m. At other locations in this area the Arenig sediments are strongly reduced to 0,5 - 0,8 m (e.g. the Narol IG 1 and Osuchy 1 boreholes). A thin inlier of carbonate rocks that overlies the conglomeratic - glauconitic layer yielded Upper Arenig conodonts. The conodont association is dominated by simple cone conodonts. The fauna includes *Drepanodus arcuatus Pander*, *Drepanoistodus forceps* (Lindström), *D. basiovalis* (Sergeeva), *Paroistodus paralellus* (Pander), *P. originalis* (Sergeeva), *Baltoniodus navis* (Lindström), *Oistodus lanceolatus* Pander, and *Protopanderodus rectus* (Lindström).

The presence of the multi-element species *Drepanoistodus forceps* (Lindström), *Baltoniodus navis* (Lindström) and *Oistodus lanceolatus* Pander are sufficient to refer the sediments to the Arenig. The common occurrence of *Drepanoistodus forceps* (Lindström) and *D. basiovalis* (Sergeeva) places these sediments within the *Baltoniodus triangularis* and *B. navis* zones (*sensu* Lindström 1971). The zones correspond to the lower Volkhov Stage of the Baltic region.

Also in the northeastern area the Arenig sequence starts with glauconitic and conglomeratic sediments above which succeed pure limestones and dolomitic limestones with marly interbeds and with numerous discontinuity surfaces. Only poorly preserved fragments of macrofossils have been found (Modlinski 1984) including: *Megistaspis* sp., *Symphysurus* sp., *S. dorsatus* Poulsen, *Nileus armadillo* Dalman. The sediments are 1,5 - 2,5 m thick.

In the southwestern area the Arenig sequence consists of thin carbonate sediments (below 1 m) with a conglomerate at the base. The age is documented by conodonts found in the Uszkowce 1 profile. The conodonts are: *Drepanodus arcuatus* (Pander), *Cornuodus longibasis* (Lindström), *Drepanoistodus forceps* (Lindström), *Scolopodus peselephantis* Lindstöm, *S. rex* Lindström, *Periodon flabellum* (Lindström), *Oistodus lanceolatus* Pander, *Baltoniodus navis* (Lindström), *Microzarkodina flabellum* (Lindström), *Drepanoistodus basiovalis* (Sergeeva), *Paroistodus originalis* (Sergeeva) and *Protopanderodus rectus* (Lindström).

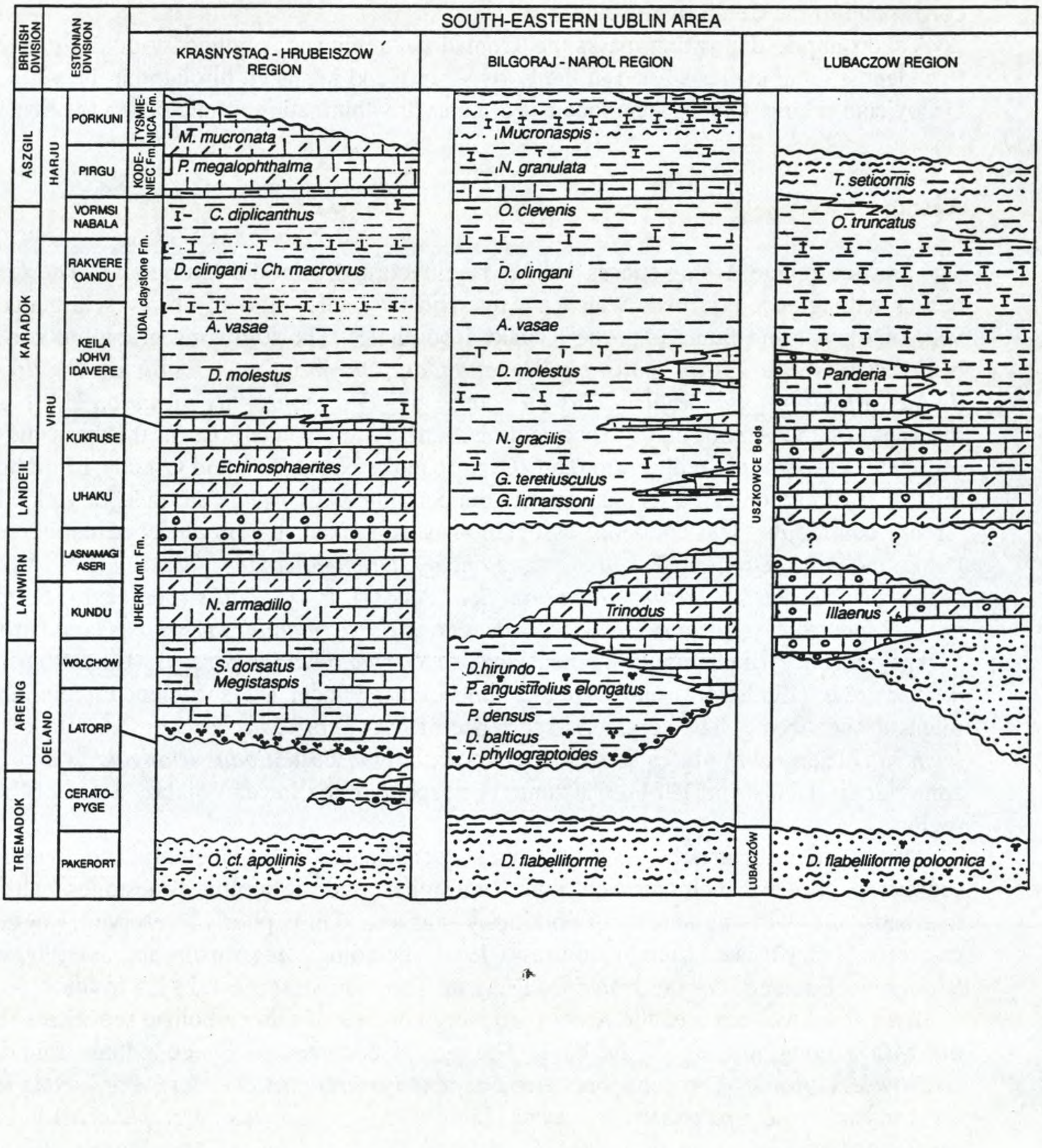

Fig. 2. Schematic correlation of the Ordovician deposits in the southern Lublin Upland

The coocurrence of the multielement species: *Microzarkodina flabellum* (Lindström), *Oistodus lanceolatus* Pander, *Drepanoistodus forceps* (Lindström) and *Baltoniodus navis* (Lindström) shows that the sediments are of late Arenig age. In the conodont zonal scheme of Lindström (1971) the fauna corresponds to the *Baltoniodus navis* Zone or to the lower part of the *Microzarkodina flabellum assemblage Zone* of Van Vamel (1974).

The presence of upper Arenig - lower Llanvirn sediment has been well-documented on the basis of conodonts in the Osuchy 1 borehole and the Lopiennik IG 1 borehole. The latter well is located in the northeastern part of the Lublin Upland (Fig. 1). In the Osuchy 1 borehole the age determination is based on the presence of multi-element species: *Amorphognathus variabilis* Sergeeva and *Protopanderodus rectus* (Lindström). The range of *Amorphognathus variabilis* Sergeeva corresponds to the Kunda Stage of the Baltic region. Its first appearance has been correlated with the uppermost Arenig (in the British zonation) by some authors (Lindström 1971, Löfgren 1978) or placed in the Llanvirn as a whole (Modlinski 1973). The Kundan sediments are referred to the *Amorphognathus variabilis* Zone cf. Lindström (1970).

The lower part of the *A. variabilis* Zone (= *Eoplacognathus? variabilis - Microzarkodina flabellum* Subzone *sensu* Löfgren 1978) is allocated to the uppermost Arenig whereas the part of the *Amorphognathus variabilis* Zone og Lindström (= *Eoplacognathus? variabilis - Microzarkodina ozarkodella* Subzone *sensu* Löfgren 1978) corresponds to the lower Llanvirn (Bergström & Orchard 1985).

The species *Amorphognathus variabilis* Sergeeva is confined to the *A. variabilis* Zone. As in the Osuchy 1 borehole this species occurs together with *Protopanderodus rectus* (Lindström). The stratigraphic range of *Protopanderodus rectus* extends into the base of the *Eoplacognathus? variabilis - Microzarkodina ozarkodella* Subzone suggesting that the sediments cannot be younger than lowermost Llanvirn and not older than latest Arenig. Coeval sediments were penetrated in the Lopiennik IG 1 borehole (NE Lublin Upland) and in the Uszkowce 1 borehole located in the southwestern part of the area (Fig. 1).

At the Lopiennik IG 1 section the dating has been based on concurrence of *Drepanoistodus basiovalis* (Sergeeva) and *Protopanderodus rectus* (Lindström).

At the Uszkowce 1 succession *Baltoniodus clavatus* Stouge & Bagnoli has been recorded. Its range corresponds largely to the lower *Amorphognathus variabilis* Zone. The presence of this species in the conodont assemblage shows that the sediments correspond to the uppermost Arenig - lowermost Llanvirn.

Llanvirn

The Llanvirn sediments are not thick in all the areas described here. The sediments are only a few metres thick and in some sections from the central area (e.g. Dyle IG 1 and Kozaki 1 boreholes) the sediments are absent. The rocks include organodetrital limestones - in places with ferruginous oolites - and marly limestones. The macrofossil fauna comprises: *Asaphus* sp., *Nileus armadillo* Dalman (at Lopiennik IG 1), "*Endoceras*" sp. (at Terebin IG 5), *Illaenus* sp. (Uszkowce 1) and *Trinodus* sp. (at Narol IG 1). These fossils are insufficient for precise age determination which was established on the basis of micro-paleontological studies.

The Llanvirn sediments have been documented on the basis of conodonts obtained from the Narol IG 1, Narol PIG 2 and Lopiennik IG 1 boreholes.

The multi-element species *Baltoniodus prevariabilis* Fåhraeus was found in the Narol IG 1. The range of this species corresponds to the uppermost *Eoplacognathus suecicus* Zone and extends into the lower part of the *Pygodus anserinus* Zone (Bergström 1971). Coexistence of this species with *Drepanoistodus basiovalis* (Sergeeva) which is unknown from above the lower part of the *Pygodus serra* Zone suggests that the sediments found in Narol IG 1 correspond to the lower part of the upper Llanvirn.

Eoplacognathus Hamar appeared to be useful in precise age determination among the conodonts encountered in the Narol PIG 2. The genus appeared in early Llanvirn and persisted into the early Caradoc. The specimen found in the Narol PIG 2 borehole does not allow an exact species identification. Its presence however shows that these sediments cannot be older than lower Llanvirn.

Apart from the taxa cited above the following conodont species have been identified in the Narol IG 1 and Narol PIG 2: *Acodus mutatus* (Branson and Mehl), *Cornuodus longibasis* (Lindström), *Drepanodus arcuatus* Pander, *Drepanoistodus venustus* (Stauffer), *Scalpellodus laevis* Dzik and *Protopanderodus rectus* (Lindström).

The upper Llanvirnian sediments have been dated by conodonts recorded from the Lopiennik IG 1 borehole. *Eoplacognathus lindstroemi* (Hamar) and *Amorphognathus kielcensis* Dzik show that these sediments correspond to the uppermost Llanvirn.

Eoplacognathus lindstroemi is the index taxon for the *Eoplacognathus lindstroemi* Subzone of the *Pygodus serra* Zone *sensu* Bergström (1971). The range of *Amorphognathus kielcensis* corresponds to the interval from the *Eoplacognathus foliaceus* to the *Amorphognathus kielcensis* Subzone. Commonly *Prioniodus (Baltoniodus) prevariabilis* Fåhraeus is associated with the two above-mentioned species. Also *Oneotodus mitratus* (Moskalenko) has been identified from this level. This taxon has been described for the first time from lower Caradocian deposits of the Siberian Platform (Moskalenko 1973). So far it has been obtained with the conodont assemblages of the North Atlantic paleozoogeographic province from the Mojcza limestone in the upper Llanvirn sediments (i.e *Eoplacognathus robustus* Subzone) and in the Llandeillo (Dzik 1976).

It is also noteworthy that ostracods occur in rather large quantities (preserved usually as internal moulds) and inarticulate brachiopods are present in the southern part of the Lublin Upland the uppermost Arenig - Llanvirn sediments. The fossils are particularly abundant in the Narol PIG 1 borehole. All the identified brachiopod specimens belong to the order *Acrotretida* Kühn.The following species are present: *Eurytreta intermedia* Biernat, *Scaphalasma subquadratum* Biernat and *Paratreta similaris* Biernat. These species come from the Llanvirn - Llandeil deposits and in the boreholes Goldap IG 1, Bartoszyce IG 1 and Ketrzyn IG 1 (northeastern Poland). *Scaphelasma subquadratum* Biernat is known also in Volkhovian deposits of Estonia (Tallinn vicinity).

Llandeilo

The most complete Llandeilo profiles are over 50 m thick and have been found in the central area. The sediments are dark grey claystones with limestone beds. The age of these sediments is documented by rich graptolite assemblages among which the following species were identified: *Glyptograptus teretiusculus* (Hisinger), *G. cernuus* Jaanusson, *G. eugluphus* (Lapworth), *Gymnograptus linnarssoni* Tullberg and others. Conodonts are present as well and *Amorphognathus inaequalis* Rhodes and *Baltoniodus variabilis* Bergström have been recorded. The conodonts were found at Narol IG 1 borehole. The species occur in the Llandeilian deposits only. The conodont zonation corresponds to *Amorphognathus inaequalis* Subzone of the *Pygodus anserinus* Zone and to lowermost part of the *Baltoniodus variabilis* Subzone of the *Amorphognathus tvaerensis* Zone. *Baltoniodus variabilis* Bergström is an index species of the *Baltoniodus variabilis* Subzone which correlates with the uppermost Llandeilo - lowermost Caradoc series, but it appears already in the *Amorphognathus inaequalis* Subzone.

In the northeastern area the Llandeilo sediments are recrystallized organodetrital limestones which are grey to dark-grey in colour and with thin interbeds of clay and marl. Macrofossils include undeterminable fragments of brachiopods, trilobites and numerous cystoids which are referred to the genus *Echinosphaerites*. In the conodont assemblage in Lopiennik IG 1 *Amorphognathus inaequalis* Rhodes was found and associated with *Baltoniodus prevariabilis* Fåhraeus. This suggests that these sediments can be correlated with the lower part of *Amorphognathus inaequalis* Subzone which is lower Llandeilo (Bergström & Orchard 1985).

The Llandeilo deposits are 6-7 m thick in the northeastern area. In the southwestern area (Lubaczów) grey limestone with ferruginous ooids passing upward into grey marls and siltstones are referred to the Llandeilo. The age of these rocks has not been paleontologically documented.

The Ordovician sediments corresponding to the uppermost Llandeilo - lower Caradoc series have been penetrated in the Narol IG 1 borehole. The concurrence of the following multi-element conodont species *Amorphognathus tvaerensis* Bergström and *Baltoniodus variabilis* Bergstöm has been mentioned. This is sufficient for the correlation with the *Amorphognathus tvaerensis* conodont Zone which embraces the uppermost Llandeilo and the lower Caradoc. According to the Estonian chronostratigraphic scheme the Kukruse, Idavere, Johvi and lower part of Keila stages are Llandeilo - early Caradoc. Three subzones have been identified on the basis of occurrence of the index species *Baltoniodus variabilis* Bergström, *Baltoniodus gerdae* Bergström and *Baltoniodus alobatus* Bergström. The two last-mentioned species were not found in the studied part of the Narol IG 1 borehole and hence these sediments presumably correspond to the *Baltoniodus variabilis* Subzone i.e. the highest Llandeilo - lowermost Caradoc.

Caradoc

The thickest Caradocian strata were found in the central area (e.g. Kozaki 1). Darkgrey to black claystones and silty claystones with interbeds of marly and organodetrital limestones dominate there. There are some rare and thin bentonite beds. Abundant graptolites allowed to establish the following zones: *Nemagraptus gracilis, Diplograptus molestus - Climacograptus wilsoni, Dicranograptus clingani* and *Climacograptus styloides*. The Caradocian sediments are about 120 up to over 200 m thick.

