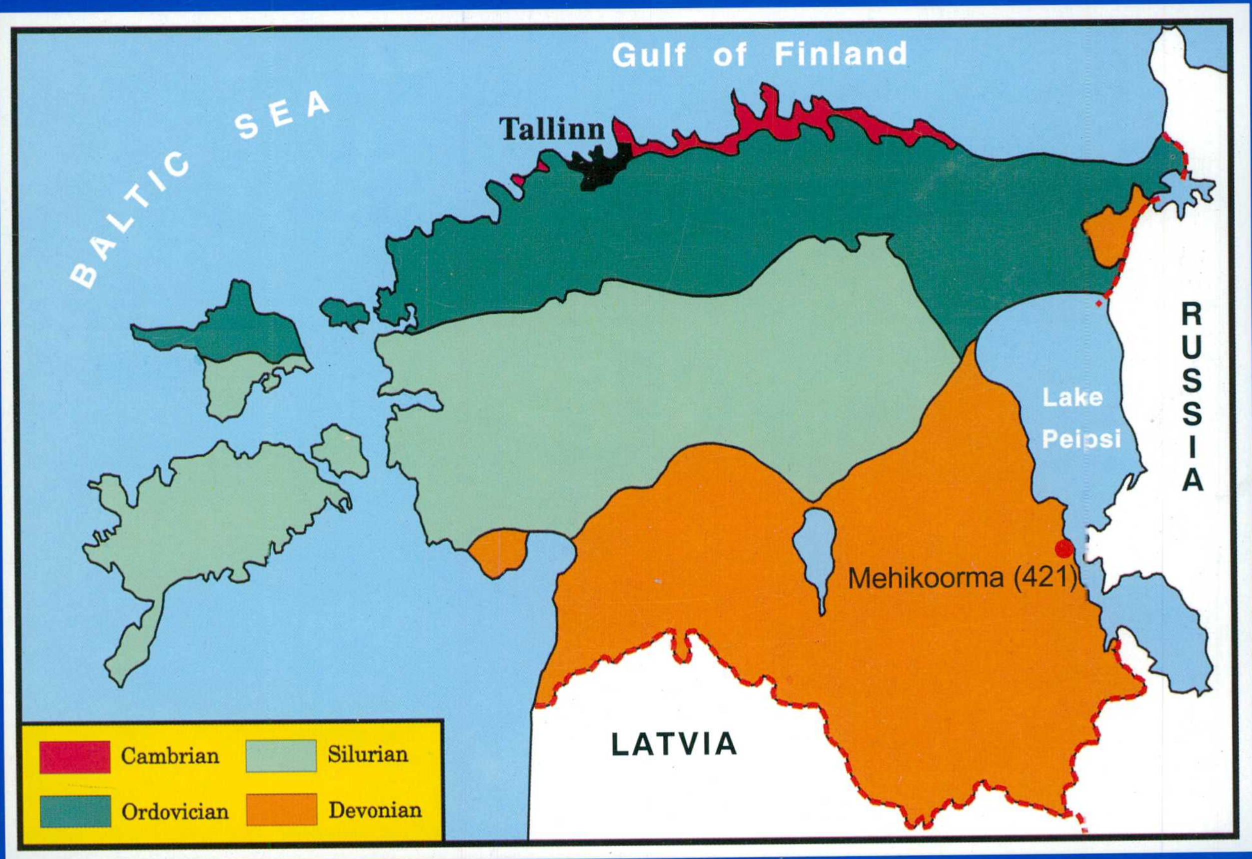
EESTI GEOLOOGIAKESKUS
GEOLOGICAL SURVEY OF ESTONIA

ESTONIAN GEOLOGICAL SECTIONS

BULLETIN 6

MEHIKOORMA (421) DRILL CORE



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PREFACE

The present issue of the journal *Estonian Geological Sections* concentrates on the Mehikoorma (421) core. The Mehikoorma drill hole on the coast of Lake Peipsi (Fig. 1) was made in the course of complex geological-hydrogeological mapping of the southeastern part of Estonia (Kajak *et al.* 1974). The core is housed in the depository of the Geological Survey of Estonia (GSE) in the settlement of Tuula in North Estonia. The source material for the present study is available in unpublished reports (Kajak *et al.* 1974; Põldvere *et al.* 1999) stored in the Depository of Manuscript Reports of the GSE, Kadaka tee 82, Tallinn. The results of earlier micropalaeontological, mineralogical and chemical investigations are used in this work together with recently obtained data.

INTRODUCTION

The Mehikoorma (421) drill hole (58° 14' 26" N, 27° 27' 09" E) penetrates the upper 34.3 m of the Palaeoproterozoic crystalline basement, Ediacaran (earlier Upper Vendian; 57.2 m), Cambrian (85.6 m), Ordovician (125.2 m) and Devonian (206.7 m) sedimentary rocks, and 39.5 m thick loose Quaternary deposits (Fig. 2). Specialists from different fields were involved in the study of the core and their invaluable assistance at various stages of the work is gratefully acknowledged.

Mati Niin (GSE) contributed the macrolithological description of the crystalline basement. Kaisa Mens from the Institute of Geology at Tallinn University of Technology (IGTUT) described the Vendian and Cambrian sediments. Anne Põldvere (GSE) wrote an overview of the Ordovician and Quaternary parts using an earlier report of geological mapping (Kajak *et al.* 1974). Anne Kleesment (IGTUT) provided the lithology of the Devonian strata (description, plus 68 grain-size, 50 mineralogical, 13 chemical and 42 X-ray diffractometry (XRD) analyses). The characterization of the core was improved using the results of laboratory studies.

To improve the stratigraphic subdivision of the Mehikoorma (421) section, Cambrian, Ordovician and Devonian sediments were additionally sampled for microfossils. All of the earlier collected samples (Kajak *et al.* 1974) were used, but 33 ostracod samples from an old collection are still under study. Cambrian acritarchs (2 samples) were identified by Ivo Paalits at the Institute of Geology, University of Tartu (IGUT), Ordovician chitinozoans (72 samples) by Jaak Nõlvak, conodonts (162 samples) by Peep Männik and Viive Viira, and macrofossils (100 samples) by Linda Hints (all from the IGTUT). Acanthodians in Devonian samples (45) were identified by Juozas Valiukevičius (Institute of Geology and Geography of Lithuania), other fossils and few acanthodian remains by Elga Mark-Kurik (IGTUT); a lingulate fragment was identified by Leonid Popov (National Museum and Galleries of Wales, Cardiff).

Tõnu Martma (IGTUT) contributed carbon isotope ($\delta^{13}C$) data of the Ordovician rocks based on the analysis of 112 whole-rock samples.

Alla Shogenova (IGTUT) provided results of wet silicate chemical (133 samples), X-ray fluorescence (XRF) spectrometry (133) analyses and measurements of physical properties (104) of the Cambrian, Ordovician and Devonian sediments. The XRD and XRF analyses of 10 Ordovician volcanic interbeds were made by Toivo Kallaste (IGTUT) and Kiira Orlova (GSE). The contents of CaO, MgO and insoluble residue in 36 Ordovician samples, and grain-size distribution and mineralogy in 5 samples were taken from Kajak *et al.* (1974).

The results of 39 grain-size, 15 mineralogical and 30 XRD analyses of Cambrian sediments obtained during complex geological-hydrogeological mapping of the southeastern part of Estonia were used (Kajak *et al.* 1974). Additionally, Kalle Kirsimäe (IGUT) performed XRD analyses of 2 samples and Terje Kespre (IGUT) XRF and XRD analyses of 7 samples.

Forty-three thin sections were made from the Cambrian, Ordovician and Devonian samples collected by Alla Shogenova. The thin sections were described by Kaisa Mens, Janika Lääts (under the guidance of Asta Oraspõld) and Anne Kleesment (all from the IGTUT). Palaeoproterozoic thin sections (7) were provided by Mati Niin (GSE).

Photos of the core were taken by Gennadi Baranov (IGTUT) and Anne Põldvere. Ene Pärn, Ranek Rohtla (both from the GSE) and Elar Põldvere (Institute of Geography, University of Tartu) provided various technical assistance.

Useful comments by Juho Kirs, Jüri Plado (both from the IGUT), Jaak Nõlvak, Asta Oraspõld, Dimitri Kaljo (all from the IGTUT) and Jaan Kivisilla (GSE) were of great help in finalizing the report.

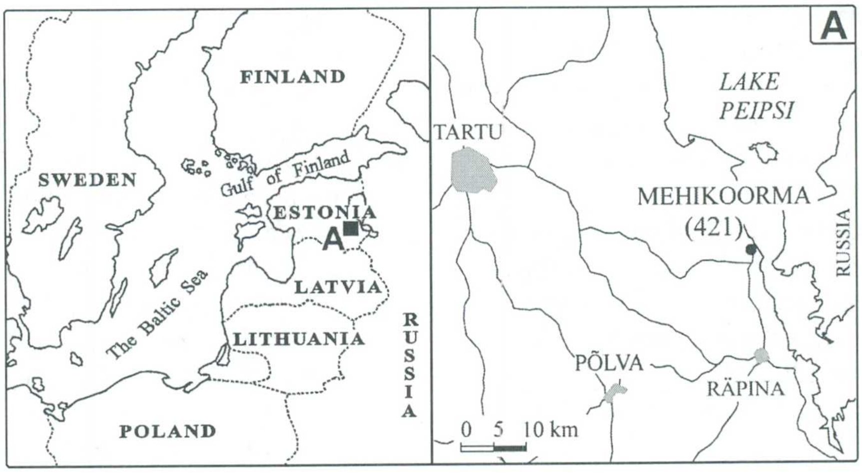

Fig. 1. Location of the Mehikoorma (421) drill hole.

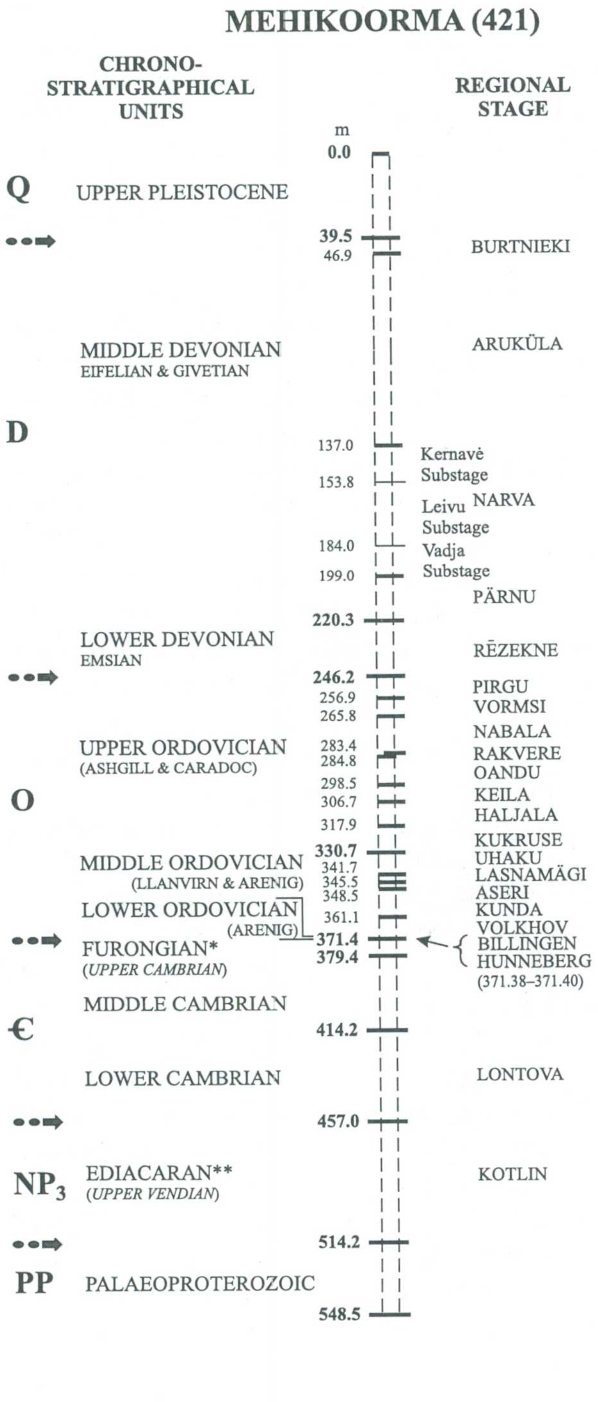

Fig. 2. Generalized stratigraphy of the Mehikoorma (421) core. PP – Palaeoproterozoic; NP_3 – Neoproterozoic III; Є – Cambrian; O – Ordovician; D – Devonian; Q – Quaternary.

* The Furongian Series was ratified in 2003 by the International Commission on Stratigraphy (ICS).

**The Neoproterozoic III System was formally replaced by the Ediacaran System in February 2004, according to a decision of the ICS.

CORE DESCRIPTION AND TERMINOLOGY

The description of the Mehikoorma (421) core is presented in the form of a table (Appendix 1) including the main lithological features of the rock. The material studied comprises 107 grain-size samples, 65 samples for mineral composition measurements, 198 chemical (149 XRF, and 49 CaO, MgO and insoluble residue analyses) and 91 XRD analyses, 104 physical properties analyses and 50 thin sections. For age specification acritarchs (3 samples), chitinozoans (72), conodonts (162), ostracods (2), fishes and invertebrates (45), miospores (3) and macrofossils (100) were used in the Cambrian, Ordovician and Devonian. Additionally, 112 Ordovician carbon isotope ($\delta^{13}C$) samples were analysed.

The degree of dolomitization of carbonate rocks was determined during field work using 3% hydrochloric acid, whereas the content of clay was estimated visually. The rocks were referred to as slightly argillaceous (insoluble residue 10–15%), medium argillaceous (15–20%) and highly argillaceous (20–25%) (Oraspõld 1975). Different contents of calcite in marlstones are denoted by terms "calcareous" ($CaCO_3$ < 25%) or "calcitic" (> 25%).

The descriptions of the textures of carbonate rocks are based on the traditional Estonian classification by Vingisaar *et al.* (1965), Loog & Oraspõld (1982) and Nestor (1990), where the relative amounts of clastic and micritic components are crucial to identification of the textures. The content of carbonaceous clasts (including bioclasts) is given in most cases in per cent.