In the northeastern area the thickness diminishes to about 25 - 60 m. The lowermost part of the succession consists of darkgrey, recrystallised and organodetrital limestones with cystoids of the genus *Echinosphaerites* and of the brachiopods *Nicolella* sp. and *N.* cf. *alliku* Oraspyld. Higher up in the sequence darkgrey less frequently greenish claystones are present. The claystones are with graptolites referred to the *Diplograptus molestus* and the *Climacograptus styloides* Zones.

At Lopiennik IG 1 some rare and badly preserved conodonts were found. *Amorphognathus superbus* Rhodes has been identified which is the index taxon for the *Amorphognathus superbus* Zone (Bergström 1971). This zone corresponds to the upper Caradoc and lower Ashgill series. The conodont fauna, however, does not allow to give a precise reference to either the upper Caradoc or to the lower Ashgill for these sediments.

In the Lubaczów area the limestones with *Panderia* are passing upwards into silty claystones and calcareous claystones corresponding to the Caradoc (Tomczyk 1962). The sediments are about 15 m thickin the Lubaczów area.

Ashgill

The lower Ashgill Series corresponding to the Baltic Pirgu Stage of the central part of the southern Lublin Upland consists of grey somewhat limy claystones with minor occurrences of limestones and marls. The upper Ashgill includes dark-grey to greenish claystones with irregular lenses and stripes of silty - arenaceous and limy-arenaceous sediments.

Organodetrial limestones, marly limestones with interbeds of marls and calcareous claystones are characteristic for the lower Ashgill in the northeastern areas. The upper part of this series is developed as dark grey marls and limy claystones with lenses of recrystallised marly limestone.

To the southwest and in the Lubaczów area only lower Ashgill sediments have been encountered. These are grey and greenish marly claystones with interbeds of arenaceous siltstones and coarse quarts sand.

The total thickness of the Ashgill in the southwest area attains about 25 m as a maximum. In the central area it diminishes to 2 - 20 m in the northeastern part and only few meters in the southwestern area.

The presence of Lower Ashgill beds in the above mentioned areas is documented by the following fossils: *Nankinolithus granulata* (Wahlenberg), *Tretaspis seticornis* (Hisinger), *Panderia megalophthalma* Linnarsson, *Liocnemis recurvus* (Linnarsson) and *Dicellograptus* cf. *complanatus* Lapworth.

The same trilobite assemblage is known from the lower Ashgill beds in the Holy Cross Mountains (Kielan 1959), northeastern Poland (Modlinski 1973) and from areas in the Scandinavian - Baltic region (Männil 1966).

Upper Ashgillan index fossils are known from the Lopiennik IG 1 and Narol IG 1 boreholes only. Those are the trilobites *Mucronaspis* sp. and *M. mucronata* (Brogniart) which are characteristic among other in the so-called "Dalmanitina Beds" of the Holy Cross Mountains (Kielan 1959) and Sweden (Jaanusson 1963). Brachiopods that represent the so-called "Hirnantia fauna" of very broad paleogeographic distribution are also represented.

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SHELL MINERALOGY OF LINGULATE BRACHIOPODS FROM THE EAST BALTIC CAMBRIAN-ORDOVICIAN "OBOLUS PHOSPHORITE"

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Introduction

In northern Estonia and Ingria (Leningrad district of Russia), mass accumulations of lingulate brachiopod valves in Upper Cambrian to Tremadocian "Obolus Sandstone form deposits of 'Obolus phosphorite'. In modern stratigraphic terms, the phosphorite-bearing 'Obolus Sandstone' corresponds to the Ülgase, Tsitre and Kallavere formations in Estonia, and the Ladoga, Lomashka and Tosna formations in Ingria. The commercial phosphorite deposits occur mostly in the Kallavere and Tosna formations (Kaljo *et al.* 1986, Heinsalu 1987, Popov *et al.* 1989).

The present article deals with the shell mineralogy of the lingulates from the 'Obolus Sandstone'. Our objective is to study the lattice parameters of apatite in lingulate shells in statistically representative material and to discuss possible pathways leading to the observed mineralogical composition.

Previous Studies on Fossil Lingulate Shell Mineralogy

In a pioneer study of chemical composition of a sample from Yamburg in Russia (presently Kingissepp, Fig. 1) Schmidt (1861) concluded that the obolid shells are composed of amorphous fluorapatite close to vertebrate bone tissue and expressed the idea of their potential use as a fertilizer. Mineral composition derived from an independent chemical analysis of a sample from the same locality by Kupffer (1870) coincided with that suggested by Schmidt. Chemical analysis of the valves of '*Obolus*' (resp. *Ungula*; see Puura & Holmer 1993) from Vikarbyn and Boda in the Siljan District of Dalarna, Sweden was carried out by Andersson and Sahlbom (1900) who concluded that the fluorapatite of the fossil valves is of primary origin because its fluorine content was close to that of Recent *Lingula anatina*.

Mickwitz (1896) observed the lamellar structure of the cross-section of the shells of '*Obolus apollinis*' (resp. *Ungula ingrica*) from Jägala-Joa by light microscopy and suggested the presence of calcite and gypsum lamellae. Niggli and Beyer (in Wrangell 1920) studied the optical parameters of apatite in thin sections and concluded that the shell mineral is an apatite containing carbonate. Providing new chemical data from Ülgase and reviewing the earlier chemical analyses and thin section observations, Öpik (1929) concluded that the obolid shells are composed of a non- homogeneous mixture of phosphatic and other minerals and proposed the term 'Obolenphosphorit' or 'Obolus phosphorite'. Koch (1958) reviewed the earlier chemical studies leaving the question of the mineralogical composition open.

Loog (1962) carried out chemical analysis and X-ray powder diffraction analysis (XRD) of the valves of *Ungula ingrica* and *Schmidtites celatus* from Iru and Ontika in Estonia, concluding that the valves are composed of francolite, a carbonate fluorapatite. The conclusion was based on chemical data and the intensities of selected XRD peaks; lattice parameters were not calculated, but cited after an earlier unpublished report (Kurman *et al.* 1955, cited after Loog 1962).

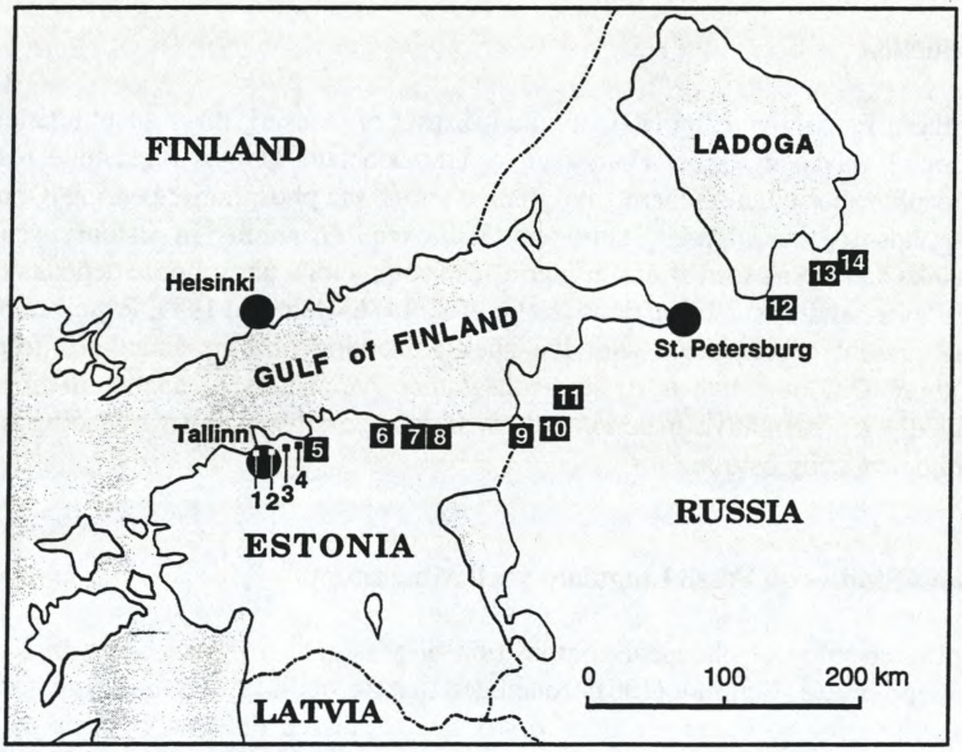

Fig. 1. Outline map of Estonia and Ingria (St. Petersburg region of Russia), with studies localities: 1 - Mäekalda, 2 - Iru, 3 - Ülgase, 4 - Valkla, 5 - Turjekelder, 6 - Toolse, 7 - Kalvi, 8 - Saka, 9 - Narva River, 10 - Kingisepp quarry, 11 - Suma River, 12 - Lava River, 13 - Volkhov River, 14 - Syas River.

Further studies have dealt with the mineral composition of phosphate in concentrates of "Obolus phosphorite" from the Maardu and Kingissepp mines, reporting the average stoichiometric formula (Veiderma and Veskimäe, 1971), lattice parameters: a=9.356, c=6.887 for Maardu and a=9.353, c=6.887 for Kingissepp and structural characteristics revealed by infrared spectroscopy (Veiderma & Knubovets 1972).

Ushatinskaya et al. (1988) carried out a comparative X-ray study of seven species from the Cambrian and Lower Ordovician of Canada, Siberia, Kirgizia and Leningrad region, including three species from the "Obolus sandstone" of Ingria (Table 1). They concluded that all the species are composed of carbonate fluorapatite with different stoichiometric proportions. Paying attention to the fact that the unit cell parameters of some species were stable in different localities, they did not make further conclusions because of the lack of statistically representative material.

Geological Setting

In the western part of the Baltoscandian palaeobasin, the Upper Cambrian and lowermost Tremadocian (Lower Ordovician) are represented mostly by alum shales, while in Estonia and Ingria, a sandstone facies prevails (Mens *et al.* 1993). Thin conglomerates in northern Öland, the South Bothnian submarine district and the Siljan district containing lingulate brachiopods identical to the species occurring in Estonia, indicate that the sandstone facies has extended westwards to Sweden (Puura & Holmer 1993).

According to biostratigraphic studies (Kaljo *et al.* 1986, 1988, Popov *et al.* 1989, Mens *et al.* 1993), the most extensive accumulations of the shelly phosphorites correspond to the *Cordylodus andresi-C. proavus* conodont zones. Most of the lithostratigraphic units related to the "Obolus Sandstone" (Heinsalu 1987) have characteristic features of high energy environments. In most sequences, only transgressive parts of the cycles, commencing with coarse-grained sandstones and terminating by fine-grained sandstones or siltstones have been preserved (Heinsalu *et al.* 1987, Popov *et al.* 1989). The sandstones contain lenses of black shales and are overlain by Dictyonema Shale, up to 7 m thick, considered to be formed mostly in an oxygen-poor environment (Heinsalu 1990). In some eastern localities in Ingria, the presence of trace fossils, e.g., *Skolithus*, suggests a more oxygen-rich environment (Popov *et al.* 1989). Lingulate shells can be found at many levels, though their higher frequency and larger average size in coarse-grained fractions indicate hydrodynamical sorting. "Obolus conglomerate", a coquina containing lingulate brachiopod shells and phosphatic pebbles occurs at the base of some transgressive cycles. The shells, redeposited from older parts of the section, and sometimes enclosed in the pebbles, can be encountered in the coquinas (Kaljo *et al.* 1986, 1988, Heinsalu *et al.* 1987, Mens *et al.* 1989, Popov *et al.* 1989).

Sampling and X-ray Analysis

Brachiopods were sampled from 14 outcrops in northern Estonia and Ingria (Fig. 1). The three most common obolid species forming mass accumulations, *Ungula ingrica* (Eichwald), *Schmidtites celatus* (Volborth) and *Obolus apollinis* (Eichwald), were systematically sampled. Selected samples of Middle and Upper Cambrian obolids *Ungula convexa* (Pander), *Oepikites macilentus* (Khazanovitch and Popov), *Oepikites koltchanovi* (Khazanovitch and Popov) and

Helmersenia ladogensis (Jeremejew) were taken for comparative analyses (Table 1), as were some phosphatized pebbles.

Shells were selected for X-ray powder diffraction analysis from all the samples. The shells or pebbles were powdered in an agate mortar and then analyzed with an X-ray diffractometer DRON- 0.5 with Ni-filtered Fe K[alfa] radiation using quartz as an internal standard. In the interval of 12-76 ° (2 [THETA]), 25 reflections of apatite were registered. The angle correction for apatite reflections was calculated against the 101 and 112 reflections of the quartz standard. Lattice parameters were calculated using the least squares method, including the data of all the 25 reflections (Aruväli, 1990). The structural CO_2 content was calculated by the formula proposed by Gulbrandsen (1969).

The results of the X-ray diffraction analysis and the comparative data from Ushatinskaya et al. (1988) are presented in Table 1. The presence of pyrite in some shells, observed by light microscopy, was also detected by X-ray analysis.

Table 1. Lattice parameters and calculated CO_2 content of apatite in lingulate brachiopod valves from the East Baltic Cambrian-Ordovician "Obolus phosphorite". Locality numbers correspond to those in Fig. 1.

Species	*a*	*c*	CO_2	n
Obolus apollinis	9.333-9.350	6.877-6.889	1.3-3.2	14
Schmidtites celatus	9.351-9.363	6.884-6.894	1.2-2.4	13
Ungula ingrica	9.335-9.366	6.890-6.905	1.0-3.4	12
Ungula convexa	9.346-9.367	6.880-6.896	1.4-2.2	6
Oepikites fragilis	9.348-9.356	6.878-6.895	1.0-2.0	3
Oepikites macilentus	9.362-9.363	6.881-6.896	0.3-2.1	2
Ungula inornata	9.352	6.890	2.1	1
Helmersenia ladogensis	9.350	6.886	1.9	1
Phosphatized pebbles, Ingria	9.342-9.347	6.885-6.888	2.3-2.5	3
Phosphatized pebble, Tallinn	9.36	6.903	3.13	1
Comparative data from Ushatinskaya et al. (1988)				
Helmersenia ladogensis	9.35	6.88		1
Oepikites koltchanovi	9.36	6.89		1
Keyserlingia buchii	9.33	6.88		1

Discussion

The data relevant for interpreting the factors and pathways governing the mineral composition in fossil lingulates can be obtained from a knowledge of shell structure and mineralogy of Recent lingulate brachiopods, ion substitutions in apatite during post- mortem mineralogical changes of apatitic skeletal parts, mechanisms of bacterial degradation and phosphatization of organic tissues, and apatite precipitation in sea water and pore space.

Structure and Mineral Composition of Recent Lingulate Shells

A study by X-ray diffraction, infrared absorption and X-ray microprobe analyses has shown that the shell mineral of Recent lingulates is F-containing carbonate-OH-apatite (LeGeros *et al.* 1985). The lattice parameters of the Recent lingulate shell mineral are between those of hydroxyapatite and fluorapatite (Table 2, comp. Slansky 1986, Table 5, p. 31).

In general terms, the shells of Recent lingulates of the genera *Lingula* and *Glottidia* are composed of alternating layers of organic tissue and carbonate-apatite (Iwata 1981, 1982, Watabe & Pan 1984). The proportion of the organic vs. mineral tissue appears to be taxon-specific: the organic content in the shells of *Glottidia pyramidata* and *Lingula anatina*_reaches 60 and 50 wt%, respectively (Pan & Watabe 1988a,b, Iwata 1981). The carbonate content in *Glottidia* apatite is higher than that in *Lingula*: 3.6 vs 1.8 wt%; the fluor content varies according to the shell layer and averages 2.58 in highly calcified layers of *Lingula* and *Glottidia* (Watabe 1990). A study of *Discina* by Williams et al. (1992) revealed that the mineralized and organic layers actually correspond to the zones of differential mineral content, termed as laminae. Still, the compact laminae, corresponding to the mineralized layer of earlier authors have significantly higher mineral content than other four types of laminae, equivalent to organic layer of earlier authors. For the purpose of the following discussion of diagenetic changes in buried lingulate shells, we use the simplified model of the alternating mineral and organic layers discussed above.

Table 2. Reported unit cell parameters of Recent lingulate brachiopod shell apatite.

Species	a	c	Reference
Lingula anatina	9.402	6.880	Watabe and Pan, 1984
Lingula anatina	9.389	6.880	LeGeros et al., 1985
Lingula anatina	9.373	6.869	Zezina et al., 1993
Lingula anatina	9.395	6.867	Zezina et al., 1993
Lingula anatina	9.383	6.859	Iijima et al., 1991
Lingula adamsi	9.396	6.880	LeGeros et al., 1985
Lingula reevei	9.383	6.871	Zezina et al., 1993
Lingula shantougensis	9.381	6.863	Iijima et al., 1991
Glottidia pyramidata	9.380	6.890	LeGeros et al., 1985
Glottidia pyramidata	9.394	6.890	Watabe and Pan, 1984
Discinisca lamellosa	9.383	6.871	Zezina et al., 1993

Mineralogy of fossil lingulate shells

The lattice parameters of fossil lingulate shells studied here and reported by Ushatinskaya et al. (1988) are in the range corresponding to apatite species between fluorapatite and carbonate fluorapatite (Table 1, comp. Slansky, 1986, Table 5, p. 31).

Compared to Recent lingulate brachiopods (Table 2), the studied fossil lingulates have considerably lower values of the unit cell parameter a. This systematic difference can be explained by substitution of different ions into the apatite lattice (Hughes *et al.* 1989).

The fluor-containing carbonate-OH-apatite in hard tissues of living brachiopods and vertebrates has a tendency for crystallographic maturation by replacement of OH by F and PO_4^{3-} by CO_3^{2-} and F^-. Thus, additional F is incorporated to the lattice with carbonate, as the apatite becomes a carbonate-fluorapatite. According to Lucas and Prévôt (1991), this aquisition of F and carbonate is initiated *in vivo* and continues during diagenesis.

Provided that most apatite species produced in skeletal biomineralization in Recent invertebrates and vertebrates have relatively high hydroxyl contents (Watabe 1990, Skinner 1991), we are inclined to interpret the significantly different mineralogy of fossil shells as influenced in some extent by diagenetic alteration.

Other processes that can influence the composition of shell apatite are bacterially mediated replacement of the organic tissue by a new generation of apatite, degradation of organic matter and subsequent precipitation of carbonate-F-apatite to the free space. Furthermore, the substitution of CO_3^{2-} vs PO_4^{3-} in the course of diagenetic alteration decreases crystal size and increases apatite solubility (Jahnke 1984) that in turn increases the possibility of recrystallization. The precipitated cryptocrystalline carbonate apatite (CCP) may also dissolve if the geochemistry of the microenvironment changes.

Pyritized and phosphatized fossil bacteria-like bodies, as well as living Recent chemolitotrophic bacteria, have been isolated from the studied fossil lingulates (Nemliher 1993). Experimental work has demonstrated that bacteria are able to mediate apatite formation through the action of their enzymes (Lucas & Prévôt 1991). In some cases, rapid phosphatization can lead to exclusive preservation of organic tissues, as has been demonstrated both by laboratory experiments and fossil record, e.g., Upper Cambrian 'orsten' arthropods from the western alum shale facies of "Obolus phosphorite" basin (Briggs & Kear 1994, Walossek 1993 and references therein). In other cases, organic tissue can be completely degraded by anaerobic bacteria. An experimental study by Arnosti et al. (1994) has shown that degradation of polysaccharides by consortia of anaerobic bacteria can be rapid in anoxic environments, that is, by different groups of bacteria working in concert in transformation of complex substrates. Possible pathways of bacterially mediated mineralization of organic tissue depending on burial conditions and early diagenetic environment can be viewed in the context of the sequence of oxic, suboxic, anoxic zones in the bottom water and sediment column (e.g., Allison 1988).

After burial in sediment, a pore-space like microenvironment may develop in a lingulate valve, where the organic matter or the space left after its degradation is enveloped by the mineral part. Light microscopy observations supported by X-ray analysis and back-scattered electron imaging show that the space occupied in vivo by organic tissue is filled with pyrite in many valves. The sedimentary pyrite precipitation is considered to be initiated by sulphate reducing bacteria (Berner 1984) and to occur below oxic-suboxic interface in the water or sedimentary column or in pore space with similar geochemical regime (Allison 1988). The observation of distinct alternating layers of phosphate and pyrite in thin sections provides a clue for understanding analogous precipitation of secondary apatite, as close phosphate phases cannot be visually distinguished.

In microenvironments richer in oxygen, an expected mineral to replace the organic tissue is carbonate-fluorapatite (syn. francolite; Clark 1993). We assume that the organic matter was first degraded by bacteria and the free space was subsequently filled with secondary apatite. This assumption is supported by SEM observations in *Ungula ingrica* by Holmer (1989, Fig. 14, p. 32): the cryptocrystalline calcium phosphate (CCP) fills the space between rod-like baculae, the preserved parts of the original structure. As yet, no evidence has been found for phosphatization of organic tissues of the "Obolus Sandstone" lingulates.

Infraspecific variation range of the lattice parameter a is from 0.01 to more than 0.03 Å, being most stable in *Schmidtites celatus* and *Obolus apollinis* and least stable in the species of the genus *Ungula*. For instance, the variation of the lattice parameters of *Ungula ingrica* is as wide, and in about the same range, as lattice parameters for eight Cambrian and Early Ordovician lingulates reported by Ushatinskaya et al. (1988).

Preliminary results of the shell structure studies to be published separately have revealed that among the studied genera, *Ungula* has the thickest organic layers and, consequently, the highest content of organic tissue that has been subsequently replaced by secondary apatite. The varying amount of secondary apatite (CCP), differentially filling the space between mineral layers, results in a varying bulk composition of the shells.

Compared to *Ungula*, the shells of a smaller obolid, *Schmidtites celatus* had thinner organic layers. The variation of lattice parameters a and c is only slightly over 0.01 Å (Table 1). The lattice parameters of a related genus *Oepikites* are approximately in the same range as for *Schmidtites.*

The shells of *Obolus apollinis* from Ingria have the lowest values of lattice parameter a and a high carbonate content, close to the corresponding values of the phosphatized pebbles from the same area. Preliminary SEM observations suggest that the shell apatite has been recrystallized in many cases. Thus, it can be suggested that the carbonate content in the recrystallized shells approached the equilibrium with seawater (Jahnke 1984). The question of the impact of possible change to shell mineralogy in non-marine conditions remains open.

Conclusions

In contrast to the opinion expressed in most earlier studies, the mineralogical composition of fossil and Recent lingulate shells differs significantly. Recent lingulate shell mineral is F-containing carbonate hydroxylapatite with a F-content higher than in dahllite with the lattice parameters a=9.38-9.40 and c=6.87-6.89. The bulk composition of the studied fossil shells corresponds to the apatite species between fluorapatite and carbonate fluorapatite with the lattice parameters a=9.33-9.36 and c=6.87-6.90. Previously reported lattice parameters of fossil lingulates (Ushatinskaya et al., 1988) are also within this range.

The original composition of the fossil lingulate shells was possibly close to that of the Recent shells. In each particular case, the change of the mineralogical composition of a fossil shell could have been caused by one or a combination of a variety of processes, such as (1) diagenetic substitution of OH^- by F^- and PO_4^{3-} by CO_3^{2-} and F- , (2) degradation of organic tissues by bacteria and precipitation of CCP (or pyrite) in the free space and (3) dissolution and recrystallization of CO_3^{2-}- enriched fossil shell apatite due to high solubility.

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THE LATEST ORDOVICIAN RUGOSE CORALS IN BALTOSCANDIA

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Introduction

In late Ordovician Hirnantian times one of the largest periods of extinction in the Phanerozoic of families, genera and species of both benthic and planctic marine faunas is well documented. Alterations in the sedimentation during Late Ordovician clearly indicate a marked eustatic sea-level change in form of a stepwise regression, in all estimated to be approximately 100 metres (Brenchley & Newall 1984, Brenchley 1984). The lowering of the sea level was most probably due to an intense glaciation on northern Gondwana where landbased glacial deposits (e.g. tillites) are found in Saharan Africa, in South America and in Saudi Arabia (see Brenchley & Newall 1984 for further references). The falling sea level during the Hirnantian times significantly reduced the habitable areas for benthic forms, and the biotopes probably became condensed into narrow bands on the continental slope. Marked sea-level changes influence on depths, bottom conditions, sedimentation and food supply of marine benthos. Areal reduction of shallow water biotopes and alteration of ecological factors at a falling sea level will often lead to marked extinctions of sessile benthos. Even deep water benthos can be influenced by starvation due to less production of deep sea plankton. (Rohling, Zachariasse & Brinkhuis 1990).

Contemporaneous changes to lower ocean temperature as expected in Late Ordovician may have affected the plankton production considerably and probably caused extinctions due to starvations of the filterfeeders and their predators. The deep flooding of the continental shelves at the very end of the Hirnantian, and release of fresh water from melting ice caps, forming a surface layer of low salinity on the oceans, can declare the late Hirnantian extinctions events (Brenchley 1984).

It is well-documented that all these alterations of the marine environment in latest Ordovician led to pronounced worldwide alteration and extinctions of rugose and tabulate coral.

In North America extinctions and origins of solitary rugose corals taxa in latest Ordovician and earliest Silurian have been reported by Elias in several papers and were summarised by him (Elias 1988). Elias & Young (1992) gave further information about tabulate corals and colonial rugose corals.

Lin & Webby (1989: 207-218) summarised the biostratigraphic relationships of Australian and Chinese Ordovician corals and stromatoporoids including information on the Hirnantian faunas from those areas.

Kaljo (1982: 52-55) has briefly summarised the early Palaeozoic rugose and tabulate corals from Estonia including the faunas from the Hirnantian Porkuni stage. These faunas were earlier described in numerous papers by Dybowski, Reiman, Kaljo and Klaamann and Weyer.

Lindström (1880) described a few species of Ordovician rugose and tabulate corals from Sweden. Some taxa were also described from Norway and especially from the Oslo Region by Kiaer (1899), Wedekind (1927), Scheffen (1933), and Spjeldnæs (1961). Neuman (1968, 1969 and 1975) treated several taxa of Upper Ordovician solitary, rugose corals from Scandinavia. Some Hirnantian colonial rugose corals and tabulate corals were described by Hanken (1974, unpubl. thesis) from the Ringerike area in the Oslo District, and were also mentioned by Hanken & Owen (1982). Later Hanken (1979: 97-100) described *Rhabdotetradium frutex* Klaamann from that fauna. Neuman (1982: 34) summarised the Upper Ordovician rugose corals from the Oslo Region, and major trends in the evolution of rugose corals were later discussed by him (Neuman, 1984).

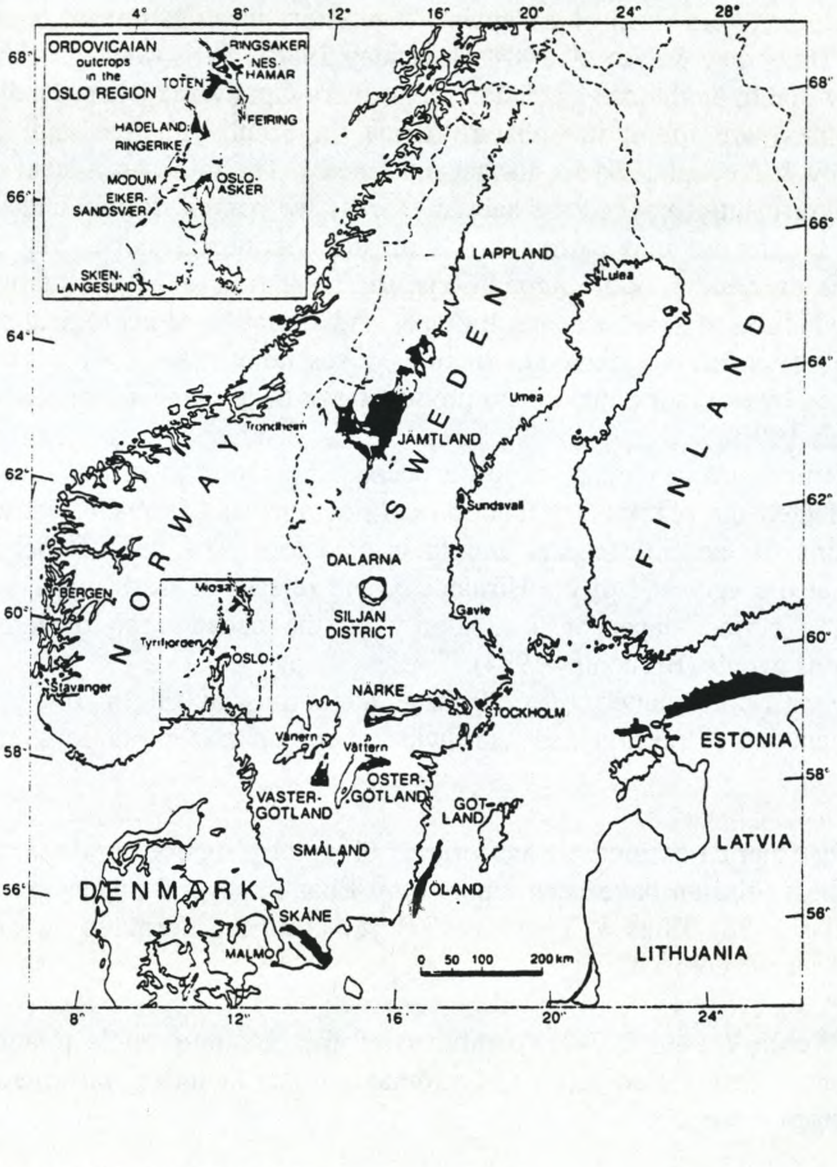

Fig. 1. Ordovician outcrops in Baltoscandia

Terminology

The terminology used in this paper is described by Hill (1983) and Neuman (1984 and 1988) and some are here recapitulated. *Fixosessile* corals have well-developed holdfasts with which they are fixed to hard bottom for all their life. *Liberosessile* corals lack holdfasts and (but for a short period on larval stage) live recumbent on soft bottom. A *fossula (sensu stricto)* is marked pit in the calice around e.g. the cardinal septum formed by depressed tabulae, tabellae or dissepiments. In a *septofossula* the pit is formed by a considerably shortening of the cardinal septum in latest precalicular stage and tabular structures are not depressed. A *pseudofossula* is the term for unusually large interseptal spaces around septa where septal incersion takes place.

Baltoscandian Corals

The Baltoscandian faunas of rugose and tabulate corals are often clearly facies-related and the following combinations between facies and taxa are seen in the latest Ordovician.

Sweden

The Hirnantian Tommarp Fm. bearing corals is found in Östergötland (locs Borenshult and Råsnäs), Västergötland (loc. Ålleberg N.) and in the Siljan District (Locs Kullsberg and Nittsjö). This formation (Jaanusson 1982a and b) mainly consists of calcilutites and mudstones and rugose corals are normally very rare (Neuman 1969, 1975). Liberosessile streptelasmatids like the pseudofossulate *Streptelasma linnarssoni* and the septofossulate *Streptelasma unicum* are represented. The coral fauna in Östergötland is more diverse and in addition to *S. Unicum* the following endemic species are present; *Streptelasma ostrogothicum, Helicelasma simplex* and *Borrelasma crassitangens*. The presence of corals in these beds could be due to that the Hirnantian regression shaped new shallow biotopes.

Erratic boulders of carbonate rock from Oil (or Ölje) Myr on Gotland, probably derived from the bedrock of Prokuni age north of the island, , contains the solitary, liberosessile streptelasmatid *Bodophyllum oilense* Neuman (1969:59) and a few other still undescribed streptelasmatid corals.