The particles with the diameter > 0.05 mm are described as grains. Skeletal remnants of organisms or their fragments (bioclasts) are mainly < 1 mm in diameter. The size of chemogenic or biochemogenic ooliths is usually < 1 mm, while the size of carbonate intraclasts is > 1 mm. For the major part of the core, the amount of grains was determined with the magnifying glass on the slabbed surfaces of the core. The micritic component consists of particles < 0.05 mm in diameter. The terms used for textures are explained in Appendix 1. Depending upon the degree of recrystallization, several transitional textures can be observed (secondary textures occurring as patches or spots). In case of mixed texture, the word marking the dominant component is given last, while those marking less important components are placed before the basic word. The same principles were followed in descriptive terms for other characteristics of the rock as well.

The textures recorded are illustrated in photographs of thin sections and details of the Mehikoorma (421) core in Appendixes 2, 3 and 4 (on the CD-ROM).

Several sedimentary structures are described in the style used in the previous issues of the bulletin (see Põldvere 2001, 2003). The relationships between different parts of rock are given in Appendix 1. In some cases the term "lamina" (maximum thickness 1 cm; Boggs 1995) is used for a very thin layer of sedimentary rock, which is of different mineralization or lithology, with uniform thickness and is visually separable from the host rock or other layers. The variation of these structures in the Mehikoorma (421) core is illustrated in Appendixes 3 and 4 on the CD-ROM.

Classification of sandstones is based on the 5-fractional classification of Pettijohn *et al.* (1987), which defines 0.05 mm as the finest diameter for sand particles. Fractions and terms for clay, silt and sand are described in Appendix 1. The term "terrigenous" is essentially synonymous with "noncarbonate" (e.g. *terrigenous sand* vs *carbonate sand*) and is applied to sediments originating from the land area and transported mechanically to the basin of deposition (Scholle 1978).

A peculiar feature of sandstones of the Narva and Aruküla stages is the occurrence of strongly dolomite cemented sandstone globules (diameter 2–3 cm) in up to 10 cm thick irregularly cemented interbeds.

The names of igneous rocks are based on their genesis, and chemical and mineralogical composition.

For parts with insufficient core yield natural gamma ray log was used to specify the lithology of rocks and the extent of lithostratigraphical units. The exact boundaries of lithostratigraphical units were determined by discontinuity surfaces and specific beds (K-bentonites) in the section, followed over a wide area. For the first time gamma log has been included in the bulletin (Appendix 5 on the CD-ROM).

GENERAL GEOLOGICAL SETTING AND STRATIGRAPHY

The bedrock succession in the Mehikoorma (421) section is divided into four general parts: the Palaeoproterozoic crystalline basement, the Ediacaran–Cambrian terrigenous sediments, and the Ordovician carbonate strata and the Devonian, predominantly terrigenous rocks. The Devonian sediments are overlain by the Quaternary cover (Fig. 2; Appendix 1).

The Mehikoorma (421) borehole is located in the granulite facies area (Puura *et al.* 1997). It penetrates 34.3 m into greyish-red fine- to medium-grained quartzose migmatite granite (Appendix 6) of the Svecofennian orogenic metamorphic complex of the **South Estonian block** (interval 514.2–548.5 m; Appendix 1, sheets 19, 20). The rocks were formed under high pressure, and have cataclastic and gneissic fabric and consist mainly of quartz (50–70%), microcline (10–30%), plagioclase (0–20%), chloritized biotite (5–10%) and muscovite (0–5%) with zircon as an accessory mineral and sericite, epidote and iron oxides as secondary minerals. The weathering crust of Palaeoproterozoic rocks is 6.8 m thick.

The eroded and weathered surface of the Palaeoproterozoic crystalline basement of Estonia is overlain by the **Ediacaran** (Upper Vendian) rocks of Proterozoic age (Fig. 2). The Ediacaran sequence is represented only by its uppermost part corresponding to the Kotlin Stage (Mens & Pirrus 1997a). The thickness of the stage is 57.2 m in the Mehikoorma (421) core (interval 457.0–514.2 m; Appendix 1, sheets 14–18). Based on differences in rock composition (Appendix 2, T-43; Appendixes 7–11) and variation in colour, three successive formations are distinguished in the Kotlin Stage: (from below) the Gdov (sandstones; thickness 6.0 m), Kotlin (silty claystone; 46.8 m) and Voronka (siltstone; 4.4 m) formations.

The thickest and stratigraphically the most representative Ediacaran sections occur in northeastern

Estonia. They thin out rather rapidly and change in lithology in a westerly direction, being absent in southwestern Estonia and some local structures (Mens & Pirrus 1997a, fig. 11).

Unlike many regions of the world, the Ediacaran sequence of the East European Platform is represented by a noncarbonate complex of siliciclastic sediments (Pirrus 1993). The investigation of Ediacaran sediments, especially of their authigenic mineral assemblage and sedimentary structures, has shown that the Kotlin sedimentation took place in brackish environments under cool and humid climatic conditions (Mens & Pirrus 1974; Pirrus 1992). Fossils are of uneven distribution and low diversity, occurring usually in grey rocks in the middle of the stage (Paškevičiene 1980).

The Ediacaran sediments in the Mehikoorma (421) section are overlain by Cambrian rocks – the Lower and Middle Cambrian and Furongian (Upper Cambrian).

Lower Cambrian silty claystones and sandstones (interval 414.2–457.0 m; Appendix 1, sheets 15, 16; Appendix 2, T-42; Appendixes 7–12) belong to the Lontova Formation corresponding to the Lontova Stage that represents the lowermost Lower Cambrian in Estonia (Mens & Pirrus 1997a). The lower boundary of the series is distinct in eastern and central Estonia. It is marked by the appearance of primitive mineralized skeletal fossils and changes in the assemblage of ichno- and phytofossils and in the mineral composition (Pirrus 1993; Mens & Pirrus 1997b).

The Lower Cambrian succession is incomplete in Estonia and varies largely from region to region. The so-called pretrilobite Lower Cambrian (Lontova Stage) sediments are widespread and in places (northern Estonia) over 80 m thick. They are missing only in southern Estonia and on some peninsulas on the south coast of the Gulf of Finland (Mens & Pirrus 1997a, fig. 14).

In southeastern Estonia a kaolinitic crust of weathering is observed in the uppermost part of the Lontova Stage (Mens & Pirrus 1997b). The chemical weathering in the interval of 414.2–418.5 m in the Mehikoorma (421) core points to subaerial erosional processes and intensive weathering known from tropical climate (Kespre 2002). In the weathered part the content of quartz and kaolinite increases, while chlorite and K-feldspar disappear completely. Considerable vertical changes take place also in the chemical composition of oxides, except for TiO_2 (Appendix 12).

Most of present-day Estonia was submerged in the Lontova Age. A small proportion of sandstone (the basal beds of the Lontova Stage) formed not far from the shoreline points to a slow marine transgression and smooth topography of both the basin bottom and surrounding source area. Later, massive accumulation of argillaceous deposits took place in the Lontova Age over the entire East European Platform under quiet hydrodynamic conditions farther from the coast (Mens & Pirrus 1986, 1997b, fig. 134B). In southeastern Estonia the crust of weathering in the uppermost part of the Lontova Stage shows that the territory stayed for a long time above sea level, which caused subsequent subaerial weathering of the topmost beds of the Lontova Stage. This process is marked by distinct changes in the fossil assemblage (Mens & Pirrus 1997b).

Middle Cambrian sandstones (interval 379.4–414.2 m; Appendix 1, sheets 14, 15; Appendixes 7–10) in the southeastern part of Estonia belong to the Paala Formation (see Mens & Pirrus 1997a). In the Mehikoorma (421) section, the basal beds of the formation consist of clayey siltstone.

The Paala Formation is distributed mainly in southeastern Estonia and is missing in northern and northwestern areas (Mens & Pirrus 1997a, fig. 22). Its registered maximum thickness is 40 m. Usually the grain-size of the rocks in the lower and upper parts of the formation is coarser than in its middle part. Muscovite and feldspar are rare (Appendixes 7, 8). In the group of heavy minerals zircon (Appendix 9) is always the index mineral (Mens & Pirrus 1997a).

The Middle Cambrian sandstone overlies older rocks with a hiatus. The succession of Middle Cambrian rocks in the East European Platform is incomplete and the evolution of the area is difficult to follow (Mens & Pirrus 1997a). The data available show that slow transgression of sea advanced from west to east and deposits accumulated probably in offshore conditions (Mens & Pirrus 1997b, fig. 135B).

Furongian (Upper Cambrian) sandstones and silty claystones (interval 371.4–379.4 m; Appendix 1, sheet 14; Appendixes 7–10) in southeastern Estonia belong to the Petseri Formation (see Mens & Pirrus 1997a), corresponding to the global Paibian Stage and representing the lowermost Furongian in Estonia. The rocks lie between older and younger ones with hiatuses. The sedimentation took place between two rapidly subsiding areas in the east and west (Mens & Pirrus 1997b, fig. 135C).

The Petseri Formation with a maximum thickness of 10.7 m is known only in core sections in a limited area in southeastern Estonia. In complete sections the formation can be subdivided into three parts: upper and lower sandstones and a silty claystone complex between them, in the middle of the formation. The sandstones contain skeletal fragments of inarticulate brachiopods and glauconite (Mens &

Pirrus 1997a). The claystone has yielded shells and fragments of lingulates and acritarchs (see Paalits in this volume).

In the Mehikoorma (421) core, the Petseri Formation is represented by the lower sandstone (374.3–379.4 m) and the middle claystone complex (371.4–374.3 m). The latter is multicoloured in its upper part (grey with ochreous and ruddy spots, stripes and bedding planes), indicating uplift of the territory accompanied with erosion and subaerial weathering in the Late Cambrian.

The Furongian sediments are overlain by **Ordovician** (Lower, Middle and Upper) dolostones, limestones (in the middle of the section) and marlstones (interval 246.2–371.4 m; Appendix 1, sheets 9–13; Appendixes 3, 4).

The Ordovician fauna is well investigated in Estonia and widely employed in biostratigraphical correlations. Systematic data on conodonts, chitinozoans and macrofossils are used also for the biostratigraphical subdivision of the Mehikoorma (421) section (see Männik & Viira, Nõlvak, and Hints in this volume).

The present paper makes use of the correlation chart for the Ordovician of Estonia by Nõlvak (1997, p. 54, table 7). During the Ordovician, the Baltoscandian area was the northern part of a shallow cratonic sea, surrounded by the Fennosarmatian land (Männil 1966; Põlma 1982; Jaanusson 1995; Nestor & Einasto 1997). The Mehikoorma (421) core log represents the transitional sediments between the North Estonian and Central Baltoscandian Confacies belts located in the marginal area of the Livonian Tongue (Fig. 3).

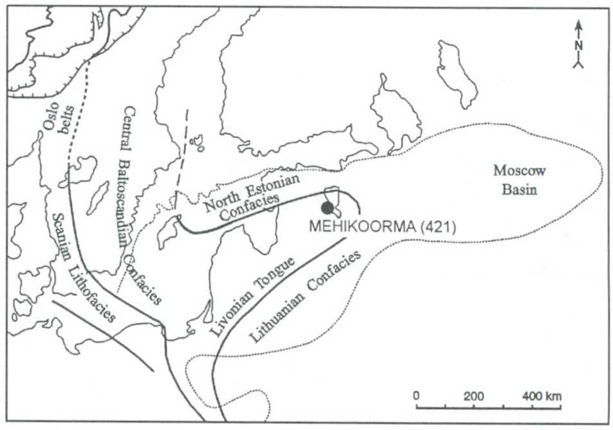

Fig. 3. Baltic Ordovician confacies belts (after Jaanusson 1995, modified from Nõlvak 1997).

Lower Ordovician glauconitic dolostones (interval 371.38–371.4 m; Appendix 1, sheet 13) are represented by the Zebre Formation corresponding to the indivisible Hunneberg–Billingen stages. The formation spreads in a limited area in southeastern Estonia. Its thickness increases towards the east up to 6.2 m (Meidla 1997). In the Mehikoorma (421) section the lower boundary of Ordovician rocks is marked by a sharp wavy yellow goethitized discontinuity surface (Appendix 3, D-21). The lowermost Ordovician (Tremadoc) rocks known from northern Estonian sections (Heinsalu & Viira 1997) are missing here.

Middle Ordovician dolostones and the uppermost limestones (interval 330.7–371.38 m; Appendix 1, sheets 12, 13; Appendixes 4, 13) are represented by the Kriukai, Šakyna, Baldone, Segerstad, Stirnas, Väo and Kõrgekallas formations corresponding to the Volkhov, Kunda, Aseri, Lasnamägi and Uhaku stages. Because of secondary dolomitization, lithological features of rocks (variation of colours, content of grains and clay, bedding planes, disposition of marlstone interbeds) are indistinct in the Mehikoorma (421) core (Appendix 2, T-32...40; Appendix 4). Dolostones of the lower and middle parts of the Middle Ordovician are brownish-red, except for those of the Šakyna Formation (Appendix 3, D-13...20). The upper part comprises grey limestones and dolostones.

Grey dolostones of the Šakyna Formation contain scattered glauconite grains (diameter mainly 0.3 mm; Appendix 3, D-19). The lower part of the Väo Formation contains elliptical, poorly preserved iron (mostly goethitic) ooliths less than 0.5 mm in diameter. On this level ooliths occur usually in the sections of the transitional area and northern Estonia (Põlma 1982). Mainly pyritized weak discontinuity surfaces are present in the lower and upper parts of the Väo Formation.

Middle Ordovician sediments have formed at the boundary between the marginal and central confacies belts (see Fig. 3) in variable conditions between the regression and transgression maximums (Nestor & Einasto 1997), subsequently in the open shelf environment.

Upper Ordovician limestones with marlstone interbeds (interval 246.2–330.7 m; Appendix 1, sheets 9–12; Appendix 2, T-21...31; Appendix 4) belong to the Dreimani, Tatruse, Kahula, Variku, Rägavere, Mõntu, Saunja, Tudulinna and Jonstorp formations corresponding to the Kukruse, Haljala, Keila, Oandu, Rakvere, Nabala, Vormsi and Pirgu stages.

Limestones of the Dreimani Formation contain pyritized bioclasts (trilobites, bryozoans, brachiopods, ostracods, gastropods, echinoderms), mostly accounting for 10–25%, in places for 30–40% of rock (Appendix 2, T-31 and T-32). Rare slightly kerogenous limestone and marlstone interbeds (thickness 1–20 cm; Appendix 3, D-13), as well as three slightly limonitized, indistinct dicontinuity surfaces in the uppermost part of the formation, point to an analogy with the northern Estonian sections (see Appendix 4).