In the Siljan District the Boda limestone is approximately equivalent to the Pirgu and Porkuni stages in Estonia and to the Jerrestad and Tommarp fms in Swedish areas. The Boda limestone represents a stromatactis-bearing carbonate mound facies. These mounds are 100-140 m thick, and are very numerous in the Siljan District (Jaanusson 1982a: 25-29). Their flank deposits consist of stratified, variegated calcarenites with argillaceous intercalations, locally with numerous corals. Solitary streptelasmatid corals are well represented by the pseudofossulate liberosessile *Bodophyllum osmundense* and the pseudofossulate fixosessile *Streptelasma primum, S. cyrtum, Grewingkia billateralis, G. contexta* (see Neuman 1969). Sparsely represented are the very large solitary, liberosessile streptelasmatids with a cardinal fossula *sensu stricto Grewingkia* cf. *buceros* and *Kaljolasma giganteum*. Non-streptelasmatid fixosessile species described are *Tryplasma spinulatum* and *Paliphyllum suecicum* (Neuman, 1968). No colonial rugose corals seem to be present in the Boda Limestone. A loose specimen of *Entelophyllum* sp. found in the Kallholn quarry is most probably emanating from the Kallholn Fm. of early Llandoverian age (Jaanusson 1982a¨32).

Norway

From carbonate beds of Rawtheyan age in the Oslo District the following solitary liberosessile streptelasmatid corals were described by Neuman (1969): *Streptelasma primum, S. eccentricum, Grewingkin buceros, G. anguinea*, and *Bodophyllum euthum.* In addition the solitary liberosessile lambelasmatid coral *Densigrewingkia pyrgoidea* is sparsely found. Most of these species are common in the Oslo-Asker area in the Langåra Fm., in the Skien-Langesund area in upper part of the Herøya Fm. and in the Bønsnes Fm. in Ringerike. The Langåra Fm. in the Asker area reaches high up in the Hirnantian (Brenchley & Newall 1975) and specimens of *Palaeophyllum* sp. was collected and reported by Kiær (1901: 7 and 14) from two localities.

In the Oslo District Hirnantian regressive shoalfacies is represented by the Langøyene sandstone-shale Fm. in Oslo-Asker, and Skien-Langesund areas (Brenchley & Newall 1975). Rugose corals adapted to such shallow water high energy environments, represented by more or less calcareous laminate sandstones, are liberosessile forms like *Bodophyllum duncanae*, the septofossulate *Ullernelasma svartoeyensis,* and some very small species of *Helicelasma* and *Borelasma.*

In the Ringerike area (locs Ullerntagen, Stamnestangen and Vestre Svartøy), a sequence of Hirnantian age was thoroughly mapped and described by Hanken (1974, unpubl. thesis) and shortly summarised as the unnamed Fm. by Owen & Hanken (1982) and tentatively brought to the Langøyene Fm by Owen et al (1990: 44). Hanken (1974) destinguished four succeeding main facies and clear indication of erosion episodes between these. In his oldest facies type 1 of sandstone and biosparitic calcarenite no corals were present. Facies 2 (by Hanken called coral-stromatoporoid-algal facies) is a shallow water carbonate ramp with biostromes and very small bioherms developed in the upper part where several colonial rugose corals are present, viz. *Palaeophyllum incertum, P.* sp., *Tryplasma regularis*, and *Cyathophylloides ringerikensis* as well as numerous tabulate corals including the earliest tetradiid coral *Rhabdotetradium frutex.* Facies type 3 of calcareous sandstone only contained the solitary, septofossulate streptelasmatid coral *Ullernelasma svartoeyensis*, and in facies type 4 of sandy biosparitic calcarenite rare specimens of *Palaeophyllum* sp. and some tabulate corals were found.

In an offshore drillcore south of Kristiansand in southern Norway Upper Ordovician and Lower Silurian beds were penetrated. The core is rich on rugose corals from the Rawtheyan and Hirnantian interval. As in the Oslo-Asker, Skien-Langesund and Ringerike areas *Grewingkia buceros, Streptelasma primum* and *Bodophyllum euthum* were present in beds of Rawtheyan age. The lambelasmatid *Densigrewingkia pyrgoidea* is very common in siltstone and shale beds of Rawtheyan, Hirnantian and even Rhuddanian age. *Palaeophyllum* cf. *incertum* is found already in late Rawtheyan (one specimen). *Bodophyllum duncanae, Ullernelasma svartoeyensis* and at least two species of *Helicelasma* are present in the Hirnantian stage only.

Conclusions

The present knowledge of the latest Ordovician and early Silurian rugose coral faunas is far from complete and a recently started Estonian-Norwegian cooperation project will include detailed research of coral-stromatoporoid faunas in the latest Ordovician of Baltoscandia. Some important trends in the development of these faunas are, however, given below.

In Rawtheyan times the Baltoscandian rugose coral fauna in more or less argillaceous limestones, often nodular and interbedded by shale, was dominated by liberosessile solitary species of the streptelasmatid genera *Steptelasma, Grewingkia, Helicelasma,* and *Bodophyllum.*

Species of lambelasmatid corals are rare in Rawtheyan and Hirnantian e.g. *Desigrewingkia pyrgoidea* and *Neotryplasma rhizobolon* and seem to be long-ranging taxa in mudstone environment. In the Siljan District the flanck facies of the Boda limestone carbonate mounds of Rawtheyan-Hirnantian age has a pronounced dominance of fixosessile pseudofossulate solitary streptelasmatid corals of the genera *Streptelasma* and *Grewingkia.* Also fixosessile species of *Tryplasma* and *Paliphyllum* are represented. Liberosessile species are rare (e.g. *Bodophyllum osmundense* of Hirnantian age) also including two large fossulate forms.

Corals with a fossula *sensu stricto* (all are liberosessile are very rare in the Upper Ordovician of Baltoscandia. *Grewingkia buceros* from Rawtheyan beds all over Baltoscandia, *Bighornia orvikui* from Pirgu stage and *Kkaljolasma giganteum* from Porkuni stage in Estonia. The first and the last species is also rarely found in the Boda Limestone in the Siljan District. Several species of Rawtheyan rugose corals became extinct before Hirnantian.

During Hirnantian times the repeated regression events shaped several new biotopes and taxa adapted to these new biotopes appeared.

In Sweden the Hirnantian Tommarp Fm. of more or less argillaceous limestone rarely contain liberosessile solitary streptelasmatids like the septofossulate *Streptelasma unicum* and pseudofossulate *Streptelasma.* In Östergötland the comparatively rich Borenshult fauna includes endemic species of *Streptelasma, Helicelasma* and *Borelasma.*

In the Oslo Region Hirnantian is developed in several facies due to the regression events (Brenchley & Newall 1975).

Shoal facies, represented by the clastic Langøyene Fm. containing laminated sandtone and thin limestone beds, gradually became more and more dominant and a great number of sedimentation brakes are registered. In such facies solitary streptelasmatid septofossulate species of *Ullernelasma* and *Streptelasma* and several small pseudofossulate streptelasmatid corals of the genera *Bodophyllum, Helicelasma, Borelasma* are common.

The carbonate ramp facies described by Hanken (1974) from the Ringerike area (Hanken 1974) also included biostromes and small bioherms mainly consisting of algae, stromatoporoids tabulate corals e.g. *Rhabdotetradium frutex* and colonial rugose corals e.g. *Palaeophyllum incertum, Palaeophyllum* sp., *Tryplasma regularis,* and *Cyathophylloides* sp.

In Estonia similar trends as in the Oslo Region is found in the Hirnantian Porkuni Beds (Kaljo 1990:54). Also there several species of colonial rugose corals appeared, viz. *Palaeophyllum,* *Strombodes* and *Holacanthia* (= ? *Tryplasma*) as well as tabulate corals including *Rhabdotetradium frutex.*

Notable is this first appearance of palaeophyllid and tetradiid corals Baltoscandia as such forms are much earlier represented in China, Australia and North America.

The very marked Early Silurian Transgression is easily observed in most Baltoscandian areas but only in the Oslo District and in Estonia the basal Silurian has coral-bearing facies.

In the Oslo Region the rugose corals of the basal Sælabonn and Solvik Fms are mainly small, liberosessile streptelasmatids species of *Helicelasma, Borelasma, Streptelasma*, and the septofossulate *Rhegmaphyllum.* Some few Ordovician forms remain in these beds e.g. *Bodophyllum euthum, Ullernelasma svartoeyensis,* and the lambelasmatid *Densigrewingkia pyrgoidea* although most Ordovician forms seem to become extinct. Among new faunal elements is a species of *Calostylis.*

The diversity increases rapidly due to the upwards shallowing sequences and in Idwian and Telychian when most genera typical for the rest of the Silurian are developed.

Based on the facts above the cold water currents bringing the *Hirnantia* fauna apparently did not to any large degree influence the subtropical Baltoscandian shallow sea areas and the rugose coral fauna. During the Hirnantian the diversity of colonial rugose coral genera increased rapidly. Palaeophyllid and tetradiid corals which now migrated to Baltoscandia are looked upon as beeing warm water indicators. Hirnantian oolitic limestone and organic buildups also support a high temperature shallow sea areas.

Although the very large transgression at the Ordovician - Silurian boundary reduced suitable biotopes for rugose corals, most of the Ordovician solitary streptelasmatid genera are still represented in basal Silurian beds often with new species. Colonial rugose corals and tabulate corals suffered much more of the loss of shallow water biotopes and several genera become extinct.

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LOWER/MIDDLE ORDOVICIAN CONDENSED PHOSPHATE-RICH BED IN THE PODLASIE DEPRESSION, EASTERN POLAND

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Summary

At the transition between the Lower and Middle Ordovician in the Podlasie Depression, eastern Poland, a thin layer of phosphatic bio-oomicrite and bio-oo-pelmicrite exists. It contains phosphatized bioclasts, coated grains including ooids, discontinuity surfaces and a rich assemblage of endolithic microborings. The stratigraphic sequence shows the features of condensed deposits with litho- and paleontological condensation and stratigraphical gaps.

Organic matter resulting from the accumulation of dead organic material, together with organisms inhabiting the bottom sediment constituted an organic-rich layer during the slow or no deposition. The decomposition of organic matter caused the liberation of phosphorous and thus the enrichment of bottom and pore waters with phosphatic ions. The process of decay changed the pH and Eh conditions which could led to the precipitation of phosphates from the supersaturated with phosphatic ions bottom water. There were two processes of the formation of apatite; biochemical precipitation from supersaturated bottom- or pore water with calcium phosphate and replacement of calcite by calcium phosphate.

The phosphate-rich deposit originated probably in photic zone and relatively calm condition.

Geological setting

The Podlasie Depression is situated in the south-eastern part of the Baltic Basin and the lithological and stratigraphical development of Ordovician deposits is most closely comparable to that of North Estonia and Lithuania (Fig. 1). The main type of sediment is grey biomicrite with thin marly interbeds, except the most lower part of Ordovician which is sandy and glauconitic. The thickness of the whole Ordovician varies from 60 to 100 metres and is known only from boreholes at a depth of 400 to 700 m.

The most complete Lower-Middle Ordovician sections come from the following six boreholes; Warpechy Nowe IG-1, Rajsk IG-2, Widowo IG-1, Troscianica IG-1, Tyniewicze IG-1, Zubowo IG-1, drilled near Bielsk Podlaski (eastern part of the Podlasie Depression). The uppermost lower and the lowermost Middle Ordovician of the investigated area is represented by phosphatic biomicrite, bio-oomicrite and bio-oo-pelmicrite up to a dozen or so cm thick with discontinuity surfaces sometimes of hardground character and rich in phosphatic grains (Fig. 1).

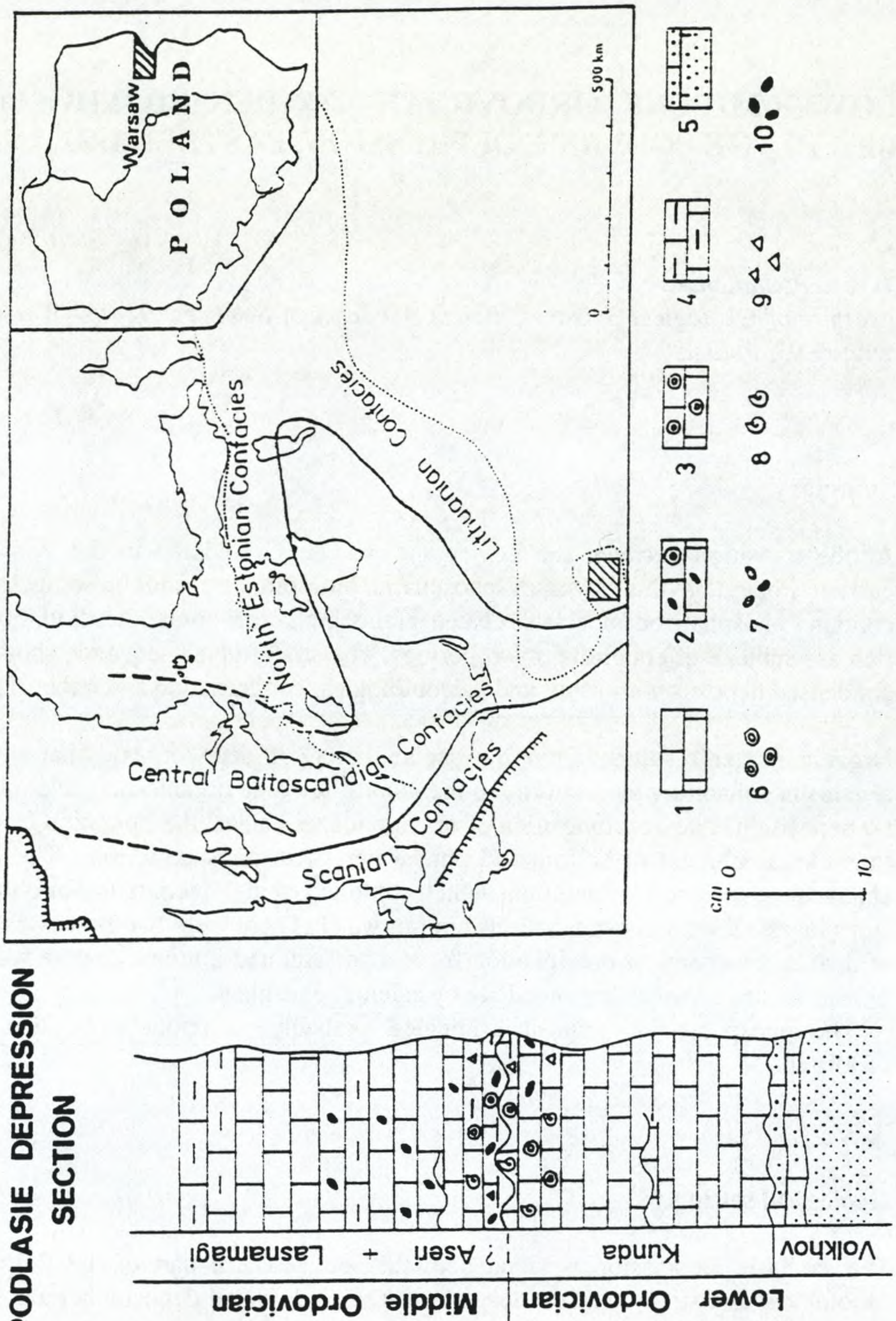

Fig. 1. Locality map and generalised stratigraphical section across Lower/Middle Ordovician boundary in the Podlasie Depression (eastern Poland). 1) biomicrite, 2) phosphatic biooopelmicrite, 3) phosphatic biooomicrite, 4) marly biomicrite, 5) glauconite and glauconitic limestone, 6) phosphatic ooids and superficial ooids, 7) phosphatic peloids, 8) phosphatized bioclasts, 9) phosphatic grains of unknown origin, 10) goethitic coated grains.

The stratigraphy of the transitional beds has been based on conodonts, whose distribution proves paleontological condensation. The following conodont zones have been distinguished; *Amorphognathus variabilis, Eoplacognathus suecicus* and *Eoplacognathus reclinatus*. In some sections conodont fauna is mixed or one conodont zone is missing.

Litho- and paleontological condensation, great amount of the skeletal and non-skeletal grains, presence of microborings, phosphatization of grainy components and discontinuity surfaces are the main features of the condensed section near the Lower/Middle Ordovician boundary in the Podlasie Depression.

Phosphatic components

As a rule grains constitute most of the phosphatic part of the deposit, the matrix being usually calcitic or clayey. Phosphatic matrix also occurs, but is rather rare and is usually limited to thin layer close to the discontinuity surfaces.

Phosphate replacement of carbonate skeletal debris together with the occurrence of primary apatitic grains are a major feature of the deposit.

The following types of phosphatic grains have been distinguished; phosphatized bioclasts, ooids and superficial ooids, peloids and ovules with microboring casts inside.

1. Phosphatized bioclasts.
   In the case of calcitic bioclasts we have completed replacement of calcite by calcium phosphate. The most frequent are; gastropods dominated by one species, ostracods, fragments of bryozoa and echinoderms and trilobites. Organisms are perfectly recognisable in every detail, even though their chemical make-up has changed.

2. Ooids and superficial ooids (Fig. 2).
   The ooids are usually ellipsoidal in form most often with bioclasts (echinoderm or trilobite fragments) as the nuclei. Glauconitic or goethitic grains may also form the nuclei of the ooids. Skeletal grains are usually earlier goethitized. Electron microprobe analysis shows that iron, calcium phosphorous, aluminium and silica dominate in the nucleus with a small amount of titanium, potassium and sulphur. In true ooids ooid lamination is well developed and consists of several fine layers. Here phosphorous and calcium dominate while silica and aluminium are in small amounts being restricted mainly to the area between phosphatic laminae (Fig. 2C). Structures, which eventually, could be interpreted as organic remnants were observed sometimes between phosphatic laminae. In some grains two phases of coating can be observed (Fig. 2A). In that case smaller phosphatic ooid occurs between the phosphatic laminae of greater one. Taking into account the morphology of whole structure it must be mentioned that small grain is; 1) *in situ* in relation to the big one and 2) it had to lie calm on the sea floor during the coating. Thus I would suggest relatively clam bottom condition during the growing of phosphate laminae.

3. Peloids.
   Among the phosphatic grains, peloids are the most common and characteristic component. The peloids are undifferentiated light to dark brown ovoid grains. Most are well-rounded with smooth boundaries, uniform in size and shape, well-sorted. Their diameter does not

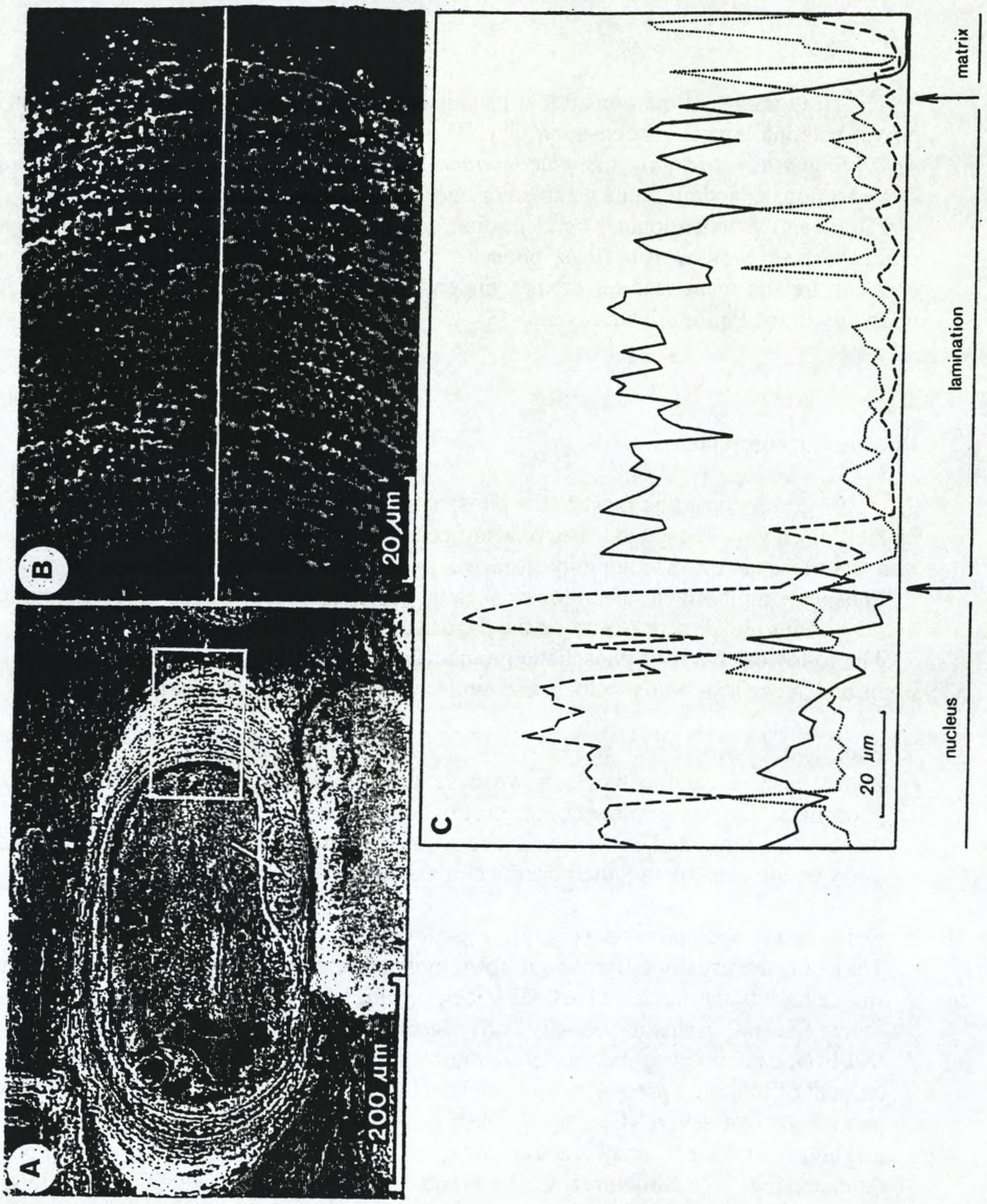

Fig. 2. Phosphate-rich ooid. Warpechy Nowe IG-1 section, depth 641,3 m, uppermost Lower Ordovician. A, B-SEM microphotographs, C-microprobe profile.

exceed 1 mm. A systematic study of these peloids indicates that they are the result of convergent diagenetic processes, the more important being; growth around a foreign grain, phosphomicritization of skeletal fragments, phosphatization of faecal pellets (Cook, Elqueta 1986, Balson 1980, Soudry, Nathan 1980).

Some peloids under study have a grain in their centre. In almost all cases this is a skeletal grain, more or less phosphatized with phosphate coating. Complete phosphatization leads to a mixing of centre and coating, and the end product is a peloid which cannot be differentiated from other peloids of probably different genesis.

Some peloids occurring in the investigated sections formed by the phospho-micritization of trilobite fragments. This process is very similar to the one described by Soudry (1979), and concerning the phosphomicritization of bone fragments by algae in the Negev (Upper Cretaceous - Lower Eocene) phosphorites.

4. Ovules with microboring casts inside.
The next very common type of phosphatic component are grains from outside resembling the peloids, but having a rich assemblage of microborings inside. By comparison with recent ones, on the basis of shape, dimensions and mode of branching several types of endolithic forms have been distinguished. The nature of the microboring filaments may suggest that they were made probably by autotrophic organisms most probably cyanobacteria or algae even though some of the most discrete forms may belong to heterotrophic micro-organisms. Studies of modern and ancient endolithic assemblages suggest that they are useful as environmental and paleoenvironmental indicators especially of the conditions at the sediment water interface during deposition (Boekschoten 1966, Budd, Perkins 1980, Golubic *et al.* 1984, Hoffman 1985, Podhalanska 1984, Rooney & Perkins 1972). After comparing available data concerning recent and ancient endolithic forms with those found in the investigated sediments, it can be stated that the material for pelletal grains seems to have formed within the photic zone, probably in the intertidal facies zone by means of presence of autotrophic micro-organisms in observed assemblage of endoliths. I would suggest that these described pelletal grains might be crushed, reworked, ab-raded and slightly redeposited fragments of microbial (algal?, cyanobacterial?) mat.

A. General view of the investigated ooid. Note 2 stadia of coating (older stadium is arrowed).
B. A part of ooid lamination marked on photos A. White line microprobe profile.
C. Distribution of studied elements along microprobe profile on photos B. 1-P, 2-Fe, 3-Si. Note the variability of intensity of studied elements in cucleus, lamination and matrix. Microprobe.

X-ray investigations show that all the boring tunnels are coated or infilled with apatite. Apatitic coatings developed around borings usually have the form of small sperical structures. This ball-shaped ultrastructure is thought to be a result of rapid precipitation of an amorphous calcium phosphate within a micro-environment supersaturated with phosphatic ions (Krajewski in press). The precipitation of calcium phosphate occurred immediately after the decomposition of endolith filaments and is strictly connected with the process of decay of organic matter as a source of phosphorous.

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Phosphogenic event near the lower/middle Ordovician boundary, general model

The Lower/Middle Ordovician limestone in the Podlasie Depression, contains phosphatic constituents such as clasts, ooids, peloids and phosphatic hardgrounds. The upwelling model presented by Kazakow (1937) and modified by Bentor (1980) and Sheldon (1981), is not appropriate for wide and shallow epicontinental Baltic Sea with extremely low relief (Jaanusson 1973, 1984). An alternative model assumes a great role of microbial activity in phosphogenesis. The source of the phosphate is generally considered to be the phosphorous liberated by decaying protoplasm of marine organisms (Birch 1980, Gulbrandsen 1969, Froelich *et al.* 1979, Krom, Berner 1981). The liberation of phosphorous from degraded organic compounds on the sea floor is a predominant microbial process involved in apatite formation. A close association of benthic microbial activity and the formation of apatite can be widely traced in recent and ancient phosphogenic environments. The role of the biological activity has also been stressed by Nordlund (1988) for the Lower Ordovician phosphatic series of northern Öland, Sweden.

Event of high organic productivity caused a change of chemical state near the sediment-water interface. The more decay of organic matter proceeded the lower was pH of the environment thus exceeding the limit of calcium carbonate precipitation. Event of high supersaturation with phosphatic ions localised in space an time increased the likelihood of phosphate precipitation close to the bottom water boundary.

I believe that in the case of the investigated deposits apatite formation is related to a pulse of supersaturation and that microbial redox processes helped to precipitate apatite, creating different phosphatic bodies (ooids, peloids) and improving preservation of microbiota (earlier described microborings and bioclasts) (Podhalanska 1992).

Probably, there was an initial step to produce the exceptionally high organic productivity system. The primary input of nutrient water could be connected with a land weathering or volcanic activity. In some areas of the Baltoscandia contemporary phosphatic facies is associated with volcanic ash (Sturesson 1992).

Final remarks

1 The section of the Podlasie Depression described here contains all the components of a classical condensed section; hiatus, phosphorite bed, discontinuity surface sometimes of hardground character, accumulation of organic matter. It was connected with a marked period of low or non-deposition near the Lower/Middle Ordovician boundary of the investigated area.

5 Probably, there was an initial step to produce the high organic productivity system (enhanced nutrient input from land or fumaroles).

6 The origin of the phosphate-rich layer at the Lower/Middle Ordovician boundary is a combination of biological, chemical and mechanical factors and was strictly connected with

- a) high organic productivity
- b) slow sedimentation
- c) formation of fairly anoxic organic-rich bottom sediment
- d) liberation of phosphorous as a result of decomposition (bacterial oxidation) of organic matter, enriching the bottom and pore waters with phosphate in mildly reducing environment
- e) precipitation of calcium phosphate in mildly reducing environment
- f) formation of phosphatic grains and phosphatization of bioclasts and matrix
- g) eventual reworking and transport of the grains

7 Possible processes of the formation of apatite;

- a) primary process - biochemical precipitation from supersaturated bottom or pore water with calcium phosphate (ooids, peloids, microbial mat, microboring coatings?)
- b) replacement - in case of calcitic bioclasts we can see complete replacement of calcite by calcium phosphate. Organisms are perfectly recognisable, even though their chemical make-up has changed.

8 The preferential occurrence of phosphate on organic substrates suggests a dual role of organic matter in phosphogenesis

- a) as a source of phosphorous
- b) as a substrate on which phosphate preferentially nucleates

Phosphate-rich sediment originated probably in intertidal environment, in photic zone and is relatively calm condition. In general my impression is that for the occurrence of phosphatic facies, at the transition between the Lower and Middle Ordovician, chemical changes of bottom water were directly responsible. These chemical changes were directly connected with a pulse of overproduction of organic matter, which was not fully consumed and fell to the bottom in the for of a “rain of dead remains”. This formed a thin, slightly anoxic mat which was the direct source of phosphate.

The primary input of nutrient water connected with changes of sea level or with volcanism should be taken into account.

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A MERIDIONAL FACIES CROSS-SECTION OF THE LLANVIRN - LOWER CARADOC, EASTERN ESTONIA

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Introduction

The Llanvirn - Lower Caradoc (Kunda - Kukruse age) of Estonian Ordovician (largely corresponding to the Whiterock Series of North America (Ross & Ethington, 1992)) can be distinguished in several aspects as a specific epoch. It appears as the main section of: (1) ooid distribution; (2) kukersite distribution; (3) the most frequent distribution of discontinuity surfaces and the best expressed facial zonation of them (based on their impregnation types). Near the lower boundary of the above section the content of glauconite decreases distinctly, near its upper boundary the first K-bentonites are registered.

The boundaries between different lithofacies are transitional. The thickness of different stratigraphic units appears to be spatially rather stable and their changes - comparatively regular. All the stages (especially the Uhaku[2] and the Kukruse) (Männil 1986) have an excellent subdivision, the separate beds and complexes of beds are well traceable over long distances. Along with the widely exposed strata at the clint area there is a great number (>150) of preserved and rather well investigated drillcores especially from the northern and central Estonia. The performed cross-section (Figs 1 & 2) is based on the results of the aforesaid data.

Distribution of the characteristic lithological features

Kukersite is a specific mixed rock type characteristic especially of the Middle Ordovician (from the Kunda to Rakvere stages) of the North Estonian confacies belt (here and further the confacies belts according to Jaanusson (1976); Fig.1). It consists of approximately equal quantities of organic (kukersine = kerogene of kukersite), terrigenous (argillaceous) and carbonate components (Kõrts & Einasto1990, Kõrts 1992). In most stages kukersite is mainly represented by several interbeds. Aserian and Lasnamägian rocks contain kukersine only locally in burrow fillings as a very slight admixture. In the Kunda Stage kukersite occurs widely only in the north western Estonia. Its spatial distribution is the widest in the Uhaku and especially in the Kukruse Stage, where it comprises up to 30% of the whole thickness in places. The both stages have a cyclic structure, represented by rhythmical alternation of marl and/or argillaceous limestone, limestone and kukersite. In the coastward areas the cycles are often completed by a pyritic discontinuity surface on top of the kukersite beds. These two stages also have an excellent subdivision: more than 50 kukersite beds and complexes with synchronous borders are registered here (Männil & Bauert 1986). Correctness of the correlations is proved by

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[2] The Lasnamägi and Uhaku stages are in this paper used in their former (Orviku, 1940: Rõõmusoks, 1970) limits.

micropalaeontologic data: more than 60 chitinozoan (species of the group *Cyathochitina campanulaeformis*) zonules and interzonules have been established in this section (Männil 1986). The kukersite beds occur as east-westward prolongated, extremely gently sloping lenses, genetically wedging out to the north. In the other directions they are replaced by limestone (Männil, Bauert & Puura 1986; Saadre & Suuroja 1993). The beds of marl and limestone have similar distribution pattern.