The Mehikoorma (421) borehole in East Estonia is located in the marginal area of the known occurrence of volcanic ashbeds (Bergström *et al.* 1995). The limestones of the Idavere and lowermost Jõhvi substages (see Appendix 1, sheet 11) are intercalated with seven 0.2–4 cm (one 15 cm) thick beds with minor volcanic component (see also Appendix 4). K-bentonite at the lower boundary of the Keila Stage is a reliable marker level in Estonian sections. In the Mehikoorma (421) section it lies at 306.7 m (observed thickness 2 cm). X-ray diffractometry (see Kiipli & Kallaste in this volume) has revealed the identity of this bentonite bed with the widely distributed Kinnekulle bed, but correlation of biostratigraphical data needs additional study (see Nõlvak in in this volume). Pure K-bentonite has been identified only 70 cm higher, in a 1 cm thick bed at a depth of 311.4 m. In the middle of the Pirgu Stage (251.8 m; Appendix 1, sheet 9), K-feldspar and some biotite flakes of volcanic origin are found in a polished bedding plane of dolostones.

In southern Estonia, in the zone of confacies transition, the argillaceous limestones and marlstones of the Kahula Formation are overlain by alternating silty marl- and claystone, clayey siltstone and argillaceous limestones of the Variku Formation (Ainsaar & Meidla 2001). The lower boundary of the Variku Formation is marked by the disappearance of limestone interbeds and notable increase in the silt-sized component of the upper rocks.

In the Mehikoorma (421) core, the Variku Formation (interval 286.4–298.5 m; Appendix 1, sheet 11) is represented by silty and sandy marlstones with clayey siltstone, argillaceous limestone and dolostone interbeds. The content of insoluble residue increases markedly on the lower and upper boundaries of the formation, accounting for 36–76% in silt- and sand-containing marlstones and siltstones (Appendixes 13, 14). A thin section (Appendix 2, T-29) and the insoluble residue of chitinozoan and conodont samples contain silt- and sand-sized quartz grains whose numbers vary largely in the core.

The marlstone of the Variku Formation usually contains bioclasts less than 10%. In the Mehikoorma (421) section, the upper half of the formation includes up to 5 cm thick layers with up to 30% fine and coarse bioclasts of brachiopods and bryozoans (Appendix 3, D-11). Bioclast-bearing layers are very common in the uppermost 3.5 m and similarity with marlstones of the Hirmuse Formation known from northern sections of Estonia is remarkable. This indicates transition between different confacies belts in East Estonia, near Mehikoorma. Besides, in places fine pyritized lines and pyritized crusts point to the bioturbation of sediments.

In the Mehikoorma (421) core, the Variku Formation is overlain by very finely crystalline and microcrystalline limestone (Appendix 3, D-10) and dolostone of the Rägavere Formation (interval 283.4–286.4 m; Appendix 1, sheets 10, 11). Only the lowermost limestone contains silt. The insoluble residue of chitinozoan and conodont samples taken from different levels has yielded glauconite, which is missing in the Rägavere Formation in northern sections. Glauconite is known from the upper part of the Mossen Formation in southern Estonia, where the Rägavere or Mõntu Formation (interval 278.0–283.4 m in the Mehikoorma section) is underlain by argillaceous limestones and calcitic marlstones (Põldvere 2001; Põldvere *et al.* 2003). The correlation of the Rägavere Formation is complicated because the lithology of rocks is variable in the facies transition area, showing the characteristics of both northern and southern sections.

Calcitic dolostone of the Mõntu Formation (interval 278.0–283.4 m; Appendix 1, sheet 10) contains glauconite, found in the insoluble residue of chitinozoan and conodont samples and in a thin section (Appendix 2, T-26). In its lower part, bluish-grey impregnated dolomitic marlstone interbeds with secondary Fe-chlorite occur at 282.7 m (Appendix 3, D-9). Vuggy dolostones are in places porous and comprise pyrite like the overlying light beige dolostones of the Saunja Formation (interval 265.8–278.0 m; Appendix 1, sheet 10; Appendix 3, D-7 and D-8).

Argillaceous dolostones and dolomitic marlstones of the Tudulinna Formation (interval 256.9–265.8 m; Appendix 1, sheet 10) are in places bioturbated and contain rare fine bioclasts (Appendix 2, T-23). Some layers in the lower part of the formation are with violet tinge.

Brownish-red argillaceous dolostones and dolomitic marlstones of the Jonstorp Formation (interval 246.2–265.8 m; Appendix 1, sheet 9), with violet, yellow and light grey interlayers, contain fine and coarse bioclasts, locally up to 25% (Appendix 2, T-21 and T-22; Appendix 3, D-2…6). In places rare quartz grains are found. Carbonate clasts occur in the upper part and under the discontinuity surface at 250.8 m. The siltstone interbed at 247.8–248.0 m comes actually from the lower beds of the Devonian sequence and was placed on this level due to technical mistake made during drilling (see Kurik in this volume).

The Upper Ordovician sediments in the Mehikoorma area have formed in the conditions of gradually deepening or shallowing open shelf (Nestor & Einasto 1997). A notable change in sea level and water turbidity took place in the late Keila Age, causing an increase in the accumulation of silty, sandy and clayey sediments. The terrigenous silt- and sand-sized quartz component in the sediments of the Oandu Age is explained as reworking product of shelf se-diments during the sea level drop and following flooding (Ainsaar 2001). A hiatus on the Ordovician–Devonian boundary resulted from the pre-Devonian denudation, that is why the lowermost Devonian rocks contain both Devonian and Ordovician fossils (see Kurik in this volume). Rocks of the upper half of the Upper Ordovician are secondarily dolomitized in the Mehikoorma (421) core. Features of chemical weathering are missing in the uppermost Ordovician, but are observed in redeposited Devonian rocks (Kleesment *et al.* 1980).

The **Devonian** sediment complex (Emsian, Eifelian and Givetian stages) is 206.7 m thick (interval 39.5–246.2 m; Appendix 1, sheets 2–9) and lies unconformably on the Ordovician dolostones. The lithology of the Devonian part of the section is described by A. Kleesment in a separate chapter of this volume.

In the Devonian, the Estonian territory was covered by terrigenous sediments of epicontinental shallow sea, transported mainly from the northward Scandinavian Caledonian foreland basin (Kleesment *et al.* 1980; Kleesment 1997; Plink-Björklund & Björklund 1999). After a long denudation period and break in sedimentation, the siliciclastic underwater delta plain deposits of Rēzekne (Appendix 3, D-1) and Pärnu ages accumulated on the Upper Ordovician carbonate rocks. At the end of the Rēzekne Age, delta deposits in southeastern Estonia were replaced by nearshore tide-influenced mixed carbonate-siliciclastic sediments with features indicative of local sedimentation breaks.

The following transgression of the Narva Age reached its maximum in Vadja time. At the end of Vadja time the sea retreated and denudation of earlier deposited layers (local break) took place in its northern part. By the beginning of Leivu time the sea again flooded the area. Regression started in the conditions of the sea level fluctuation. First the nearshore carbonate and mixed carbonate-siliciclastic sediments accumulated. Kernavė time shows a clear regressive trend. Influx of terrigenous material and influence of the freshwater streams became more pronounced. The sedimentation history of the Narva Age has been examined in more detail in Tänavsuu (2004).

The regression continued in the Aruküla Age. The accumulation of sediments was fluvially influenced by freshwater streams and the sediments formed in places of distal and proximal delta fronts. However, the cyclic structure of the sequence suggests frequent sea-level fluctuations and fluvial influence on shallow marine sediments (Kleesment 1994).

Accumulation of the delta plain sediments continued on the Estonian territory in the Burtnieki Age. The regression of sea was characterized by short-term breaks in sedimentation.

The **Quaternary** cover in the Mehikoorma (421) core is 39.5 m thick (Appendix 1, sheets 1, 2) and is formed of Upper Pleistocene glaciolacustrine sediments. The Middle Devonian claystone is overlain by a 0.7 m thick till bed with diorite and porphyrite pebbles, covered by whitish-grey silty medium- to fine-grained sand. Sandy sediments of the uppermost part (7.7 m) are brownish-grey.

DEVONIAN

The Devonian sequence in the Mehikoorma (421) core (Appendix 1, sheets 1–9) is well studied (Appendixes 15–19), especially the contact with the Ordovician and the lower half of the section comprising the Rēzekne, Pärnu and Narva regional stages.

The Rēzekne Stage is separated from the underlying Ordovician rocks with a major stratigraphical gap. The lower Devonian boundary is lithologically not clear. The boundary at 246.2 m has been defined on the basis of natural gamma log data (Appendix 5; Kajak *et al.* 1974) and previous investigations (Kleesment *et al.* 1975; Ljarskaja & Kleesment 1981; Valiukevičius 1998).

The uppermost Ordovician rocks are not weathered. The grey silty dolomitic marlstone of Pirgu age is overlain by a 20 cm interbed of dolomitic siltstone, covered with a complex (245.0–246.0 m) of dolomitic marlstone, silty dolostone and sandstone. A similar succession of rocks has been registered at the Ordovician–Devonian boundary in many cores of Southeast Estonia and studied in detail in six cores (Kleesment *et al.* 1980). The lithological and palaeontological study of these sections shows that the upper Ordovician rocks were destroyed by pre-Devonian denudation. Partly the Ordovician material was redeposited in the Devonian. Thus, Ordovician crinoid fragments occur together with Devonian fossils at a depth of 245.3 m in the Mehikoorma (421) core.

Siltstone with Devonian fossils was found at 247.8–248.0 m in the Mehikoorma (421) core (see Kurik in this volume). Its composition resembles that of the rock in the interval of 246.0–246.2 m (Appendix 15) at the Ordovician–Silurian boundary. Greenish-grey and brownish-red dolostones and dolomitic marlstone between these levels (thickness 1.6 m; outwardly typical of the Pirgu Stage) contain conodonts of Pirgu time. Judging from all data available, the lower siltstone interbed (247.8–248.0 m) is actually of Devonian age and was mistakenly placed into the underlying Upper Ordovician (Pirgu Stage) section during drilling.

The **Rēzekne Regional Stage** in the Mehikoorma (421) core (220.3–246.2 m, Appendix 1, sheets 8, 9; core yield 75%) is the stratotype of the Mehikoorma Formation recognized in eastern Estonia. Kalju Kajak (Geological Survey of Estonia) proposed to establish this formation in the regional Devonian correlation chart. The Devonian working group of the Estonian Commission on Stratigraphy approved the new chart in 1999 (manuscript version).

The Rēzekne Stage in the Mehikoorma (421) core yielded a rich fish and invertebrate fauna and miospores (Kleesment *et al.* 1975; Valiukevičius 1998; see also Mark-Kurik in this volume). Juozas Valiukevičius (Institute of Geology and Geography of Lithuania) has proposed the Mehikoorma section as a stratotype section for the acanthodian succession (Valiukevičius 1994, 1998). The real stratotype of the Rēzekne Stage is the interval of 361.8–487.8 m in the 5–Akniste core in southern Latvia (presently destroyed) and the neostratotype is the interval of 427.0–446.0 m in the 15–Ludza core in eastern Latvia (Ljarskaja & Kleesment 1981).

The basal 1.2 m (245.0–246.2 m) of the Mehikoorma Formation in the Mehikoorma (421) core is represented by yellowish- to bluish-grey sandstones and violetish-grey dolomitic marlstones with light grey dolo- and siltstone interbeds, followed upwards by 16.7 m thick (228.3–245.0 m), mainly light grey, strongly to medium-cemented sandstone with interbeds of greenish-grey claystone, grey strongly to medium-cemented siltstone (Appendix 2, T-20; Appendix 15), and dolomitic marl- and dolostone. The upper 8 m (220.3–228.3 m) is represented by greenish-grey thin-bedded silty dolomitic marlstone with interbeds of sand-, silt- and dolostone. Marlstone and dolostone are horizontally thin-bedded; some bedding surfaces of dolostones resemble discontinuity surfaces (Appendix 2, T-19). In places interlayers of argillaceous silt- or sand-containing dolomitic marlstone occur (Appendix 14, samples 225.4 and 226.6 m).

According to the mineralogical composition the sandstones of the Mehikoorma Formation (Appendixes 15, 16) qualify as subarkoses and arkoses (Folk 1980). The content of micas is notable only in siltstone layers. In the heavy fraction transparent allothigenic minerals dominate. Pyrite occurs as a significant component in dolostones and dolomite-cemented sand- and siltstones (Appendixes 16, 17). Garnet, followed by zircon, is clearly dominating among transparent heavy minerals (Appendix 18). The clay fraction sometimes contains large amounts of authigenic feldspars (Appendix 19), which is explained by a relatively high K_2O content in many samples from this level (Appendix 14). Authigenic feldspars occur in silty dolomitic marlstones and dolomite-cemented siltstone, especially in thin intercalated complexes.

The **Pärnu Regional Stage** (=Formation; 199.0–220.3 m, Appendix 1, sheet 8; core yield 60%) is represented by grey and greyish-brown, fine- to medium-grained, mainly weakly to medium-cemented sandstone (Appendix 15). Natural gamma log shows the presence of some claystone interbeds.

Cement of sandstones is represented by goethite/haematite-pigmented clay (Appendix 2, T-15...18), in places by dolomite or clay with a marked dolomite admixture (Appendix 2, T-13 and T-14). Some levels contain abundant dolomitic ooliths of concentric structure (Appendix 2, T-14). Judging from the mineralogical composition, sandstones are mainly subarkoses, with a notable content of transparent allothigenic heavy minerals. Siltstone is rich in micas and pyrite (Appendix 16, 17). Quartz grains are often fractured (Appendix 2, T-13 and T-16). The heavy transparent mineral spectrum is dominated by garnet, accompanied by zircon, tourmaline and apatite (Appendix 18). The clay fraction is dominated by illite, with considerable amounts of montmorillonite/illite mixed-layer varieties. Some samples of fine-grained rocks contain authigenic feldspar (Appendix 19).