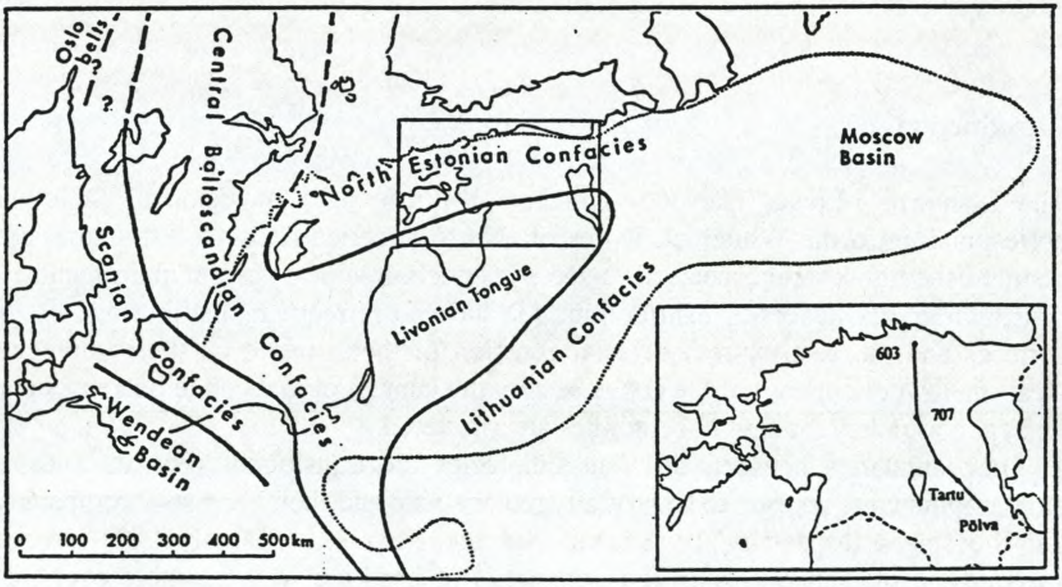

Fig. 1 Location of the facial cross-section with respect to the confacies belts of the Balto-Scandian region (by V. Jaannusson, 1976, Fig. 7).

The discontinuity surfaces (DS`s) in Estonian Ordovician have different types of mineralization: pyritic, phosphatic, limonitic (+ goethitic), hematitic and with glauconitic film; they can also be nonimpregnated (Põlma 1982; Saadre 1992). In this particular section the first three of them and the nonimpregnated DS's are represented (Saadre 1993). The number and the type of impregnation of DS´s appears to depend on the facial position of the section. The pyritic DS´s , being the most common in the Ordovician of Estonia occur only in the upper part of the section under consideration. In the Uhaku and Kukruse stages they occur only in the northern, coastward part of the section close to the outwedging area. The DS´s of the aforesaid type are registered mainly on the upper boundaries of the kukersite beds and complexes, where their maximum number in one core section reaches 10 in the Uhaku and 20 in the Kukruse Stage. The morphology of the surfaces is quite different. Beside the level and slightly undulating DS´s, numerous robustly undulating surfaces with a network of different size cavities (mostly 3-5 cm deep and up to 5 cm wide) can be observed as well. The thickness of the impregnation zone of the DS's is 1-20mm (5mm on the average). In the Lasnamägi Stage in Eastern Estonia the pyritic DS´s are distributed only in the basinward side of the section (cores Laeva, Tartu, Põlva), where their number ranges from 5 to 25. The DS´s are mostly broken, their pieces occur as in more or less rounded pebbles (up to 5x10 mm) which are located horizontally or irregularly.

The phosphatic DS´s are widely distributed in the treated part of the section, missing only in the Aseri Stage. In the other stages their maximum number in one core section exceeds 20. They are represented in the parts of the section where the thickness of the units (beds, members, stages) is stable and the clay content of the rocks is low (compared to the adjacent areas). The phosphatic DS´s are mostly slightly undulating (the difference between the ups and downs < 1 cm), their impregnation is comparatively weak and the thickness of the zone is mostly 1-2mm. Nonimpregnated DS´s are registered in some levels in the Uhaku and Kukruse stages, northward of the distribution area of the phosphatic DS´s.

Limonitic and/or goethitic DS´s are extensively represented in the oolitic limestones of Kunda (Sillaoru and Napa formations) and Aseri stages (up to 20 in both). They are distributed amongst the redcoloured and variegated (greenish gray, yellow and red) rocks of Kunda, Aseri and Lasnamägi stages also. These DS´s are also slightly undulating, the thickness of their impregnation rim reaches 1 cm.

Most of the above described DS´s can be regarded as firmgrounds, except most of the pyritic surfaces which show true evidence to have formed as hardgrounds.

Ooids are widely distributed in the Ordovician sediments of Estonia, especially in the interval from Kunda to Idavere stages. Most of the ooids are ferriferous (goethitic) but in some stages and regions carbonate (in the Porkuni Stage) and phosphatic (francolitic) ooids occur (Põlma, 1982). Along with the ferriferous ooids, or in the same levels in adjacent areas ferruginized skeletal grains (pseudo-ooids by Orviku (1940)) are distributed. In Estonia two most important ooid levels are distinguished, both cropping out in the clint area: at the base of the Kunda Stage - the Sillaoru Formation (occupying also the extremely thin -<10cm - Pada Member from the top of the Volkhov Stage) =the Lower Oolitic Bed, and the Aseri Stage = the Upper Oolitic Bed (Lamansky 1905). In the north-easternmost Estonia the latter includes also the Napa Formation at the top of the Kunda Stage. All these sedimentary bodies can be traced for more than 70 kilometers southward at the presented cross-section. In the deep seated area three more elliptic bodies can be observed in the section: in the Lasnamägi, Kukruse and Idavere (Tatruse Formation) stages. The thickness of the elliptic bodies varies mainly between 0,5-3m (predominantly 0,5- 1 m), content of ooids - 10-30 % (reaching 80 % in patches - Sillaoru Formation in the Ontika section (Mägi 1990)).

It is proved that in the Aseri Stage the real ooids tend to occur mostly in the north -western part of Estonia (Orviku 1940; Sturesson & Bauert 1994). Toward the east and south they are gradually replaced by the ferrugenised skeletal grains predominantly. Compared to the Sillaoru and Asery ooid bodies, the content of the real ooids in the other levels is considerably lower. The distribution of ooids in the elliptic bodies is rhythmically changing. In some levels and areas the structure of the oolites is cyclic. In most cases the ooids are abundant just above the DS`s and decrease gradually upward. Differently from the widely distributed recent sparry elliptic sands and limestones, the above discussed Fe-ooids are scattered in muddy (micrite) matrix. They were plausibly formed in some distance from the deposition area in high energy environment (on the hard- or firmground basinfloor), afterwards transported and buried in the lime mud (from personal discussion with R. Einasto)

Facies cross-section

The cross-section (Fig. 2) situated in the eastern Estonia extends from the clint area of the North Estonian confacies belt to the central belt (Livonian Tongue, Fig.1). It is compiled basing on the assumption that the Lasnamägi Stage, being extremely stable in thickness and lithology, must

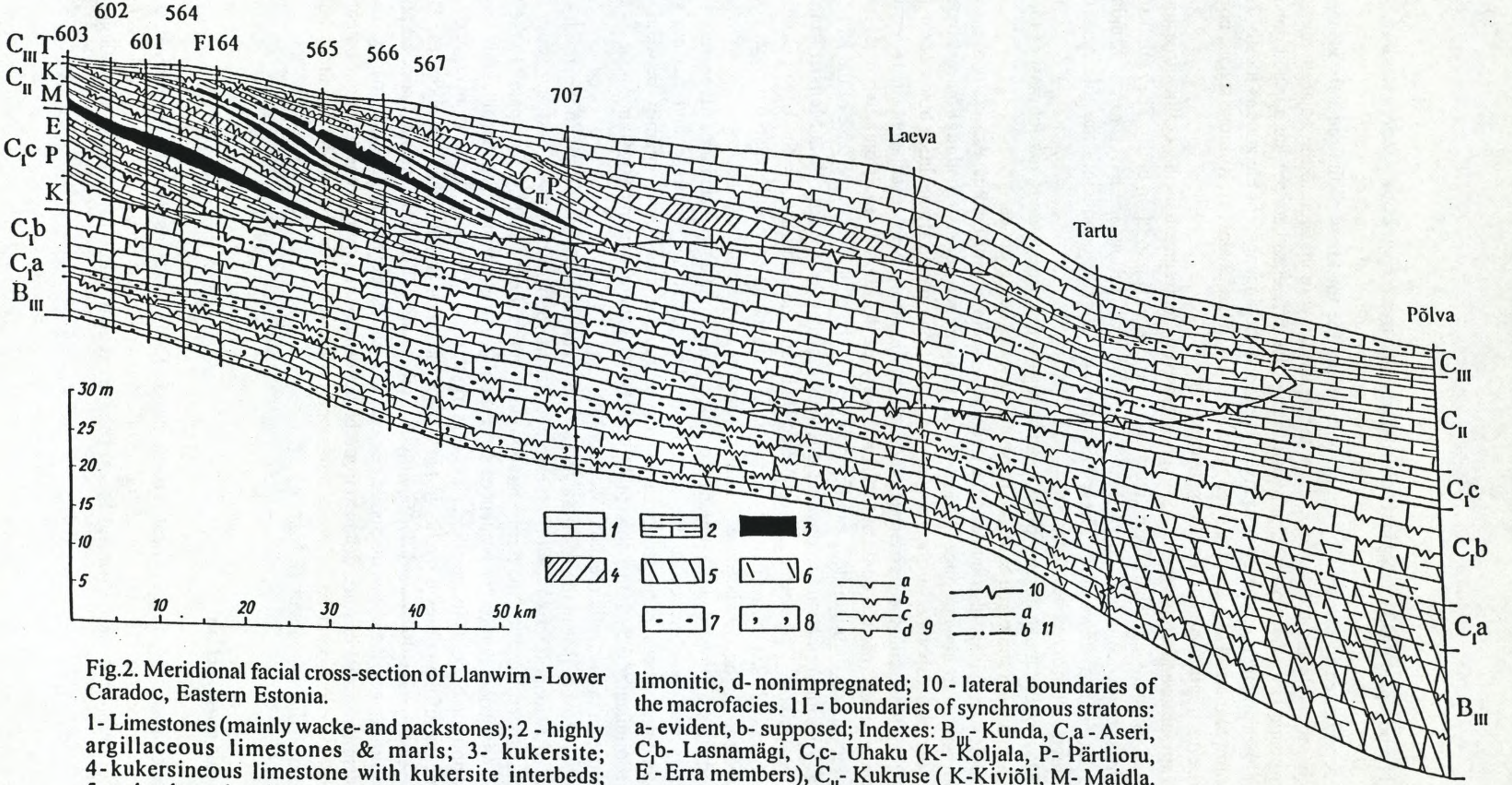

Fig.2. Meridional facial cross-section of Llanwirn - Lower Caradoc, Eastern Estonia.

1 - Limestones (mainly wacke- and packstones); 2 - highly argillaceous limestones & marls; 3 - kukersite; 4 - kukersineous limestone with kukersite interbeds; 5 - red-coloured rocks; 6 - variegated rocks; 7 - Fe-ooids & ferrugenised skeletal grains; 8 - glauconite; 9 - discontinuity surfaces : a- phosphatic, b- pyritic, c- goethitic & limonitic, d- nonimpregnated; 10 - lateral boundaries of the macrofacies. 11 - boundaries of synchronous stratons: a- evident, b- supposed; Indexes: B_{III} - Kunda, C_Ia - Aseri, C_Ib - Lasnamägi, C_Ic - Uhaku (K- Koljala, P- Pärtlioru, E - Erra members), C_{II} - Kukruse (K-Kiviõli, M- Maidla, P- Peetri members), C_{III} Idavere (T- Tatruse Formation) stages.

have been formed in comparatively stable conditions - in the gently sloping or equal in depth part of basin with even seafloor. Therefore, the upper boundary of the stage has been taken on the cross-section a straight line. All the thicknesses of the sedimentary bodies up-and downward are counted from this level. The cross-section is inclined so that the outwedging parts of the sedimentary lenses in the Kukruse Stage would stay horizontal or have a slight basinward bend.

In the section the following areas and corresponding macrofacies can be distinguished (basinward from the costal area):

1.The distribution area of coastward outwedging sedimentary lenses is represented only in the Uhaku and Kukruse stages. The lenses are characterized by extremely regular shape, well exposed cyclicity, the highest clay and kukersine content compared to the basinward areas and by numerous pyritic DS's.

2. The area of stable thicknesses is present in the Kunda (Valgejõe Member), Lasnamägi, Uhaku, Kukruse and Idavere (Tatruse Formation) stages.The macrofacies is represented mainly by pure (insoluble residue < 10 %) or slightly argillaceous limestones (wackestones or packstones) with numerous phosphatic DS's.

3.The ooid distribution area. Real ooids are registered mainly in the coastward area in the Kunda and Aseri stages; basinward they are mostly replaced by the superficial ooids and pseudo-ooids (= ferruginized skeletal grains). Pseudo-ooids predominate also in the Lasnamägi, Kukruse and Idavere stages.

4. The red-colored sediments are distributed in the central confacies belt, in the lower part of the section (Kunda and Aseri stages). On its coastward side a zone is located where the rocks are alternately of greenish gray, yellow (from goethite) and red (from hematite) colour. In the Lasnamägi Stage only the zone of variegated rocks is represented.

Discussion

Constant regression during the Uhaku and the Kukruse time with regard to gradual basinward migration of the sedimentary lenses replaced by a new transgression in the beginning of the Idavere time was first stated by Männil and Bauert (1984). The distribution pattern of the macrofacies in the cross-section (Fig.2) (gradual migration of the Fe-oolites and red-coloured sediments) allows us to surmise that the regression started at the beginning of the Asery time already. This presumption fits well to that stated for the Livonian Tongue of the central confacies belt (Ulst 1982), although a week transgressive pulsation was proposed for the lastmentioned area.

The general view of the cross-section performs the following: the coastward side - regular progradation of the sedimentary lenses, resulting from the gradual accommodation of the space available for sedimentation; in the central part - extremely stable thicknesses of sedimentary bodies; on the basinward side - increasing thicknesses. Based on the aforesaid, we assume that the northward and the central part have acted as a tectonically stable block at least from the Àseri to the Idavere time, its rate of subsidence corresponding to that of sediment accumulation in the long run. Sediment compaction is considered a minor factor and is therefore neglected.

Based on these assumptions, the following relief elements are distinguished there:

Zone A represents periodically emerging nonsedimentation area.

Zones B - D perform the upper ramp of the open shelf (Nestor 1990), embracing the North Estonian confacies belt and the transitional zone. Zone B represents a narrow, extremely gently sloping (nearly horizontal) area, which in the performed section has preserved only in the upper part of the Kukruse Stage (elsewhere it is cut down by the post Ordovician erosion) where the outwedging ends of the lenses were deposited beneath the wave base. Due to the extremely gently sloping seabed relief, the depth of the latter could have been only a few meters. Zone C is the next, much wider and the steepest part of the upper ramp where the sedimentary lenses (which have preserved in the Kukruse and partly in the Uhaku stages) were formed. According to the calculations the depth of deposition at the foot of the lenses could have been about 20 - 30 m. Zone D is the widest and again extremely gently sloping part of the upper ramp.

Zone E performs the upper part of the lower ramp, facially belonging to the central belt already. In this part of the basin the subsidence rate was evidently higher, especially up to the end of the Kunda age (Männil 1966; Ulst 1982) which has resulted in the greater thicknesses (Fig. 2).

Acknowledgments

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AUTOCHTHONOUS AND ALLOCHTHONOUS COMPONENTS IN THE BILLINGENIAN CONODONT ASSEMBLAGES OF THE EAST BALTIC

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Abstract

Comparative analysis of a distribution of settling velocities for conodont elements and quartz sand grains from various types of lithology in three Billingenian sections of the East Baltic have shown that the original composition of assemblages is ususally modified by various reworking processes. Autochthonous conodont assemblages have been recovered only from interlayers of clay in the Päite Member (Upper Billingenian), whereas assemblages from limestones in this unit are usually reworked. Autochthonous component may be also recognized in some mixtured assemblages from the glauconitic sand of the Mäeküa Member. Reworking processes affect strongly a local range of some species and taxonomic composition of assemblages in the Lower Ordovician sections of the East Baltic.

Introduction

Sorted and reworked conodont assemblages are recorded, possibly from all the main types of lithology in a shoreline-to-basin transect. Ususally, a presence of allochthonous components in a conodont assemblage may be recognized on a basis of differences in preservation, color alternation index, as well as in a mixture of taxa, which strongly differ in their stratigraphic range and ecology (Chernyshuk 1989). In the environments with a very low rate of sedimentation, which are typical for the Early Ordovician of the Baltic Basin, the original characters of conodont assemblages sometimes are strongly modified by the various reworking processes (Lindström 1984), but the scale of sorting, transportation and redeposition, as well as their influence on a primary composition of the assemblage is difficult to estimate.