The **Narva Regional Stage** (137.0–199.0 m, Appendix 1, sheets 5–7; core yield 60%) is represented by the Vadja (Substage), Leivu (Substage) and Gorodenka (Kernavė Substage) formations, which can be traced over a large territory in East Baltic and western Belarus. The distinction of these units is grounded on lithological as well as palaeontological data (Valiukevičius *et al.* 1986; Kleesment *et al.* 1987).

The lower, **Vadja Formation** (184.0–199.0 m, Appendix 1, sheet 7; core yield 58%) is characterized by intercalated light grey dolostone, grey dolomitic marlstone and dark grey dolomitic claystone (Appendix 13), and transitional varieties between these rock types (Appendixes 14, 15). The complex is indistinctly wavy-bedded, in places with dessication cracks. Dolomitic marlstone includes numerous slickensides, covered with dark grey or brown clay films. Aphanocrystalline dolostone is penetrated by intricate fracture sets, filled with authigenic dolomite (Appendix 2, T-11 and T-12). The lower 3 m is a breccia layer containing large amounts of clastic material (Appendix 14).

Judging from the mineralogical composition the sandstone is mica-rich arkose (Appendix 16). Pyrite is dominating among heavy minerals (Appendix 17). Garnet and zircon are the main constituents of the transparent heavy mineral spectrum. Some levels show a notable admixture of titanite (Appendix 18). In addition to illite–chlorite, the clay fraction contains mixed-layer montmorillonite/chlorite (Appendix 19).

The **Leivu Formation** (153.8–184.0 m, Appendix 1, sheets 6, 7; core yield 56%) is represented by intercalated dolomitic marlstone, and silt- and sandstone containing thin interbeds of dolo- and claystone. Rocks are mostly mottled, usually reddish-brown, with grey partings and spots. Only the lower part of the section is grey. Dolomitic marlstone forms about 50% of the section and is predominantly silty, showing transitional varieties up to siltstone (Appendix 2, T-6…8 and T-10; Appendix 15). Grains of feldspar and quartz, also mica flakes often form thin clastic-rich laminae (Appendix 2, T-6 and T-10). Sometimes patchy pigmentation by goethite/haematite follows the bedding planes (Appendix 2, T-7 and T-10). In the lower part of the formation aphanocrystalline dolostone (Appendix 2, T-9; similar to the Vadja Formation) with filled vugs and fractures forms about 5% of the section.

Thin-bedded siltstone and fine- to very fine-grained sandstone with transitional varieties concentrate in the upper part of the section and form 42% of the formation. Dolomite-cemented varieties intercalate with clay-cemented varieties (Appendix 2, T-8; Appendix 15). Silty claystone forms 3% of the section.

According to the mineralogical composition the sandstone is mica-rich arkose (Appendix 16). Iron hydroxides and micas are dominating among heavy minerals (Appendix 17). Tourmaline–apatite, in some levels garnet prevail in the transparent heavy mineral spectrum. An additional major constituent is zircon, in one case titanite (Appendix 18). Illite predominates in the clay fraction (Appendix 19).

The upper, **Gorodenka Formation** (137.0–153.8 m, Appendix 1, sheets 5, 6; core yield 68%) is represented by intercalated reddish-brown and grey sandstone, siltstone and varicoloured dolomitic marlstone, with interbeds of dolo- and claystone. Sandstone forms 60% of the section and is mostly thin-bedded and very fine-grained (Appendix 15). In the lower part of the formation sandstone is in places strongly cemented by dolomite and has globular structure. Siltstone makes up 10% of the section. Dolomitic marlstone (21%) is usually silty. Grains of clastic minerals concentrate on bedding planes; in some cases rock contains grain-rich and goethite-pigmented patches (Appendix 2, T-4 and T-5).

According to the mineralogical composition the sandstone is arkose with a notable content of micas (Appendix 16). The composition of heavy minerals varies greatly. Iron hydroxides, micas or transparent allothigenic minerals predominate (Appendix 17). The transparent heavy mineral spectrum is represented by the zircon–garnet–tourmaline–apatite assemblage (Appendix 18). The tourmaline content is higher than usual in the rocks of the Gorodenka Formation (Valiukevičius *et al.* 1986; Kleesment & Kurik 1997). Similar to the Leivu Formation, the clay fraction of the Gorodenka rocks shows a stable illite-dominated composition (Appendix 19).

The **Aruküla Regional Stage** (Formation; 46.9–137.0 m, Appendix 1, sheets 2–5; core yield 57%), consisting of three members, is complete in the Mehikoorma (421) core (Kleesment 1994).

The lower, **Viljandi Member** (112.0–137.0 m, Appendix 1, sheet 5) is represented by reddish-brown, fine- and very fine-grained sandstone (share 35%) and brownish-grey and grey siltstone (55%), with thin interlayers of varicoloured silty clay- and marlstone (10%). Transitional rocks between sand- and siltstones are common (Appendix 2, T-2). Weakly- to medium-cemented sand- and siltstones (Appendix 15) are horizontally thin-bedded, in places cross-bedded. Bedding surfaces are often wavy, mica flakes or clay films usually concentrate on bedding planes. Cement is mostly goethite/haematite-pigmented clay. Rare interbeds are strongly cemented by dolomite (Appendix 2, T-2 and T-3; Appendix 13).

The **Kureküla Member** (88.5–112.0 m, Appendix 1, sheet 4) is represented by a horizontal thin-bedded complex of reddish-brown and grey siltstone (share 68%) and reddish- and yellowish-brown sandstone (30%) with a few interbeds of multicoloured claystone. Rocks are mainly weakly to medium-cemented by clay, and all varieties from medium-grained sandstone to siltstone are present (Appendix 15). Slickensides in siltstone beds are covered with clay films.

The upper, **Tarvastu Member** (46.9–88.5 m, Appendix 1, sheets 2–4) is represented by a complex of alternating greyish- and reddish-brown siltstone (share 48%), reddish-, purplish- and yellowish-brown sandstone (22%) and multicoloured silty claystone (30%) typical of this unit. Sand- and siltstones (Appendix 15) are weakly to medium-cemented by goethite/haematite-pigmented clay. Rare strongly dolomite-cemented interbeds, sometimes with globular structure, are present (Appendix 14). Dolomite-cemented silty claystone layers include transitions to dolomitic marlstone laminae (Appendix 2, T-1).

The sandstones of the Aruküla Stage, having the mineralogical composition typical of this formation (Kleesment 1994), are mostly subarkoses, with mica-rich interlayers at some levels (Appendix 16). In the heavy mineral spectrum mostly ilmenite and transparent allothigenic minerals prevail. Some samples from the Viljandi Member and one from the Tarvastu Member show a high content of micas. Besides, iron hydroxides occur in notable amounts in the Viljandi Member and the leucoxene content is somewhat higher than average in the Kureküla and Tarvastu members (Appendix 17). The heavy transparent mineral suite is represented by the zircon–apatite–tourmaline assemblage with a considerable admixture of garnet. Contrary to zircon, the apatite content is the highest in the Viljandi Member and in the lower part of the Tarvastu Member (Kleesment 1994; Kleesment & Mark-Kurik 1997). The rocks of the Kureküla and Tarvastu members contain also rutile in significant amounts and staurolite admixture is found in the Tarvastu Member (Appendix 18). The mineralogical composition of the clay fraction of the Viljandi Member is similar to that of the Gorodenka Formation, with clear predominance of illite. Tarvastu rocks are characterized by a considerable admixture of kaolinite (Appendix 19).

The **Burtnieki Regional Stage** (39.5–46.9 m, Appendix 1, sheet 2; core yield 50%) is represented only by the basal part of its lower unit, the Härma Member. Its lower part contains brownish-grey sandstone, upsection grading into greenish-grey siltstone and reddish-brown multicoloured rubbly claystone. Only one single grain-size distribution analysis was made from this interval (Appendix 15).

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DISTRIBUTION OF DEVONIAN FOSSILS

The Devonian sequence of the Mehikoorma (421) core (Appendix 1, sheets 2–9) is comparatively rich in fossils. A total of 45 samples from this core yielded fossil fishes (particularly fish microremains), miospores (also very rare acritarchs), invertebrates (mainly fragments of lingulates, rare conchostracans, poorly preserved ostracods) and trace fossils (Appendixes 20–23). Samples were collected by Anne Kleesment (Institute of Geology at Tallinn University of Technology), Kalju Kajak (Geological Survey of Estonia) and Juozas Valiukevičius (Institute of Geology and Geography of Lithuania) in the 1970s and 1980s. The Devonian fossils come mainly from the lower part of the Rēzekne Stage and middle part of the Narva Stage (Leivu Substage). They are relatively numerous also in the upper part of the Narva Stage (Kernavė Substage) and lower part of the Aruküla Stage (Viljandi Member) (Appendix 20). Only one sample from the Vadja Substage (Narva Stage) contained fossils – miospores. Samples from the Tarvastu Member (Aruküla Stage) and Härma Member (Burtnieki Stage) yielded no fossils.

The lower boundary of the Devonian (and the Rēzekne Stage) in the Mehikoorma (421) core has been defined at a depth of 246.2 m (Kleesment *et al.* 1975; Valiukevičius 1994; Kleesment & Mark-Kurik 1997). In 1976 Anne Kleesment took a sample from the level of 247.9 m and in the same year Elga Mark-Kurik identified fossils in it. Co-occurrence of a psammosteid fragment and an acanthodian scale of *Diplacanthus* type with otoliths showed that the rock, light grey siltstone, was undoubtedly of Devonian age. However, new examination of the core by Anne Põldvere (Geological Survey of Estonia) and comparison of laboratory data and earlier descriptions of Anne Kleesment and Lembit Põlma (both Institute of Geology at Tallinn University of Technology) show that the sampled siltstone interval (thickness 20 cm) actually represents lower beds of the Devonian sequence and has been placed in the underlying Upper Ordovician (Pirgu Stage) section by mistake during drilling. The sample from 247.9 m (corrected depth 246.0–246.2 m) is still valuable, as it has yielded rare fish remains – otoliths.

Several samples taken above the Devonian–Ordovician boundary contain either Ordovician conodonts together with Devonian fish remains (at 244.6–245.8 m) or Ordovician fossils, mostly conodonts (at 245.2 m). At a depth of 246.1 m only Devonian fish fossils, acanthodian and sarcopterygian scales and otoliths have been found. Thus, the about 1.6 m thick interval below 244.6 m in the Mehikoorma (421) core includes mixed Devonian fishes as well as redeposited Ordovician fossils. This phenomenon was noticed already by Kleesment *et al.* (1980). Acanthodian scales, together with worn stem segments of crinoids, were recovered at 245.3 m.

Two stratigraphical units that the Mehikoorma (421) drill hole penetrates have deserved special attention and have been described in detail. These are the Rēzekne Stage (Kleesment *et al.* 1975; Valiukevičius 1994, 1998) and the Narva Stage (Valiukevičius *et al.* 1986).

Using the data by Anne Kleesment and Elga Mark-Kurik, Ljubov A. Ljarskaja divided the Rēzekne Stage (Formation) in the Mehikoorma (421) core into two members and three rhythms or cycles (Ljarskaja 1978, fig. 2). She traced these rhythms in many cores of eastern Estonia and Latvia, and in northeastern Belarus. Valiukevičius (1994, 1998) accepted the subdivision of the stage (formation) into two units – subformations – in the Mehikoorma (421) core (actually, a unit next in rank below a *formation* is a *member*; Salvador 1994, p. 36). According to Valiukevičius (1998, p. 14), the upper subformation at a depth of 220.3–229.2 m is represented by dolomitic marlstones with siltstone and sandstone interlayers. The lower subformation at 229.2–246.2 m is composed of light grey and greenish-grey fine sandstone interbedded with sandy siltstones and claystones. Its lowermost part (243.0–246.2 m) comprises greenish-grey dolomitic claystones and horizontally bedded dolomitic siltstones with lenses of marlstones and dolostones. In Belarus, the Emsian Vitebsk Regional Stage, coeval with the Rēzekne Stage, is subdivided into two units: the Obol′ and Lepel′ beds (Kruchek *et al.* 2001, pp. 193–194).

Valiukevičius (1994) chose the interval of 199.0–246.2 m, embracing both the Rēzekne and Pärnu stages in the Mehikoorma (421) core, as the reference section (stratotype) of the *Laliacanthus singularis* acanthodian Biozone. The Pärnu Stage does not contain the index species. In Estonia the Tori and Oore outcrops are considered as the localities of the upper part of the *Laliacanthus singularis* Zone (Valiukevičius 1994). The members (subformations by Valiukevičius) of the Rēzekne Stage (Formation) have yielded different miospore assemblages, initially identified by Gene Vaitiekūnienė, Vilnius, Lithuania (Kleesment *et al.* 1975) but later corrected by Tamara G. Obukhovskaya, Minsk, Belarus (Kõrts & Mark-Kurik 1997, table 30). Valiukevičius (1995, fig. 1; 2000, fig. 3) recognized the Emsian age of the Rēzekne Stage, correlating it with the *patulus* Zone, the latest Emsian Standard Conodont Zone. In several other papers (e.g. Valiukevičius 2002) he still attributes the Eifelian age to the Rēzekne acanthodian association, since many Middle Devonian acanthodian genera appeared first in Rēzekne time. E. Mark-Kurik has a different point of view. Using interregional correlation based on various placoderms, she dates the Rēzekne Stage as Emsian (for discussions see Mark-Kurik 1991, p. 19; 2000, p. 310; Valiukevičius 2000, pp. 274–276).

Three samples (from 244.1–244.6, 223.0–224.0 and 195.0–199.0 m) yielded **miospores** of Rēzekne (Early Devonian) and Narva (Middle Devonian) age (Appendixes 21, 22). The Rēzekne miospores are listed and figured in Kleesment *et al.* (1975, pls 1–3). The list of the miospores (identified by G. Vaitiekūnienė) of the Vadja Substage of the Narva Stage (sample from 195.0–199.0 m) is available in Valiukevičius *et al.* (1986). Later the lists were modified by T. G. Obukhovskaya, who gave more recent names to some taxa (see Kõrts & Mark-Kurik 1997, table 30). Two aspects are worthy of note: (1) miospores from the upper and lower parts of the Rēzekne Stage differ significantly, only *Retusotriletes simplex* and *Leiotriletes microrugosus* occur in both parts of the stage, and (2) in the Vadja Substage, acritarchs were discovered together with miospores.

Lingulates are the most common **invertebrates** in the Mehikoorma (421) core, yet preserved mainly as small fragments. Rare poorly preserved ostracods, a conchostracan probably from the genus *Glyptoasmussia* and an unidentified ichnofossil come from the lower part of the Rēzekne Stage (Appendixes 20, 21).

Fossil fishes (agnathans and gnathostomes) are represented by all groups common in the late Early Devonian and Middle Devonian of the Baltic area but, depending on the exoskeleton structure and character of squamation, the number of remains of different fishes varies significantly. The faunas are dominated by psammosteid heterostracans having tubercles more or less easily separable from exoskeletal plates. In the Mehikoorma (421) core (Appendixes 20, 21) the psammosteid remains are found either as isolated tubercles or small plate fragments, which are usually insufficient for reliable identification of the taxa they belong to. Exceptional is a fragment of *Pycnosteus palaeformis*, the key fossil of the Viljandi Member of the Aruküla Stage (Mark-Kurik 2000) found at 124.5 m. Placoderms (arthrodires, antiarchs), another important component in the late Early and Middle Devonian faunas, can hardly be identified in the core samples even on the generic level. Their remains are more common in the Leivu Substage (Narva Stage).

Small discrete scales of **acanthodians** are abundant in almost all core samples, which makes acanthodians a biostratigraphically valuable group. Sometimes fin spines of acanthodians are found. In the Mehikoorma (421) core, two spines belonging probably to *Haplacanthus* and *Archaeacanthus* were discovered in the Narva Stage (Leivu Substage) and Aruküla Stage (Viljandi Member), respectively. Most of acanthodian scales known from the Rēzekne Stage come from the lower part of the stage (Appendix 23).

According to Valiukevičius (1994), the boundary between the lower and upper parts of the Rēzekne Stage lies at the level of 229.2 m. Eleven acanthodians, almost all belonging to different genera, come from the lower unit. The distribution of these species is rather variable in the Baltic area (Valiukevičius

2000, fig. 1). For example, *Nostolepis gracilis* occurs throughout the Lower Devonian and ranges into the Rēzekne Stage. "*Pruemolepis wellsi*" has a similar range but reaches even the top of the Narva Stage. *Acanthoides*? sp. C, *Ectopacanthus flabellatus* and *Markcanthus parallelus* come from the Lower Devonian Kemeri Stage and the overlying Rēzekne Stage. The first of these acanthodians reaches even the Vadja Substage (Narva Stage). In the Mehikoorma (421) core *Acanthoides*? sp. C is found in both parts of the Rēzekne Stage, being fairly numerous in the lower part. *Ptychodictyon ancestralis* is known only from the Rēzekne Stage (Appendix 23).

Laliacanthus singularis, *Diplacanthus kleesmentae*, *Rhadinacanthus primaris* and *Cheiracanthus gibbosus* characterize both the Rēzekne and Pärnu stages (Valiukevičius 1994). *Laliacanthus singularis* is the key fossil of these stages. Scales of *Diplacanthus kleesmentae* are more numerous than those of the other listed acanthodians. The two last mentioned species are found in both parts of the Rēzekne Stage in the Mehikoorma (421) core. *Nostolepis* sp. ranges up to the Narva Stage (Vadja Substage) in other sections (Appendix 23).

Cheiracanthus longicostatus, *C. brevicostatus*, *Acanthoides*? sp. B and D are very long-ranging acanthodians (Valiukevičius 1994). In the Baltic area, they occur in the Rēzekne Stage and range further up, reaching the Burtnieki Stage and even the Amata Stage (*Acanthoides*? sp. B and D). These *Acanthoides*? species are the most frequent ones in the Mehikoorma core, particularly in the Leivu Substage (Narva Stage). *Acanthoides*? sp. B occurs also in the Pärnu Stage, otherwise poor in fossils (Appendixes 21, 23).

Haplacanthus and *Archaeacanthus*, two acanthodian genera based on fin spines, have also a long range – from Narva (Vadja Substage) to Gauja stages. *Ptychodictyon rimosum* ranges from the Leivu Substage (Narva Stage) to the Burtnieki Stage, being particularly characteristic of the former unit and serving as its key fossil. *Ptychodictyon distinctum*, *Rhadinacanthus balticus* and *Cheiracanthus intricatus* appear in the middle part of the Leivu Substage (Narva Stage) and range into the lower part of the Aruküla Stage (Valiukevičius 1994). In the Mehikoorma (421) core *Rhadinacanthus balticus* is the most common among these three species, showing its full range (Appendix 23).

Cheiracanthus sp. and *Diplacanthus* sp. are long-ranging in other Baltic sections (Valiukevičius 1994) (up to the Burtnieki and Amata stages, respectively). In the Mehikoorma (421) core they were identified in the Leivu Substage of the Narva Stage (Appendix 23). *Acanthoides*? sp. A has a similar long range. It is rare in that core, occurring in the Kernavė Substage (Narva Stage) and the Kureküla Member (Aruküla Stage). *Ptychodictyon sulcatum* is found in the Mehikoorma core in the Viljandi Member (Aruküla Stage), but is known to range elsewhere from the Leivu Substage (Narva Stage) to the lower part of the Aruküla Stage (Valiukevičius 1994). *Rhadinacanthus multisulcatus* appears commonly in the Aruküla Stage and ranges up to the Gauja Stage. In the Mehikoorma (421) core it comes only from the Kureküla Member of the Aruküla Stage (depth 105.5 m). To sum up, most of the acanthodian remains have been found in the Mehikoorma core in the units represented by a larger number of samples with fish fossils, i.e. the lower part of the Rēzekne Stage, the Leivu and Kernavė substages (Narva Stage), and the Viljandi Member (Aruküla Stage).

Fragmentary **sarcopterygian remains** cannot often be identified on the genus, moreover on the species level. In the Mehikoorma (421) core (Appendixes 20, 21) only a scale fragment of *Glyptolepis* and *Onychodus* teeth were identified from the Rēzekne Stage and the Leivu Substage (Narva Stage), respectively. The most frequent are cosmin covered rhombic osteolepidid scales, particularly in the Leivu Substage. Rare dipnoan remains were found in the Leivu and Kernavė substages of the Narva Stage.

Actinopterygians, represented by the genera *Orvikuina* and *Cheirolepis*, occur in the Leivu and Kernavė substages (Narva Stage) and the Viljandi Member (Aruküla Stage) in the Mehikoorma (421) core (Appendix 20). In outcrops *Orvikuina vardiaensis* is common in the topmost part of the Narva Stage. It is not excluded that *Orvikuina* specimens from the Viljandi Member (Aruküla Stage) belong to another species. After thorough study actinopterygian scales could also be attributed a biostratigraphical value, as they are relatively small, with compact structure and are met rather often in core samples.

Fish **otoliths** (ear-stones), discovered in four Devonian rock samples of the Mehikoorma (421) core, are of special interest (Appendixes 20, 21). They all come from the lower part of the Rēzekne Stage (Valiukevičius 1994, fig. 3). The otolith specimen from the uppermost sample (236.7–237.1 m) is the first find of a fossil fish otolith from the Devonian of the Baltic area (Kleesment *et al.* 1975, pl. 4, fig. 7). The other otoliths were found from the basal Devonian in the Mehikoorma (421) core at 244.6–245.8, 246.1 and 247.9 m (corrected depth 246.0–246.2 m). According to Nolf (1995), only a very small number of phosphatized otoliths are known from Palaeozoic strata. Valentina Talimaa (in Nolf 1985, p. 36) reported phosphatic otoliths from the Rēzekne and Upper Narova (= Narva) stages of the Baltic Region (Latvia, Leningrad District), as well as from the approximate age equivalents of the Rēzekne Stage in Belarus (Vitebsk Regional Stage), Severnaya Zemlya (Albanov Formation) and Spitsbergen (Wood Bay and Grey Hoek formations). She supposed that the otoliths, always occurring together with scales of *Orvikuina* and related forms, belong quite probably to actinopterygians (paleoniscoids). Coates (1993, p. 138) rejected Talimaa's interpretation concerning the chemical composition of the above otoliths, and Nolf (1995) considered their relationships with skeleton-based fossil fish taxa unknown.

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DISTRIBUTION OF ORDOVICIAN CONODONTS

Conodonts were studied in 159 samples collected as three different sets (Appendixes 4, 24, 25). The size of the samples provided by Kalju Kajak (Estonian Geological Survey) in 1974 varied from 85 to 155 g, and of those collected later by Jaak Nõlvak (in 2003) and Peep Männik (in 2004) from 250 to 1220 g. All samples but one (from a depth of 345.2 m) yielded conodonts, mostly well preserved and with CAI = 1. The number of specimens per sample varied from less than 10 to several hundreds. In general, the lower part of the studied section (up to the upper boundary of the *Amorphognathus tvaerensis* Zone) is characterized by rich and abundant conodont faunas, whereas in the strata above that level the yield of conodonts in samples decreases notably. Conodonts from the lowermost part of the studied section, up to the *Eoplacognathus suecicus* Zone, were studied by Viive Viira and above that level by Peep Männik. The study of conodonts was supported by the Estonian Science Foundation grants Nos 5406 and 5920, and by target-financed project No. 0332524s03. The collection of conodonts is stored at the Institute of Geology at Tallinn University of Technology.

The *Oepikodus evae* Zone *sensu lato*. This zone was first established by Sergeeva (1964, 1966) as a subzone in the upper part of the Billingen Stage, in the upper Billingen Substage ($B_1\beta$). In his zonal scheme Lindström (1971) defined the *O. evae* Zone as an interval between the first appearance of *O. evae* below and *Baltoniodus triangularis* above. He also noted that *O. evae* was absent in the upper part of the zone. Based on the succession of different faunas, Löfgren (1993) divided the *O. evae* Zone into four successive subunits. For the upper part of the *O. evae* Zone, where the zonal species is missing, several different names have been proposed. Viira (1974) described this interval as the *Trichonodella flabellum* (= *Periodon flabellum*) Subzone, the upper subzone of her *Oistodus lanceolatus* Zone. Bagnoli & Stouge (1997) distinguished the upper part of the *O. evae* Zone as the *Trapezognathus diprion* and *Microzarkodina* sp. A interval subzones. The last two taxa have not been found in North Estonia but *Periodon flabellum* is quite abundant in this region (Viira *et al.* 2001). In Estonia, Svend Stouge has identified *T. diprion* from the Tartu (453) core (Põldvere *et al.* 1998).

The rich and very well preserved conodont fauna in the sample from 371.1 m includes numerous specimens of *Oistodus lanceolatus* and *Drepanoistodus forceps*, also *Protopanderodus rectus*, *Drepanodus arcuatus* and *Periodon flabellum*. *Scolopodus striatus*, *Decoriconus peselephantis* and *Cornuodus longibasis* are rare. The specimens of *O. evae* are not found, but the conodont assemblage of this sample seems to refer to the upper part of the *O. evae* Zone *sensu lato*. This zone has been widely recognized all over the world. In the Taga-Roostoja (25A) core, *O. evae* was identified in a sample from the upper part of the Leetse Formation (Viira & Männik 1999). In the Mäekalda section the *O. evae* Zone *s.l.* has been identified in the Mäeküla (based on the occurrence of *O. evae*) and Päite (based on the occurrence of *Periodon flabellum*) members (Viira *et al.* 2001).

The *Baltoniodus navis* Zone. This zone was established by Lindström (1971) and informally subdivided by Löfgren (1993, 1995). Bagnoli & Stouge (1997) defined its lower and upper boundaries as the levels of the first appearance of *B. navis* and *Microzarkodina parva*, respectively. In the sample from 370.8 m *Drepanoistodus forceps* is still dominating and *Oistodus lanceolatus* is quite common. Other taxa, represented in this sample by up to ten specimens and known also from lower levels, are *Protopanderodus rectus*, *Drepanodus arcuatus*, *Decoriconus peselephantis*, *Scolopodus striatus* and *Cornuodus longibasis*. *Microzarkodina flabellum* (characteristic of the Volkhov Stage) and *B.* cf. *navis* (type species of the *B. navis* Zone) appear at this level. In general, the conodont assemblage at 370.8 m seems to be indicative of the lowermost *B. navis* Zone. In the Mäekalda section, the *B. navis* Zone occupies a short interval in the lower part of the Saka Member (= lower part of the Volkhov Stage) (Viira *et al.* 2001). In the Tartu (453) core, S. Stouge identified *B. navis* in the lower part of the Kriukai Formation (Põldvere *et al.* 1998).

The sample marked as 369.6 m most probably comes from a higher level. The diverse and quite numerous fauna of this sample indicates probably the upper Volkhov *B. norrlandicus* Zone and is slightly dominated by *Drepanodus arcuatus*. Other taxa identified include *B. norrlandicus*, *Lenodus* cf. *antivariabilis* (An), *Trapezognathus* cf. *quadrangulum* Lindström, *Protopanderodus rectus*, *C. longibasis*, *Drepanoistodus* aff. *forceps*, *M.* cf. *parva* Lindström, *Scalpellodus gracilis*, *Semiacontiodus* cf. *cornuformis* and *Parapaltodus simplissimus* Stouge. Conodonts from this sample are not indicated in Appendix 25.

The *Paroistodus originalis* Zone. The zone was defined by Lindström (1971). According to him "…It may be appropriate to refer to the range of this species (of *P. originalis*), and the main range of *Scandodus brevibasis* (=*Triangulodus brevibasis*) as the *Prioniodus originalis* zone" (Lindström 1971, p. 32). Löfgren (1995) recognized a succession of five different conodont faunas (phases) in this zone and specified its lower and upper boundaries. In Estonia, the fauna of this zone occurs in the middle part of the Volkhov Stage (Viira 1974). Conodonts in the sample from 366.9 m are typical representatives of the lower part (phases 1 and 2 according to Löfgren 1995) of the *P. originalis* Zone. Together with the zonal species, *Triangulodus brevibasis*, *B. navis*, *M. flabellum*, *Protopanderodus rectus* and *Drepanodus arcuatus* have been identified. The fauna of the *P. originalis* Zone was recognized in two samples from the Taga-Roostoja (25A) core and in one sample from the Tartu (453) core, from the middle part of the Volkhov Stage (Põldvere *et al.* 1998; Viira & Männik 1999). In the Mäekalda section, the main part of the Volkhov Stage, from the upper Saka Member below up to the lower Lahepere Member above, is assigned to the *P. originalis* Zone (Viira *et al.* 2001).