As it was shown by Broadhead *et al.* (1990), settling velosity of conodont elements is one of the important parameters, which is correlated strongly with a hydrodynamic regime, and may be used to define the autochthonous, allochthonous, mixtured or sorted conodont assemblages in samples. Settling velocity of various types of conodont elements is correlated with their shape and size (Fig. 1). Therefore, comparative analysis of distributional patterns for conodonts and sand grains, which usually are present in the sediment, may give an important information on the relative abundance of the allochthonous and autochthonous components in conodont assemblages. It makes it possible to estimate the influence of the selective sorting on the relative abundance of conodont elements with different morphology and to recognize conodont assemblages which are entirely transported or reworked.

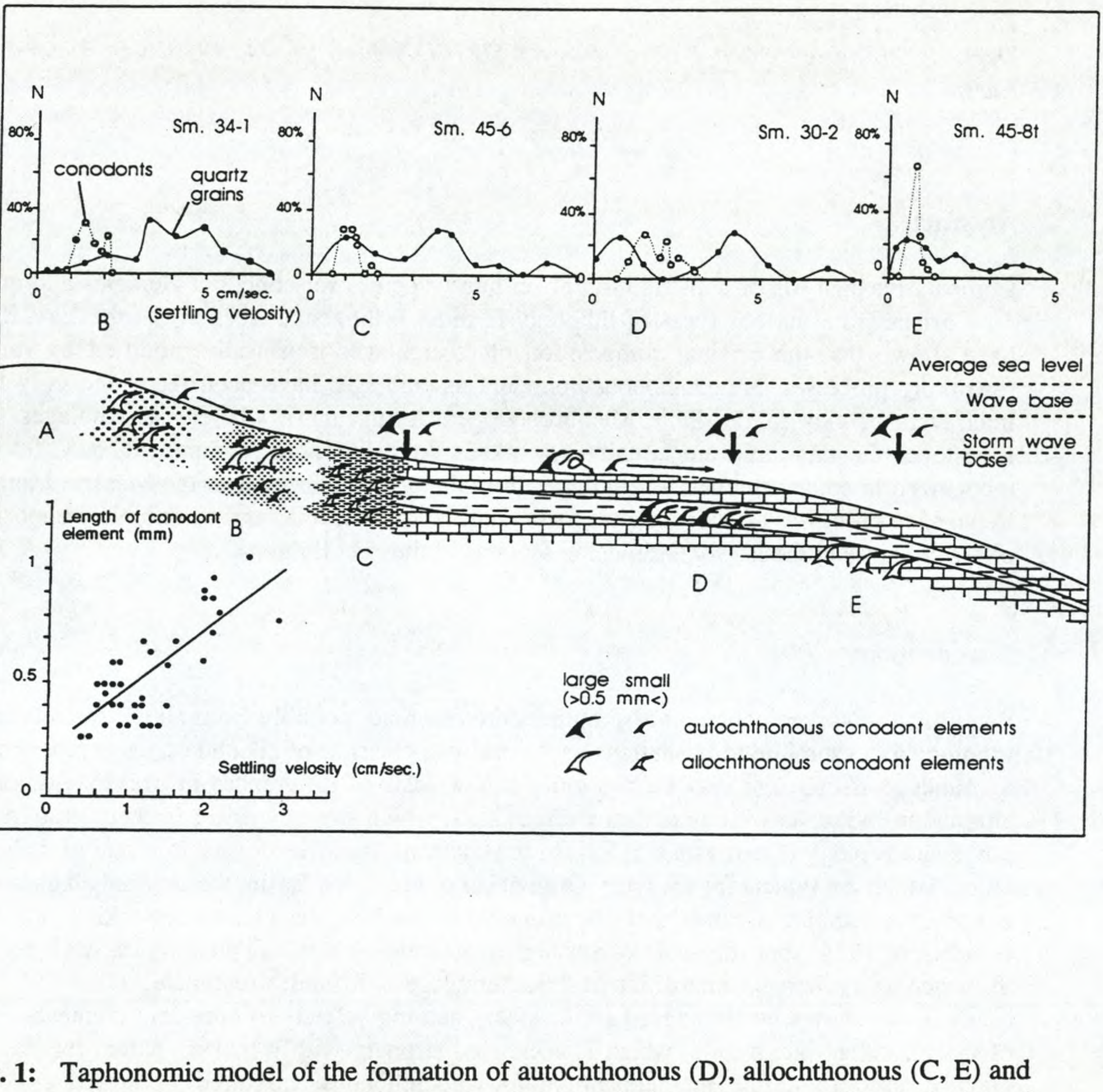

Fig. 1: Taphonomic model of the formation of autochthonous (D), allochthonous (C, E) and mixtured conodont assemblages showing correlation between settling velocity of conodonts and their size, various types of a distribution of settling velocities of conodont elements and quartz grains (B-E) and their relation to various types of hydrodynamic. Nearshore depositional environments (A) absent in studied sections of Billingenian strata.

Material and Method

This study is based on about 50 samples from three sections of Billingenian strata situated in the eastern part of Baltic-Ladoga Clint, northwest Russia (Fig. 2). Two localities are situated in the phosphorite quarry west of Kingisepp and south of St. Petersburg-Tallinn highway. A third locality is situated near a waterfall about 0,5 km above the stream to the mouth of Sablinka river and a section on the right side of Tosna river studied and described by Bergström (1988).

Billingenian strata in these localities are represented by a condensed sequence of coarse to fine-grained quartz glaucanitic sand, calcareous sandstone and marlstone (Mäeküa Member) in the base and glauconitic limestone with thin intercalations of clay (Päite Member) to the top (Fig. 2), which were formed during the early Arenig transgression mainly within the interval of *P. elegans* and *O. evae* zones. Differences in the the lithology from the base to the top and numerous discontinuity surfaces suggest on varying hydrodynamic conditions and depositional environments.

The main purpose of this study is a comparative analysis of the distribution of settling velocities for conodont elements and associated quartz sand grains. Settling velocities of quartz grains are calculated according to the Onishchenko's (Onishchenko & Jampolskij 1990) formula. A distribution and regression line for settling velocities of conodonts were deduced on a basis of experimental studies (Fig. 1).

Results

There are large variations in characters of a settling velocity distribution for conodonts and sand grains in samples, which suggest on their formation under the different hydrodynamic regimes. High and low water energy zones and an erosion zone, as well as some storm generated deposits in the predominantly low energy depositional environments may be recognized.

Distributional patterns of settling velocities of conodonts and quartz sand grains in studied samples from Billingenian may be subdivided into three main types (Fig. 2, B-E).

(1) Curves showing distribution of settling velocities for quartz grains and conodonts do not coincide and settling velocities of conodonts are usually much lower. These differences are interpreted here as the evidence for autochthonous character of conodont assemblage. However, a small additional peak is sometimes present on the conodont curve. This suggests that a small number of reworked conodonts are present in some samples. This distributional pattern occurs in samples from the intercalations of a clay within a limestone of Pite Member (Fig. 1E).

(2) Curves showing distribution of settling velocities for quartz grains and conodonts are more or less coincide, which suggest that the conodont assemblage is strongly affected by the reworking processes and should be considered as essentially allochthonous.

Settling velosities in the reworked conodont assemblages usually have a tendency to be close to normal distribution. In sorted conodont assemblage elements both with low or high settling velocities may be removed. Therefore sorting index may be used to estimate significance of the selective sorting of conodont elements in samples.

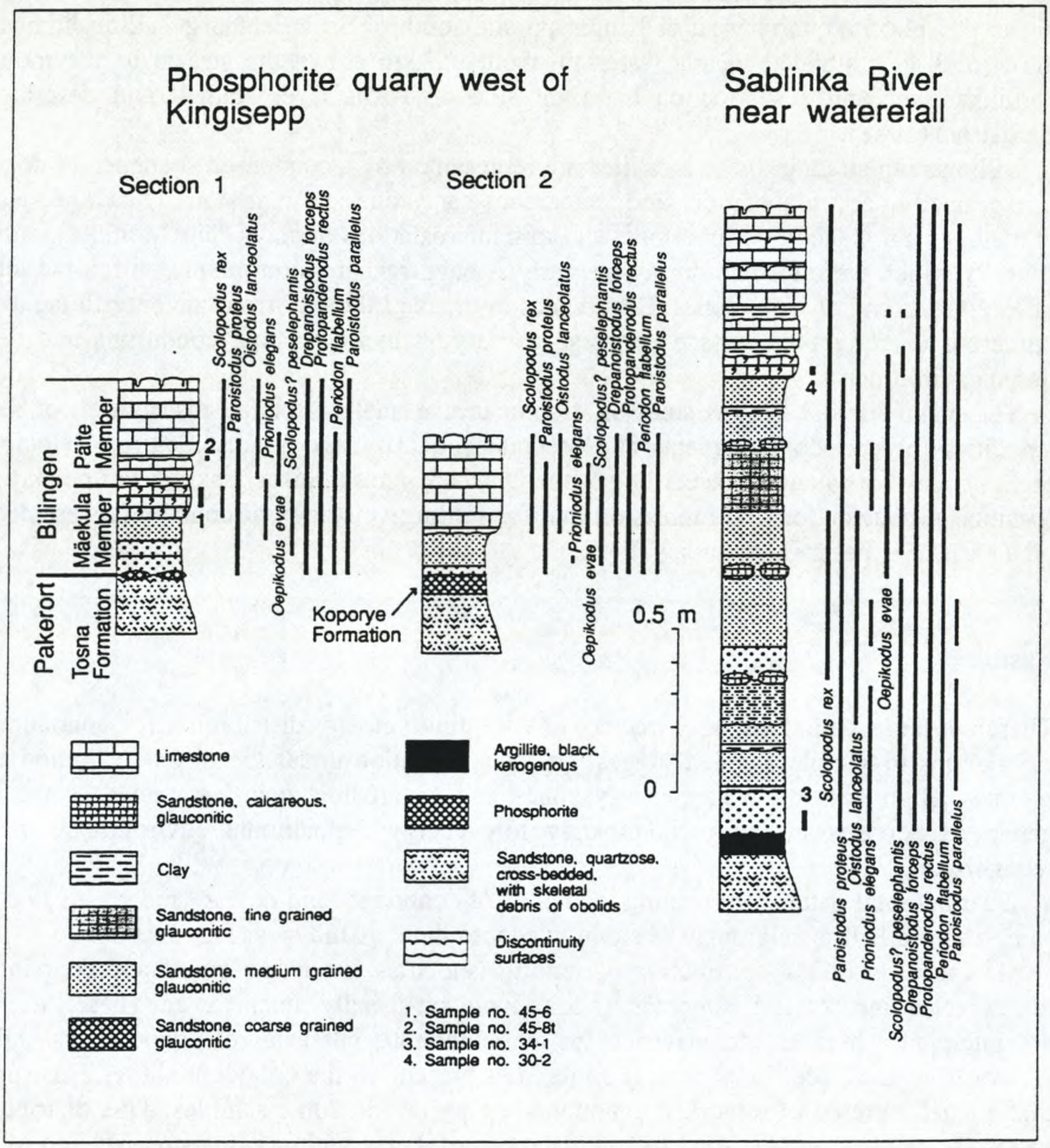

Fig. 2: Faunal log from three selected sections of Billingenian strata at the eastern part of the Baltic - Ladoga Clint showing a distribution of conodont species and position of samples used for illustration of differences in the distribution of settling velocities of conodont elements and quartz grains in the different lithology.

Only conodont assemblages from interlayers of clay in the upper part of Billingenian (Päite Member) are mainly autochthonous, whereas conodont assemblages from the limestone, comprising a main part of the unit, are usually reworked. Settling velocity distribution of conodonts in the latter is characterized by the low value of sorting index and high values of settling velocity. Comparison with assemblages from clay suggest that reworking and sorting of conodonts in the limestones of Päite Member may take place *in situ*, as a result of storm processes, and therefore this pattern (Fig. 1E) may characterize proximal tempestites.

Another distributional pattern of mainly allochthonous conodont assemblages is characterized by the high value of conodont sorting index and low values of settling velocity. It occures in a fine-grained sand and calcareous sandstone of Mäekula Member and may be typical for the accumulation zone (Fig. 1C).

A third distributional pattern of allochtonous and reworked conodont assemblages is characterized by the high value of sorting index and high values of settling velocity (Fig. 1A). It has never been recorded from the Billingenian, but sometimes occurs in cross-bedded sands of the Pakerortian Tosna Formation, which formed under the high energy, near-shore environment. Conodont elements in this type of assemblages are presumably entirely transported and redeposited.

(3) Curves showing distribution of settling velocities of conodonts have a well-defined bimodal character with one peak more or less coinciding with a distributional pattern of sand. This bimodal curve may be interpreted as an evidence for mixed character of conodonts in the assemblage with one peak corresponding to the atuochthonous component and another one produced by transported conodont elements and placed in the sediment together with clast grains. This distributional pattern occurs in samples of a medium grained quartzose glauconitic sand in the base of Mäekula Member (Fig 1B).

Discussion

Analysis of the distributional patterns of settling velocities in conodont assemblages from the Billingenian strata suggest that they are more or less modified. As a result of the original relationship between taxa have changed and distortion of numerical relations between elements belonging to the same apparatus is highly probable. For instance, a gap in the ranges of *Scolopodus rex* and *Oistodus lanceolatus* in the Sablinka section (Fig. 2) more or less coincides with a layer of fine-grained glauconitic sand. This layer contains reworked assemblages of conodonts with low values of settling velocities, which suggest that conodont elements of the relatively large species were completely removed from the assemblage.

Differences in a local stratigraphic range and relative abundance of ramiform elements of *Prioniodus elegans* and *Oepikodus evae* in Sablinka and Kingisepp quarry sections (Fig. 2), also are mainly a result of the taphonomic control. Ramiform elements are relatively rare in samples with high sorting index and settling velocities, which is typical for the interval of *Prioniodus elegans* Zone in two studied sections of Kingisepp phosphorite quarry.

Conodont assemblages in Billingenian strata of the East Baltic are usually strongly modified by various reworking processes, and it is likely that the same processes may change an original character of conodont assemblages in condensed Lower Ordovician carbonate sequences of the other parts of the Baltic Basin (Lindström 1984). Conodont assemblages from clays retain the main characters of the original conodont assemblages and proportions between taxa and different elements belonging to the same conodont apparatuses are not distorted. Similar data obtained from samples of sand and carbonate rocks may be used for reconstructions of conodont biofacies and apparatuses only when the autochthonous nature of the assemblage is confirmed by other observations.

Comparative analysis of the distribution of settling velocities for conodonts and sand grains also may be used in sedimentological studies as one of the criteria of proximal tempestites, especially when the primary structure of the rock has been changed by the subsequent bioturbation.

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LITHOLOGY OF ORDOVICIAN SEDIMENTS, KULDIGA OIL FIELD, LATVIA

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Introduction

The Kuldiga oil field is situated in western Latvia, at the eastern edge of the Baltic syneclise, within the Liepaja depression. The field is associated with the anticline of the same name forming the Kuldiga rampart together with several other highs. The main oil accumulation is associated with Middle Cambrian sandstones. In Ordovician sediments, 3 oil-bearing horizons have been observed represented by carbonate rocks having low porosity.

The Ordovician sediments of the Kuldiga field belong to the "onionskin" or "layer cake" class. They are represented by intercalation of multi-coloured clayey-carbonate rocks and are 152-180 m thick.

The deposition of Ordovician sediments was preceded by a prolonged denudation during the late Cambrian. The Ordovician sediments lie on the washed out surface of the Cambrian ones, the latter being uneven, sometimes weathered, with basal conglomerates in some sections. The thickness of the conglomerates is up to 10 cm. The pebbles are represented by reddish-brown sandstones cemented by dolomite, dolomites, quartzose sandstones with oboloid fauna and dark brown kerogen-containing argillites.

At the boundary of Ordovician and Silurian, a hiatus has been observed which, probably, was not long as, within the Kuldiga High, both gradual transition and hiatus surfaces in top of the Ordovician have been observed. Those surfaces are uneven with pockets, and are covered by a limonite crust. The rocks under the hiatus surface display spotty ferruginization down to 7 cm.