The *Baltoniodus norrlandicus* Zone. The *B. norrlandicus* Zone, as defined by Bagnoli & Stouge (1997), corresponds to the lower part of the *M. parva* Zone of Lindström (1971). The upper part of the latter zone was described as the *Lenodus antivariabilis* Zone (Bagnoli & Stouge 1997). Löfgren (2000) suggested that the whole interval between the first appearance of *B. norrlandicus* and the first appearance of *L. variabilis* (Sergeeva 1963) should be retained as the *B. norrlandicus* Zone (corresponds approximately to the *M. parva* Zone of Lindström 1971) with two subzones – *Trapezognathus quadrangulum* and *L. antivariabilis*.

The sample from 364.6 m yielded well-preserved specimens of *B. norrlandicus*, together with other conodonts characteristic of the *B. norrlandicus* Zone. Among these, *M. flabellum*, *Drepanoistodus basiovalis*, *Protopanderodus rectus* and *Drepanodus arcuatus* are known to be well represented in the Volkhov Stage, and *Semiacontiodus cornuformis* and *Scalpellodus latus* in the Kunda Stage. The next two samples, from 361.8 and 360.6 m, come probably also from the *B. norrlandicus* Zone but the number of conodonts is smaller and specimens (including the zonal taxon) are often broken. The sample from 358.4 m is the only one in the Mehikoorma (421) core where *Protopanderodus calceatus* has been found (Appendix 25). Originally, this species was described from the Horns Udde section on Öland, Sweden, from the strata corresponding to the *B. norrlandicus* Zone. Morphologically, *Protopanderodus calceatus* is closely related to *Protopanderodus rectus*. The distribution of *B. norrlandicus* is known in the Taga-Roostoja (25A) core where it occurs in the uppermost Volkhov Stage (Sillaoru Formation), and in the Tartu (453) core (in the middle part of the Kriukai Formation) (Põldvere *et al.* 1998; Viira & Männik 1999). In the Mäekalda section, the *B. norrlandicus* Zone is distinguished in a very short interval in the upper half of the Lahepere Member (uppermost Volkhov Stage) (Viira *et al.* 2001).

The *Eoplacognathus pseudoplanus* Zone. Viira (1974) proposed the *Ambalodus pseudoplanus* (= *Eoplacognathus pseudoplanus*) Zone for the upper part of the Kunda Stage. Zhang (1998) introduced the *E. pseudoplanus* Zone in its present understanding. Two subzones, *Microzarkodina hagetiana* and *M. ozarkodella* have been described in the *E. pseudoplanus* Zone (Löfgren 1978, 2000; Zhang 1998). In the Mehikoorma (421) core, typical conodont fauna of the *M. ozarkodella* Subzone of the *E. pseudoplanus* Zone occurs at 354.9, 352.9 and 351.7 m. Together with the index species, *Baltoniodus medius* and *Dapsilodus viruensis* appear at 354.9 m. Conodonts are rare in samples from 351.1 and 349.2 m, but the fauna still seems to belong to the *E. pseudoplanus* Zone. This zone is also recognized in the upper part of the Kunda Stage in the Taga-Roostoja (25A) core (Viira & Männik 1999). In the Mäekalda section, the zone corresponds to the Pakri Formation and to the lower part of the Loobu Formation, that is, to the main part of the Kunda Stage (Viira *et al.* 2001).

The *Eoplacognathus suecicus* Zone. Bergström (1971) used *E. suecicus* as an index species for the lowermost subzone of the *Pygodus serra* Zone. In the zonal scheme of Viira (1974) the *E. pseudoplanus* Zone is followed by the *E. suecicus* Zone and is considered to correlate with the Aseri Stage. According to Löfgren (1978), the lower boundary of the *E. suecicus* Zone lies in the upper part of the Kunda Stage. Zhang (1998) subdivided the *E. suecicus* Zone further into two subzones.

In the Mehikoorma (421) core, the lower boundary of the *E. suecicus* Zone lies below 348.00–348.10 m. Together with *E. suecicus* several other taxa, including *Baltoniodus prevariabilis* and *Panderodus sulcatus*, appear at this level (Appendix 25). Also the oldest specimens of *Walliserodus* in the studied section have been identified from that level. The upper boundary of the *E. suecicus* Zone is marked by the appearance of *Yangtzeplacognathus foliaceous* at 344.90–345.00 m. The *E. suecicus* Zone (the zonal species) has been identified in the Taga-Roostoja (25A) core (Aseri Stage) and in the Mäekalda section (uppermost Kunda and Aseri stages) (Viira & Männik 1999; Viira *et al.* 2001).

The *Pygodus serra* Zone. Bergström (1971) defined the *Pyg. serra* Zone and subdivided it in five subzones (from below): *E. suecicus*, *E. foliaceus* (redescribed in 1998 as *Yangtzeplacognathus foliaceus* by Zhang), *E. reclinatus* (redescribed in 1998 as *Baltoplacognathus reclinatus* by Zhang), *E. robustus* (redescribed in 1998 as *Baltoplac. robustus* by Zhang) and *E. lindstroemi*. Löfgren (1978) redefined the lower boundary of the *Pyg. serra* Zone and described it as coinciding with the level of appearance of *Y. foliaceus*. She suggested that the *E. suecicus* Subzone of Bergström (1971) should be considered as a separate zone (as proposed by Viira 1974), followed by the *Pyg. serra* Zone. Later, a new subzone, called after its nominal taxon as the *Y. protoramosus* Subzone, was described above the *Baltoplac. robustus* Subzone in the *Pyg. serra* Zone (see references in Zhang 1998). According to Viira (1974), the level of appearance of *Y. foliaceus*, and accordingly also the lower boundary of the *Pyg. serra* Zone, in Estonia coincides with the boundary between the Aseri and Lasnamägi stages; the *Y. foliaceus* and *Baltoplac. reclinatus* subzones (distinguished as zones in Viira 1974) correspond to the Lasnamägi Stage, and the lower boundary of the *Baltoplac. robustus* Subzone coincides with the lower boundary of the Uhaku Stage.

In the Mehikoorma (421) section the lower boundary of the *Pyg. serra* Zone is drawn below the sample 344.90–345.00 m in which the lowermost *Y. foliaceus* was identified (Appendix 25). Based on the occurrence of nominal taxa, all five subzones: *Y. foliaceus*, *Baltoplac. reclinatus*, *E. robustus*, *Y. protoramosus* and *E. lindstroemi* were recognized in that zone. In the studied section *Pyg. serra* appears in the *Y. protoramosus* Subzone (at 338.60–338.70 m), although the lowermost unidentifiable specimens of *Pygodus* were found already in the sample from 342.40–342.55 m, in the *Baltoplac. reclinatus* Subzone. The *Pyg. serra* Zone has been identified also in the Tartu (453), Taga-Roostoja (25A), Valga (10) and Ruhnu (500) core sections (Põldvere *et al.* 1998; Viira & Männik 1999; Männik 2001, 2003). All five subzones were recognized in the Taga-Roostoja (25A) core section. In the Valga (10) and Ruhnu (500) sections, the lowermost part of the *Pyg. serra* Zone corresponding to the *Y. foliaceus* Subzone has not been studied.

The *Pygodus anserinus* Zone. The *Pyg. anserinus* Zone was originally described by Bergström (1971) and defined as an interval between the first appearance of *Pyg. anserinus* below and of *Amorphognathus tvaerensis* above. The zone was further subdivided into two subzones: the Lower Subzone and the Upper Subzone. As defined by Bergström (1971), the boundary between these subzones corresponds to the level at which *Baltoniodus prevariabilis* is replaced by *B. variabilis*.

In the Mehikoorma (421) section, the lower boundary of the *Pyg. anserinus* Zone lies between 334.70–334.80 m (the uppermost sample with *Pyg. serra*) and 333.90–334.00 m (the lowermost occurrence of *Pyg. anserinus*). The upper boundary of the zone is marked by the appearance of *A. tvaerensis*. The boundary between the Upper and Lower subzones lies in the lowermost Dreimani Formation, below the level of 329.70–329.80 m where the lowermost specimens of *B. variabilis* were identified (Appendix 25). As in the Ruhnu (500) core, also in the studied section probable specimens of *A. inaequalis* were found in the Upper Subzone of the *Pyg. anserinus* Zone (in samples from 326.70–326.80, 325.50–325.60 and 324.60–324.70 m). In both sections, but also in the Valga (10) core, the boundary interval of the *Pyg. serra* and *Pyg. anserinus* zones is characterized by the occurrence of *Sagittodontina kielcensis*, and the uppermost part of the Upper Subzone of the *Pyg. anserinus* Zone by the occurrence of *Eoplacognathus elongatus*. In the Mehikoorma (421) and Ruhnu (500) cores, *E. elongatus* reaches the lower part of the overlying *A. tvaerensis* Zone (see below). In all these sections *Scabbardella* ex gr. *altipes* appears together with *B. variabilis*. The occurrence of *B. variabilis* below the level of appearance of *S.* ex gr. *altipes* in the Valga (10) core (see Männik 2001) was not proved by restudy of that collection.

The *Amorphognathus tvaerensis* Zone. The *A. tvae-*

rensis Zone was defined by Bergström (1971) as an interval between the first appearance of *A. tvaerensis* below and *A. superbus* above. The *A. tvaerensis* Zone was further subdivided into three subzones (from below): *Baltoniodus variabilis*, *B. gerdae* and *B. alobatus* (Bergström 1971). Recently it was proved that, at least in Estonia, the range of *A. tvaerensis* is followed by an interval (corresponding approximately to the upper Haljala and Keila stages) where *Amorphognathus* is missing or is too rare to be found in samples available from a core, and that *A. superbus* appears in the uppermost Oandu Stage and is preceded by an interval (= main part of the Oandu Stage) with *A. ventilatus* (Männik 2003, 2004).

In the Mehikoorma (421) core, *A. tvaerensis* appears at 323.80–323.90 m and is very frequent in all studied samples up the level of 312.20–312.30 m (Appendix 25). Higher, a single small M element of *A. tvaerensis* was found only in the sample from 310.60–310.70 m. However, as it is evident that this level corresponds to Datum 1 *sensu* Männik (2004) of the Mid-Caradoc Event (the abundance of conodonts decreases considerably; the fauna becomes dominated by simple-cone taxa; *A. tvaerensis* disappears), the upper boundary of the *A. tvaerensis* Zone has been drawn just above 312.20–312.30 m.

All three subzones of the *A. tvaerensis* Zone are well represented by their nominal taxa and easy to recognize in the Mehikoorma (421) core. However, the ranges of *B. variabilis* and *B. gerdae* overlap in this section (Appendix 25). The lowermost *B. gerdae* appears at 319.75–319.85 m and the uppermost *B. variabilis* was found at 317.95–318.05 m. Higher, just a few unidentifiable fragments of *Baltoniodus* were found in samples from 319.50–319.60 and 319.00–319.05 m. *Baltoniodus gerdae* is continuously present in the interval from 318.50–318.60 to 315.50–315.70 m. In all respective samples, except for that from 317.95–318.05 m, it is the only species of *Baltoniodus*. The sample from 317.95–318.05 m yielded, together with *B. gerdae*, several specimens (Pa elements) of *B. variabilis*. The studied section is the only one known in Estonia where the ranges of *B. variabilis* and *B. gerdae* overlap. It is possible that in the sections studied earlier (e.g. Valga (10) and Ruhnu (500)) the strata corresponding to this interval are missing.

The Mid-Caradoc Event interval. In the conodont sequence, the Mid-Caradoc Event is considered to correspond to the interval from the level of disappearance (considerable decrease in abundance) of *A. tvaerensis* below (= Datum 1 of the event) up to the level of disappearance of the *Semiacontiodus* lineage (= Datum 5) (Männik 2004; Appendix 25). In total, five datums (levels of major changes in conodont faunas) have been recognized in the event interval. These datums have proved useful in regional stratigraphy for recognition and correlation of several informal units (Männik 2003, 2004).

<u>The uppermost *Baltoniodus alobatus* range</u>. This informal unit corresponds to the interval between Datum 1 (level of disappearance of *A. tvaerensis*) and Datum 2 (level of disappearance of identifiable *B. alobatus*) of the Mid-Caradoc Event, and has been recognized in several sections in Estonia (Männik 2003, 2004).

Four samples of the Mehikoorma (421) core, from 311.60–311.70 to 310.20–310.30 m, come from this unit (Appendix 25). In this section, as also in the Ruhnu (500) (Männik 2003) and Taga-Roostoja (25A) cores (Viira & Männik 1999), "the uppermost *B. alobatus* range" correlates with the middle part of the Haljala Stage.

<u>The uppermost *Baltoniodus* range</u>. The unit corresponds to the interval between datums 2 and 3 of the Mid-Caradoc Event. At Datum 2 the abundance of conodonts decreases considerably and faunas become dominated by simple-cone taxa, such as *Decoriconus*, *Drepanoistodus*, *Panderodus*, etc. (Männik 2004). Ramiforms are almost missing in this interval but rare small unidentifiable fragments of *Baltoniodus* are present. The upper boundary of "the uppermost *Baltoniodus* range" corresponds to the level of disappearance of *Baltoniodus*.

In the Mehikoorma (421) section "the uppermost *Baltoniodus* range"-fauna occurs in the interval from 309.50–309.60 to 298.50–298.60 m and corresponds to the upper Haljala and Keila stages (Appendix 25). The unit has similar distribution in the Ruhnu (500) core (corresponds to the interval marked with "?" between the "uppermost *B. alobatus* range" and the *A. ventilatus* Zone in Appendix 20 of Männik 2003; actually, specimens of *Amorphognathus* spp. from this interval are, most probably, fragments of *Baltoniodus*).

<u>The *Amorphognathus ventilatus* Zone</u>. Dzik (1999) recognized an interval with *A. ventilatus* between the ranges (zones) of *A. tvaerensis* and *A. superbus*. Later, the *A. ventilatus* Zone was identified in the Ruhnu (500) core section (Männik 2003), with its lower boundary (the level of appearance of *A. ventilatus*) coinciding with Datum 4 of the Mid-Caradoc Event (Männik 2004). The upper boundary of the zone is marked by the appearance of *A. superbus*. Earlier in Estonia, the interval almost identical to the *A. ven-*

tilatus Zone was correlated with the *Ozarkodina* aff. *rhodesi–Icriodella* cf. *superba* Zone (Viira 1974).