The Ordovician sediments are represented by the formations belonging to 2 cycles of sedimentation (sedimentary complexes) - the lower Ordovician cycle and the upper Ordovician cycle. Each of the sedimentary complexes consists of 3 units; the lower unit is clayey, the middle unit comprises red carbonates, and the upper unit which is grey carbonates. In general, clayey and fine-grained carbonate rocks are predominant in the section. The sedimentary complexes are divided by the pyritized uneven surface of the underwater break. The thickness of the lower complex is 95-125 m and the upper complex is 56-68 m thick (Fig. 1; Fig.2).

The Lower complex

Unit a - thin-bedded grey claystones with graptolites, enriched by algal organic matter within layers and red claystones with hard grounds and rich trilobite and cephalopod faunas. At the base, there are greensands - dark green horizontally bedded chemogenous rocks, consisting of glauconite pellets (medium and large psammitic size) cemented by clayey-carbonate, ferrous or carbonate material.

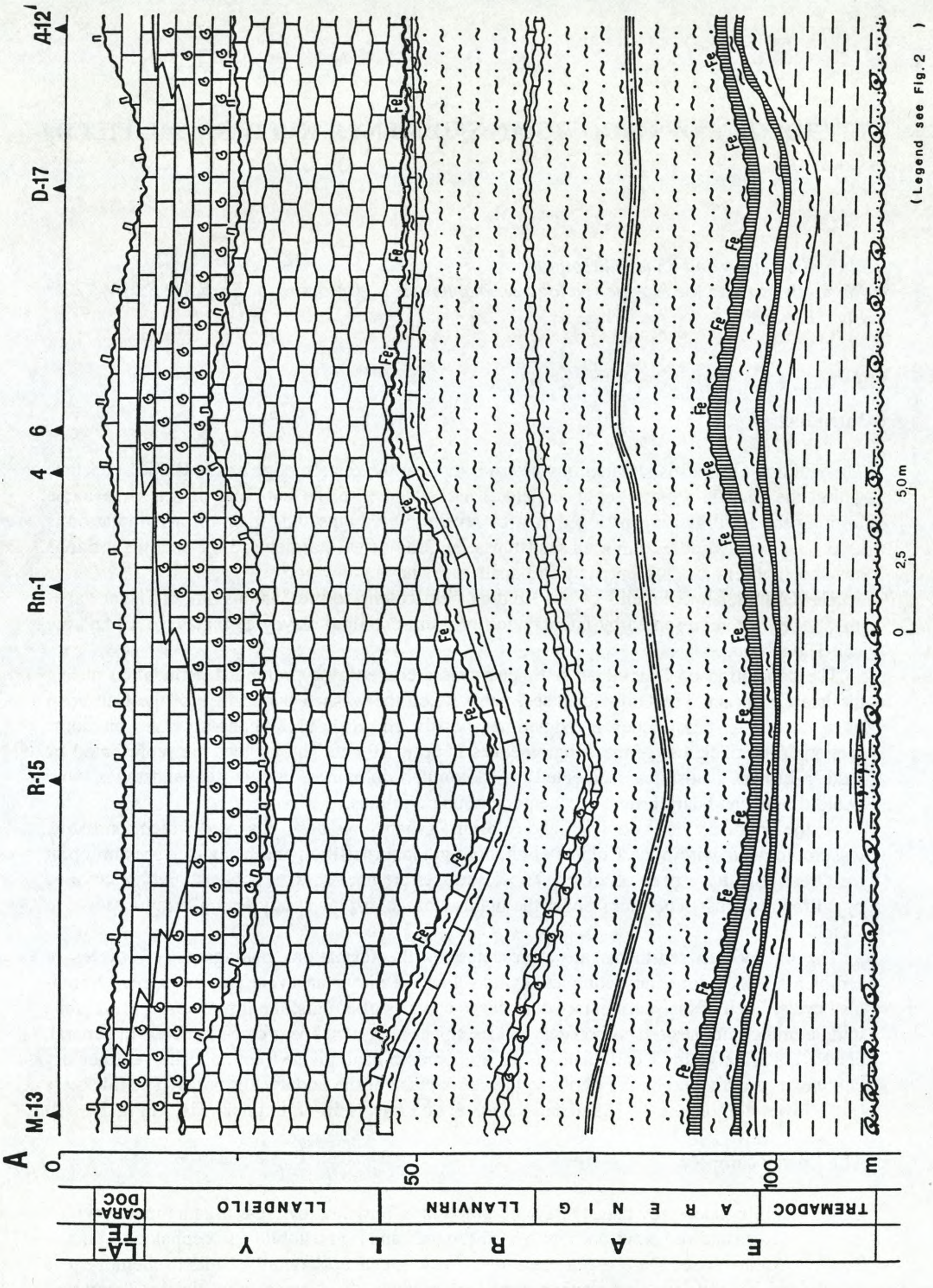

Fig. 1. Profile A-A'. Kuldiga oil field. Section of Lower Ordovician complex (see legend on Fig. 2)

Unit b - red lenticular and nodular calcisiltites overlying with washout the claystones. The calcisiltites consist of a mixture of thin - crushed organogenous detritus, clay and Fe-oxides. On the bedding planes shrinkage fractures, numerous vertical burrows belonging to Glossifungites facies typical for sublittoral environments are common. The calcisiltites are a genetic type of upper tidal zone rocks of arid areas; in the Ordovician sequence of the area under dissusion they represent the shallowest water facies.

In the lower part of red calcisiltites, there is a thin (0.1-0.65 m) layer of light greenish-grey siltstone with cross-bedding and washout signs at the top and bottom. Within the Kuldiga High, the siltstone is, as a rule, oil-saturated.

Unit c - grey micritic limestones rich in organogenous detritus, 37-50 m thick, overlying red calcisiltites; the transition is gradual. The thickness of the transition zone is 0.5-3.0 m. Lens-like interbedding of red calcisiltites and grey micrite limestones occurs. The red limestone is characterized by lensoid-wavy structure; in the lower part, extensive bioturbation, and, possibly, subaerial lithification of the formation with intraclasts occured.

The largest accumulation of organogenous detritus is associated with Kukruse interval of the section, 6-12 m thick. The organogenous detrital limestone prevails. Numerous horizontal burrowsare present in the limestone. The burrows (horizontal curving tubes), diameter 1-3 mm, can be classified as belonging to Cruziana facies and are characteristic of littoral sedimentation conditions.

The organogenous detritus is diverse and is represented by trilobites, bryozoans, brachiopods, corals and cephalopods. The top part of the grey limestone contains ash material represented both by montmorillonitized dust tuff and seldom by tuff-siltstone and as montmorillonite mixture in limestone. The montmorillonite forms glauconite-like pellets in the limestone as well as concentric nodules.

The whole thickness of the grey limestones is to a greater or lesser degree oil-saturated. The oil migrates along sedimentary and diagenetic fractures, fills pores, caverns and cavities in the faunal detritus.

The upper complex

The upper complex overlies the wash-out surface of the underlying grey limestones of the lower complex.

Unit a - Grey claystones occur at the base with a thickness of 13-17 m; it corresponds to the maximum water-level period. The structure is condensed and claystones display numerous sedimentation pauses. This unit is rhytmic and represented by a lower bed consisting of grey calcareous claystones and an upper bed of black bituminous shales. Generally, the sediments are thin-bedded. The grey claystones contain shell detritus of varying size while the black one is characterized by a fauna of pyritized graptolites and brachiopods. In the grey claystones of the upper rhythm, a coquina layer, up to 1 m thick and with uneven top and bottom is present. It has a typical appearance of a "stormy" layer and contains identically oriented and disarticulated shells of brachiopods and ostracodes. In some sections this kind of limestone is oil-bearing.

In the top part of the claystone unit, a regionally uniform sedimentary break has been observed.

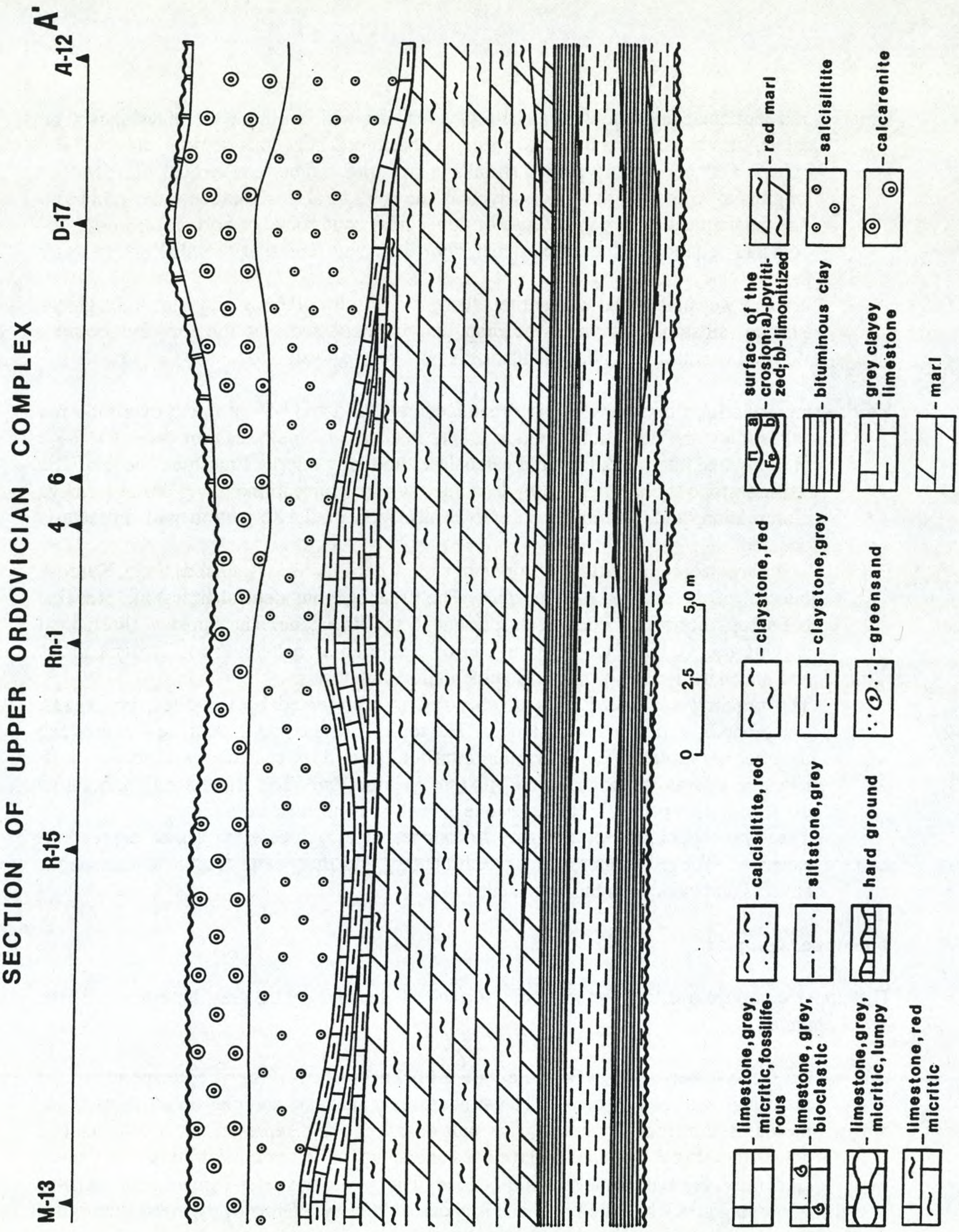

Fig. 2. Profile A-A'. Kuldiga oil field. Section of Upper Ordovician complex.

Unit b - The overlying red marls (18-24 m thick) have the same composition and structure as the underlying claystones. The transition zone, 1.5 - 3 m thick, consists of horizontally-bedded calcareous claystones and marls with linear interbeds of bituminous shales. Upwards in the sequence the content of calcite in the rocks increases, the claystones disappear and lumpy cryptocrystalline grey, higher-red, limestones appear. Upwards the latter are passing into red marls. The marls have a uniform appearance with non-distinct horizontal bedding and with a gradual enrichment of clayey material and disperse Fe-oxides upwards in the sequence. Within the wholesuccession long thin burrows, small brachiopods and crinoid detritus are present.

Unit c - Grey detrital limestones (22-29 m thick) ending the Ordovician section, have a regressive structure and consist of calcarenites, calcisiltites and limestones with numerous terrigenous carbonate and silicate materials. The transition from limestones to the underlying red marls is gradual and at the top a wavy surface of the post-Ordovician washout is present. The grey limestone unit is characterized by the Porkuni oolite structure, the sandbank retreating towards the sea. The base of the sandbank consists of a shallow-water offshore siltstones and clayey limestones, 5-8 m thick and wavy-bedded with rare burrows.

Those limestones are conformly overlain by wavy and lenticular bedded calcisiltites, with oolites, - the formations of the distal part of the sandbank. The thickness of the calcisiltites is 10-18 m. The calcisiltites contain interbeds of skeleton tempestites - silty limestones with a wavy top and even bottom up to 10-12 m thick consisting of dissected, uniformly oriented brachiopod valves.

Thin claystone interbeds, thin burrows filled by pyrite, graptolite and orthoceratite fauna are also typical.

The deposits of the sandbank itself (4-12 m thick) overlie with washout the calcisiltites. These deposits are represented by medium - and coarse-grained oolitic calcarenites with multi-stage cross-bedding and regressive structure. The upper part of calcarenite section displays clear signs of the sediments being exposed to non-marine environment.

The interbeds containing clayey material admixtures occur displaying wide pinching out shrinkage fractures; in thin sections from that part of the section, the signs of freshwater dissolution have been observed.

In the eastern sections where the sandbank section has been completely preserved from the post-Ordovician erosion, the sandbank dying surface has been observed, represented by fine-grained limestones with abundant burrows, containing numerous accumulations of algal remains on wavy bedding planes. The limestones are overlain by thin-bedded calcareous-dolomitic claystones characteristic of post-sandbank mud.

Thus during the Ordovician, the area of the Kuldiga Field was a part of a vast carbonate shelf. During the periods when the sea level was high, the distal mud zone was situated there and during the maximum sea regression the tidal flat environs prevailed. At other periods, the field area was occupied by a shallow-water shelf environs, within which a large oolitic sandbank was formed and later died towards the end of the Ordovician history known to us.

Lithification and calcite diagenesis

The oolitic calcarenites of the sandbank at the Kuldiga area display, as a rule, spotted oil-saturation. The distribution of oil in calcarenites and the intensity of oil saturation are directly related to the intensity of post-sedimentation processes.

The Ordovician deposits of the Kuldiga Field are at an early stage of post-sedimentation alteration (reflecting power in less than 0.5%, clay density - 2.04-2.31 g/cm^3). Secondary alterations are displayed, predominantly, by new calcite formations. They are widespread only in grain-supported rocks, -organogenous, -detrital limestone and calcarenites. In thin sections, 3 successive generations of calcite have been observed:

I-Orthomicrite: II-Orthomicrosparite; III-Orthosparite.

Orthomicrite is represented by sedimentation calcite and calcite formed due to the aggradation of the sedimentation calcite; the grain size is 0.001-0.005 mm. Orthomicrosparite is a fine crystalline granular calcite growing on pore walls and filling empty pore space. The size of crystal is 0.005-0.05 mm, their shape is not perfect. Orthosparite is represented by crystal druses of perfectly shaped calcite on the surface of detritus (size 0.005-0.05 mm) and calcite consisting of big crystals joining together separate parts of the rock, their size being up to 2.0 mm.

Orthomicrite and Orthomicrosparite are widespread. Orthosparite occurence in organogenous - detrital limestone is not certain. In calcarenites, the content of 2 first generations gradually diminishes upwards the section; Orthosparite is dominant at the top part, which has been under subaerial conditions. Probably, the latter metasomatically fills the caverns formed due to freshwater leaching of the calcites belonging to 2 earlier generations.

The studies have demonstrated both quantitative and qualitative dependence of authigenic calcites on structural and textural features of the rocks, i.e. on facial conditions. Such a dependence permits us to consider all 3 calcite generations to be sedimentation - diagenetic minerals and to date the end of the Ordovician sediment lithification at the Kuldiga Field as early diagenesis. The latter lithification factors have not, practically, influenced the composition, texture and structure of those sediments.