In the Mehikoorma (421) core the lower boundary of the *A. ventilatus* Zone is tentatively drawn below the level of 297.60–297.75 m at which *Amorphognathus* re-appears, although the lowermost identifiable specimen (M element) of *A. ventilatus* was found here about a metre higher (at 296.30–296.40 m) (Appendix 25). In this section, as also in the Ruhnu (500) core, the *A. ventilatus* Zone corresponds to the main part of the Oandu Stage.

The *Amorphognathus superbus* Zone. Bergström (1971) defined the *A. superbus* Zone as a unit corresponding to the interval from the first appearance of *A. superbus* below up to the first appearance of *A. ordovicicus* above. In Bergström's scheme the *A. superbus* Zone follows directly the *A. tvaerensis* Zone. As we know now there is a considerable interval between the ranges of *A. tvaerensis* and *A. superbus* (see above). In the lower part of this interval *Amorphognathus* is very rare or missing, in its upper part *A. ventilatus* occurs. Also, according to Dzik (1999), *A. superbus* is not directly followed by *A. ordovicicus* but these taxa are evolutionally linked by a form called by Dzik *Amorphognathus* sp. n., and the strata between the *A. superbus* and *A. ordovicicus* zones can be considered as a separate unit. In the zonal scheme proposed by Viira (1974), the interval identified here as the *A. superbus* Zone corresponds to the main part of the *Ambalodus triangularis frognoeyensis* Zone.

In the Mehikoorma (421) section, *A. superbus* appears at 288.60–288.70 m (in the upper Oandu Stage) and is continuously present up to the level of 279.45–279.60 m in the upper Mõntu Formation (lower Nabala Stage) (Appendix 25). Conodonts are very rare above this interval, up to 270.80–270.90 m, and *Amorphognathus* is represented only by fragments not identifiable to species level.

The *Amorphognathus ordovicicus* Zone. Originally, the zone was defined as corresponding to the total range of *A. ordovicicus* (Bergström 1971). *Amorphognathus ordovicicus* was considered to be the last member in the evolutionary lineage of *Amorphognathus* and the lineage to become extinct at the Ordovician–Silurian boundary. However, as Dzik (1999) shows, *A. ordovicicus* is not the last species in the lineage but is followed by *A. duftonus* Rhodes. So far, *A. duftonus* has not been identified in Estonia, but it has turned out that the uppermost Ordovician conodont faunas of the Valga (10) and Ruhnu (500) cores (Porkuni Stage, its lower part – the Bernati Member – excluded) differ considerably from that of the underlying strata with *A. ordovicicus* (Männik 2001, 2003). In both sections the Ordovician strata above the Bernati Member yield very poor faunas, mainly dominated by *Noixodontus girardeauensis* (Satterfield). *Amorphognathus* is missing in these strata (in the Valga (10) core) or is extremely rare and represented by unidentifiable fragments (in the Ruhnu (500) core).

In the Mehikoorma (421) core section the lowermost specimens of *A. ordovicicus* are found at 270.00–270.10 m, in the upper part of the Saunja Formation. The species is almost continuously present in all samples above this level (Appendix 25). In the light of the present data it seems very probable that in Estonia *A. ordovicicus* appears already in the upper Nabala Stage and not in the lowermost Vormsi Stage as thought earlier (Männik 1992). However, a single M element of *Amorphognathus* identical to that of *Amorphognathus* sp. n. Dzik (1999) was found in the uppermost sample from the Saunja Formation (at 269.10–269.20 m). This may indicate that the lowermost part of the *A. ordovicicus* Zone in the studied section (corresponding to the upper Nabala Stage) is an equivalent to the *Amorphognathus* sp. n. interval *sensu* Dzik (1999).

Only the lower part of the *A. ordovicicus* Zone is represented in the Mehikoorma (421) core. Younger strata are missing due to the pre-Devonian erosion.

DISTRIBUTION OF ORDOVICIAN CHITINOZOANS

Seventy-two samples from the middle and upper Ordovician part of the Mehikoorma (421) core, from the interval of 283.4–340.9 m, were processed and studied for chitinozoans (Appendixes 4, 26). Due to secondary dolomitization organic-walled microfossils in the beds below 340.9 m and higher than 283.4 m were destroyed or poorly preserved (Appendix 1, sheets 9–13).

The work was carried out at the Institute of Geology at Tallinn University of Technology, and financially supported by the Estonian Science Foundation (Grant No. 5922).

The samples varied in size from 0.3 to 0.6 kg and were up to 5 cm in vertical range. All productive samples yielded a relatively rich assemblage of acid-resistant microfossils including well to excellently preserved chitinozoans. In total, 70 chitinozoan taxa were distinguished. Their distribution is given in Appendix 27, where the taxa with open nomenclature follow those found earlier from the Taga-Roostoja (25A), Valga (10) and Ruhnu (500) sections (see Nõlvak 1999a, appendix 6; 2001, appendix 8; 2003, appendix 23). Eleven biostratigraphically important chitinozoan zones were established in the investigated part of the Mehikoorma (421) section, following the zonation schemes introduced by Nõlvak & Grahn (1993) and revised by Nõlvak (1999b).

The lowermost part of the section is represented by the *Conochitina clavaherculi* Subzone of the *Laufeldochitina striata* Zone corresponding to most of Uhaku time. In the sample from 336.5–336.55 m the stratigraphically important graptolite *Gymnograptus linnarssoni* (Moberg) was found, which is present in the lower Uhaku beds in many sections (Männil 1976; Nõlvak 2001).

The next *Conochitina tuberculata* Subzone is represented by very rare specimens. The small number of specimens seems to be characteristic of South Estonian sections (e.g. Valga (10), Ruhnu (500); see Nõlvak 2001, 2003), contrary to the northern sections closer to the stratotype area (e.g. Taga-Roostoja 25A, Nõlvak 1999a; Männil 1986, fig. 2.2.1). More data are needed to determine if this shows also the difference in age. *Conochitina tuberculata* is absent in the Furudal Limestone on Öland (Sweden) (Grahn 1981), but some lithological similarities between this unit and the interval of 335.7–338.8 m in the Mehikoorma (421) core (the Taurupe Formation in the East Baltic sections, e.g. in the Valga (10) section) can be noticed.

The boundary between *Laufeldochitina striata* and *L. stentor* lies within the uppermost part of the Uhaku Stage in most of the investigated sections. This level coincides roughly with the lower boundary of so-called Erra beds of the uppermost Uhaku Stage in North Estonia (Männil 1966, 1986).

The succeeding *Eisenackitina rhenana* Subzone can be clearly followed. The species appears a little higher than *L. stentor* together with *Calpichitina complanata* and below *Conochitina* sp. 1 (Appendix 27). The last species occurs only in early Kukruse time, when the main kukersite-bearing (kerogenous) beds were formed in northeastern Estonia. Such an order of changes in the chitinozoan assemblages can be used to define the lower boundary of the Kukruse Stage in Estonian sections (e.g. in the Ruhnu (500) core, see Nõlvak 2003, p. 23) and in Sweden (Vandenbroucke 2004). In the Mehikoorma (421) core that boundary lies at a depth 330.7 m.

The most remarkable find among graptolites is that of *Nemagraptus gracilis* (Hall) in the uppermost beds of the Kukruse Stage at a depth of 319.5–319.6 m. According to our latest finds and identifications of *N. gracilis* only in the Kukruse Stage in the East Baltic sections (Nõlvak & Goldman 2004), at least these graptolite-bearing beds belong to the Upper Ordovician. The appearance level of this species marks the global lower boundary of the Upper Ordovician Series (Bergström *et al.* 2000), which coincides with the upper boundary of the global Darriwilian Stage and the lower boundary of time slice 5a by Webby *et al.* (2004, fig. 2.1).

In the Mehikoorma (421) section the base of the Haljala Stage (Idavere Substage) can be defined precisely between two close samples at a depth of 317.9 m. Both *Laufeldochitina stentor* and *Eisenackitina rhenana* disappear in the lower sample (317.95–318.05 m) and *Lagenochitina dalbyensis* appears in the upper sample (317.70–317.80 m; Appendix 27). *Lagenochitina dalbyensis*, together with the *Belonechitina hirsuta* Zone, occurs in the Idavere Substage. However, the subdivision of the Haljala Stage into substages in terms of chitinozoan zonation is still complicated in most of the studied sections.

The base of the Keila Stage is marked in Estonian sections by the widely distributed Kinnekulle K-bentonite bed (Hints & Nõlvak 1999). In the Mehikoorma (421) core this bed was earlier described at a depth of 306.0 m (thickness 1 cm; Kajak *et al.* 1974). In the present study (see Kiipli & Kallaste in this volume) the Kinnekulle bed is identified at 306.7 m (thickness 2 cm). The well-known very brief *Angochitina multiplex* Subzone together with *Hercochitina* sp. has not been found above this probable new level (confirmed by XRD data), so the correlation in terms of chitinozoans remains open, as the number of both species is often very low.

Continuing systematic problems with the key species of the *Fungochitina fungiformis* Zone complicate the use of this zone as noticed already in the Valga (10) section (Nõlvak 2001, p. 9). Nõlvak & Grahn (1993) defined this zone as a total range of *F. fungiformis* and they did not distinguish *F. fungiformis fungiformis* and *F. fungiformis spinifera*, originally established as subspecies by Eisenack (1962). However, in the revision of *Fungochitina* by Paris *et al.* (1999) these subspecies were elevated to species rank, belonging to different genera: glabrous forms of *fungiformis* can be included into *Saharochitina*. Moreover, according to the new data from the Mehikoorma (421) and Viljandi cores (Kaljo *et al.* 2004, fig.4), *Saharochitina fungiformis* appears earlier than typical spiny *F. spinifera*. At that time we had detailed data only from the North Estonian Confacies Belt, showing a clear change in lithology and the well-known gap in the boundary beds of the Keila and Oandu stages. A clear change among chitinozoans and acritarchs in this part of the succession has been interpreted as an extinction event (Kaljo *et al.* 1996). In the Mehikoorma (421) section chitinozoan assemblages change more gradually than in the Rapla section (Kaljo *et al.* 1996) and no distinct levels of change can be defined. This shows that the North Estonian gap is fulfilled with new layers, which belong to the uppermost Keila and/or lowermost Oandu stages. At the moment the beginning of Oandu time cannot be reliably fixed on the basis of chitinozoan data. Therefore here the end of mass occurrence of the acritarch *Leiosphaeridia baltica* and specific curved population of *Euconochitina primitiva* is used to mark the possible level of the base of the Oandu Stage below a depth of 296.3–296.4 m. Such a change can be followed also in the Valga (10) core (Nõlvak 2001, appendix 8). The upper portion of the Oandu beds, above the *Ancyrochitina* sp. n. 1 Subzone and below the Rägavere Formation, seems to be younger than in the North Estonian sections. This can be proved by the *spinifera* Zone (= earlier *fungiformis*), which is known to appear in the beds belonging to the Tõrremägi Member of the Rägavere Formation (Nõlvak & Grahn 1993).

Thus, it seems reasonable to rename the *Fungochitina fungiformis* Zone (by Nõlvak & Grahn 1993; Nõlvak 1999) as the *Fungochitina spinifera* Zone (used already here in Appendix 27). It should also be noted that in the general correlation chart of Webby *et al.* (2004, fig. 2.1.) the name of this zone is absent in the corresponding time slices (5c and 5d) as a misprint in that work.

The lower boundary of the Rakvere Stage coincides in the Mehikoorma (421) core with the appearance of *Cyathochitina angusta* (Appendix 27).

In general, the distribution and diversity of chitinozoans in the Uhaku, Kukruse, Haljala and Keila stages and also in the very condensed Rakvere Stage in the Mehikoorma (421) section is in relatively good accordance with earlier data. The beds of the Oandu Stage, however, need additional study because of the risk of miscorrelation between the sections from North Estonian and Central Confacies Belts.

DISTRIBUTION OF ORDOVICIAN MACROFOSSILS

A total of 100 samples from the interval of 251.5–371.4 m in the Mehikoorma (421) core were analysed to identify Ordovician macrofossils (Fig. 4, Appendix 28). The samples were collected during the initial study of the core by Kalju Kajak in the 1970s and are presently stored at the Institute of Geology at Tallinn University of Technology. Those of the samples containing badly preserved fragments of fossils, which could be identified only to phylum level, are not preserved. Two intervals (262.7–279.6 and 336.0–354.5 m) were not sampled. The uppermost samples from the Vormsi and Pirgu stages (11 samples; see Appendix 28) up to 251.5 m, containing mainly sparse small lingulate brachiopods, are not represented in Fig. 4. Part of the material needs further taxonomic revision and comparison with the materials from other sections located in the limits of the same facies belt. In addition to macrofossils, some microfossils were identified in the studied rock samples.

The lowermost sampled interval (354.5–370.8 m), represented by mainly brownish-red and mottled dolostones (see Appendix 1, sheet 13), contains sparse skeletal fragments of trilobites and cephalopod shells. Some small lingulate ("*Conotreta*") brachiopods occur in the interval included into the Volkhov Stage. Two samples of grey-coloured rocks (from 360.9 and 356.4 m) between brownish-red and mottled rocks yielded *Leiosphaeridia*-like acritarch spores. Supposedly, these spores represent an example of exceptional preservation as spores are usually destroyed in brownish-red rocks.

Cephalopod shell fragments in samples from 354.5 and 354.8 m supposedly belong to *Cyclendoceras cancellatum* (Eichwald), usually identified in the Kunda Stage. Frequent occurrence of cephalopods in the lowermost Ordovician brownish-red carbonate rocks is characteristic of the Baltic Basin, but relatively low abundance of trilobites possibly differentiates the fauna of the Mehikoorma area from that of the central part of the basin where trilobites are more common (see Gailite & Ulst 1975).

The interval of 289.0–336.0 m, represented by more or less argillaceous limestones and dolomitic marlstones and corresponding to the stratigraphical interval from the Uhaku Stage to the Oandu Stage, contains a diverse association of fossils (Fig. 4).

The bioclast-bearing limestones of the Kõrgekallas Formation (interval 331.2–335.7 m) yielded some trilobites and small brachiopods (dalmanellids and cremnorthids). Small shells of the brachiopod *Cremnorthis*, characteristic of the Uhaku Stage in northern Estonia, were found in two samples.

The sample from a depth of 332.5 m deserves special attention. Apart from trilobites, it contains the cystoid *Echinosphaerites* and graptolite fragments. The argillaceous limestone of that sample includes small white oolith-like grains. White phosphate (francolite) ooliths appear in some SE Estonian sections, for example in the Laeva (297) and Kaagvere (1) drill

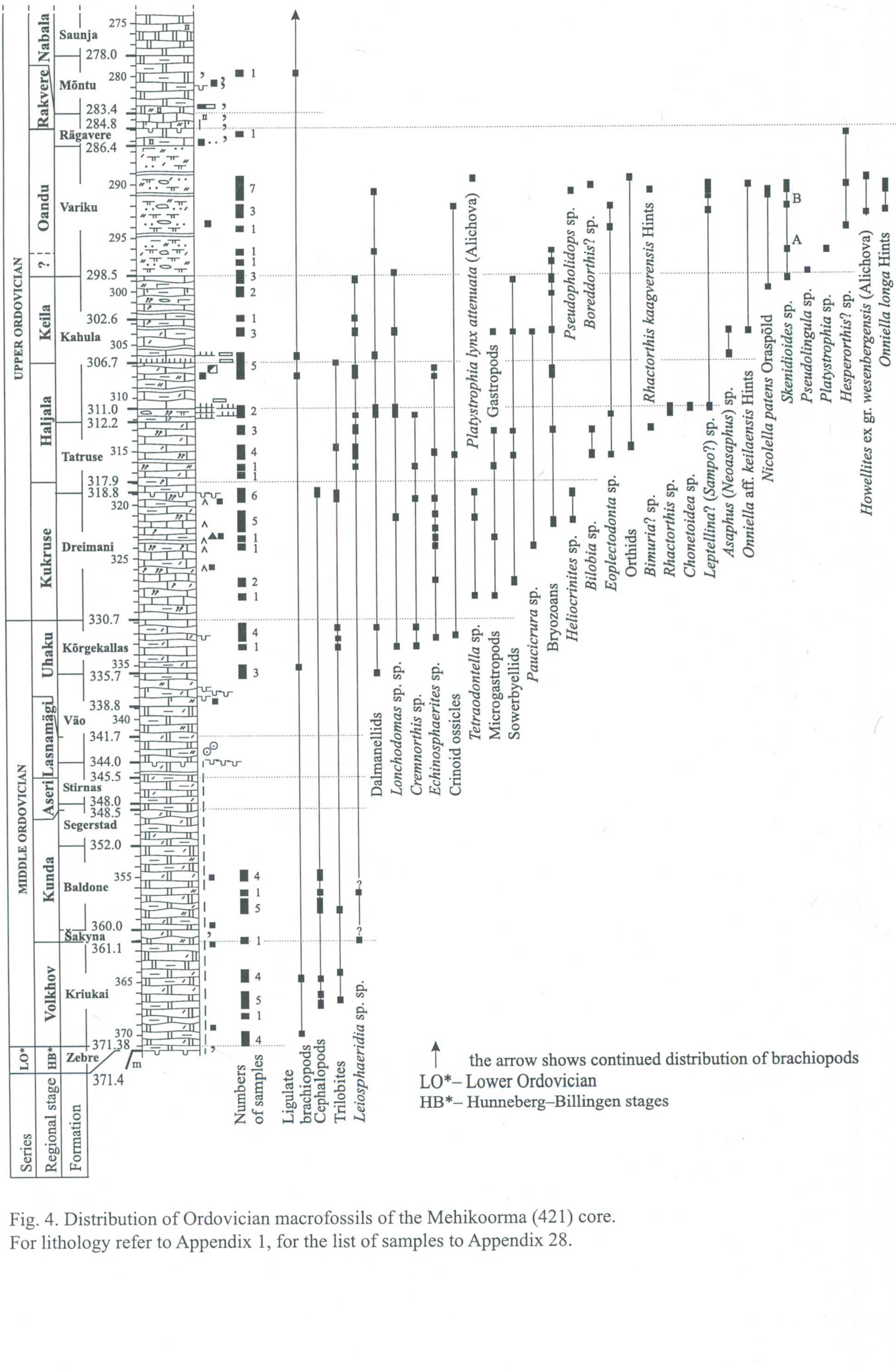

Fig. 4. Distribution of Ordovician macrofossils of the Mehikoorma (421) core.
For lithology refer to Appendix 1, for the list of samples to Appendix 28.

cores (Põlma 1982), in the boundary beds between the Lasnamägi and Uhaku stages and occur sporadically up to the topmost Kukruse Stage.

Echinosphaerites sp. is the most frequent fossil in the interval of 318.65–328.5 m. These cystoids are common in the Kukruse Stage in many sections of southeastern Estonia (Karula, Otepää, Kaagvere and Laeva cores; see Männil 1966, figs 12–15). The lowermost part of this interval yielded phosphatic (francolitic?) microgastropods (samples at 323.1 and 328.5 m), which were identified also in the thin section from 324.7 m (Appendix 2, T-32). Microgastropods are found in some sections in southern Estonia, but their stratigraphic and spatial distribution needs additional study. In the Mehikoorma (421) section the uppermost microgastropods occur at 313.0 m (Haljala Stage). The first plectambonitid (*Tetraodontella*?) and sowerbyellid brachiopods appear approximately at the same level with microgastropods.

Some new fossils, larger fragments of bryozoan colonies and cystoid *Heliocrinites* sp. appear in the uppermost part of the Kukruse Stage (317.9–321.3 m). Bryozoans have limited importance in the composition of offshore faunas and, according to drill core data, are more common only in the Keila–Oandu interval (Hints & Põlma 1981; Hints 1990).

Frequent acritarchs *Leiosphaeridia* were identified above 316.4 m in the samples with brachiopods *Bilobia* sp., *Eoplectodonta*? sp. and *Bimuria*? sp., distributed in the lower Haljala Stage (Idavere Substage; Appendix 1, sheets 11, 12; interval 311.0–317.9 m). The level of frequent appearance of *Leiosphaeridia* falls in northern Estonia very close to the lower boundary of the Haljala Stage. The thin bed of massive marlstone in the sample from 315.4 m is possibly a K-bentonite, but its volcanic origin is not proved by other data. In some sections (for example, in the Laeva core; Männil 1966, fig. 17) three K-bentonites are recognized in the uppermost Idavere Stage (=Substage).

Two samples, from 311.44 and 310.9 m, contain a similar and diverse association of fossils. Some brachiopods belong possibly to the genera *Leptellina* and *Rhactorthis*, earlier identified in southeastern Estonia in Keila Age rocks. However, the latter genus has been identified in the Haljala Stage in the westernmost Latvian section (Engure-4, unpublished data) together with cremnorthids and representatives of *Onniella* and *Chonetoidea*. The trilobite *Lonchodomas*? and graptolite fragments are also found in these samples.

The about 3 m thick upper part of the Haljala Stage (below the boundary K-bentonite bed at 306.7 m) shows the last occurrence of *Echinosphaerites*, found in association with fragments of bryozoan colonies and trilobites.

The trilobite *Asaphus* (*Neoasaphus*) sp. is found in some samples (interval 303.6–305.9 m) above the K-bentonite at a depth of 306.0 m, together with *Onniella* aff. *keilaensis* Hints, sowerbyellids and gastropods. This association suggests both the late Haljala and early Keila Age. In southeast Estonia the range of *Asaphus* (*Neoasaphus*) *ludibundus* Törnquist falls into the Kukruse and Haljala stages (Männil 1966), although few finds come from the lowermost Keila Stage.

Renovation of fauna takes place roughly at 299.0 m. Samples close to that level sometimes contain rare acritarchs *Leiosphaeridia*. According to micropalaeontological data (Fig. 4), the uppermost frequent occurrence of *Leiospaeridia* is at 298.7 m (at 296.3 m after the data by Jaak Nõlvak, see Appendix 27). Above that depth the brachiopods constitute the predominating group among macrofossils, however, they are associated with small spherical colonies of bryozoans up to the 296.2 m level. The brachiopods *Skenidioides*, *Nicolella*, *Platystrophia*, *Onniella*, *Sowerbyella*? and *Leptellina*? (*Sampo*?), identified in the Mehikoorma (421) core, are found in the argillaceous and silt-containing limestones in the Keila–Oandu interval of several other sections (see Männil 1966). The Ordovician *Skenidioides* is known mainly in the offshore facies (Harper 1986), in the composition of Silurian communities *Skenidioides* is characteristic of the fourth or fifth benthic community (Kaljo & Rubel 1982). The brachiopods identified in the Mehikoorma (421) core represent a specific association different from that of northern Estonia, although there exist some common species: *Nicolella patens* Oraspõld, *Platystrophia lynx attenuata* (Alichova). Also, this association differs from that of the central part of the Baltic Basin, where the Blidene and Priekule formations are distributed.

The sample at a depth of 294.0 m consists of burrowed calcitic marlstones with some content of dark grey (organic) material. Supposedly these marlstones represent the periphery of some part of black shales of the Mossen Member, known in areas SW of the Mehikoorma (421) drill hole (in the Karula core; see Männil 1966). If that is the case, the lower part of the Variku Formation may correspond to some part of the Blidene Formation in the Central East Baltic, where the black shales of the Mossen Formation overlie the Blidene Formation. In the interval of 289.9–292.8 m the brachiopods *Onniella* cf. *longa* Hints, *Skenidioides* sp. B, *Leptellina* cf. *indentata* (Spjeldnaes), *Rhactorthis kaagverensis* Hints, *Howellites* ex gr. *wesenbergensis* Alichova, *Boreadorthis*? sp. and some others make up a new association. The succession of brachiopods above the level of about 302.0 m in the Mehikoorma (421) core is similar to those in the Kaagvere (1) and Otepää (2) cores (unpublished data). Judging from the practise in northern Estonia, the beds with *Howellites wesenbergensis* Alichova supposedly belong to the Oandu Stage.

The uppermost samples (up to 251.5 m) contain sparse macrofossils, mainly small lingulate brachiopods found in the lower parts of the Vormsi and Pirgu stages.

ORDOVICIAN CARBON ISOTOPES

A total of 112 samples from the Middle and Upper Ordovician rocks of the Mehikoorma (421) core (Appendix 1, sheets 9–13) were analysed for stable isotopes. The entire Ordovician sequence was sampled at more or less regular intervals of 1 m, not depending on the possibility of finding any bioclasts (Martma 2003). For isotope analysis about 1 g of rock material was taken from conodont and chitinozoan samples (82), and geochemical and petrophysical samples (3), additionally 27 samples were collected (Appendix 29). Combined sampling for microfossils and isotopes ensures a more reliable correlation of different sections and data sets (see also Männik and Nõlvak in this volume).

Our study is based on the whole-rock sampling method with consideration of the stratigraphic context (lithology, unit thickness, positions of unit boundaries). The quality of the carbon isotope data based on whole-rock analyses has been discussed for several years (Brenchley *et al.* 1994; Kaljo *et al.* 1997; Martma 2003). Brenchley *et al.* (2003) investigated in detail the reliability of isotope signals in the Late Ordovician rocks of Estonia and note that major changes in isotope values reflect primary composition. The comparison of the Baltic latest Ordovician and Silurian whole-rock (Kaljo *et al.* 1998, 2001) and brachiopod shell isotope data (Marshall *et al.* 1997; Heath *et al.* 1998; Brenchley *et al.* 2003) shows only slight difference in $\delta^{13}C$ values but great similarity of the corresponding curves. This is confirmed also by Silurian data of Gotland (Samtleben *et al.* 1996). The oxygen isotope ratios are more sensitive to diagenesis (Marshall 1992; Saltzman 2002), and the Baltic carbonate rocks are mostly highly variable mixtures of calcite and dolomite, which have different oxygen isotope fractionation factors. Therefore data from whole-rock analysis are not trustworthy and we do not use Mehikoorma (421) oxygen isotope data for geological conclusions.

The methodology of carbon isotope analysis used in the isotope palaeoclimatology laboratory of the Institute of Geology at Tallinn University of Technology is explained in detail in Kaljo *et al.* (1997, 1998). Here only some comments are made on essential details of the study. The whole-rock samples were powdered to a <10 μm grain size, 30 mg of powder was reacted with 100% phosphoric acid at 100 °C for 15 min and analysed with a Finnigan MAT "Delta E" mass spectrometer. The results are presented in the usual δ notation, as per mil deviation from the VPDB standard. The reproducibility of the results is better than 0.1‰.

A full set of analytical data on the carbon isotopes obtained from bulk samples of the Mehikoorma (421) drill core will be published in a forthcoming paper (Kaljo *et al.* submitted).

The application of carbon isotopes as a tool for stratigraphic correlation and dating of rock sequences is in principle a simple method. The reliability of the results depends on how detailed and complete is the database available for comparison. A more or less complete carbon isotope trend for the Middle and Late Ordovician of Baltica and Laurentia has been ascertained on the basis of studies by Ainsaar *et al.* (1999, 2004a, 2004b), Kaljo *et al.* (1999, 2001, 2004) and Meidla *et al.* (2004). Considering the earlier data and analyses included in Fig. 5, the following main carbon isotopic events and specific intervals of the $\delta^{13}C$ temporal variation through the Ordovician of the Mehikoorma (421) core could be listed:

(1) The Aseri (Mid-Darriwilian) isotopic event or positive excursion. The carbon isotope trend is characterized by a relatively rapid rise in $\delta^{13}C$ values from 0.5‰ in the Baldone Formation (Kunda Stage) through the Segerstad Formation to a peak value of 1.7‰ in the Stirnas Formation of the Aseri Stage. The falling limb of the excursion is located in the Väo Formation of the Lasnamägi and Uhaku stages. The Aseri shift was first identified by Ainsaar *et al.* (2004b) in the rocks of the Segerstad Formation (correlates with the Aseri Stage after Nõlvak 1997) of the Jurmala (R–1) and Ruhnu (500) sections. New data from the Mehikoorma (421) core show that the excursion is considerably wider, beginning in late Kunda and ending in Lasnamägi times.

(2) The mid-Caradoc double-peaked excursion with a maximum $\delta^{13}C$ value of 1.8‰ at the boundary of the Kahula and Variku formations (Keila and Oandu stages), followed by a lower shift (1.1‰) higher in the Variku Formation. The Mehikoorma (421) data support the correlation suggested by Kaljo *et al.* (2004), based on the data from the Vasalemma and Saku area.

(3) The 1st late Caradoc isotopic event (peak value reaching 1.6‰) is confined to the thin Rägavere

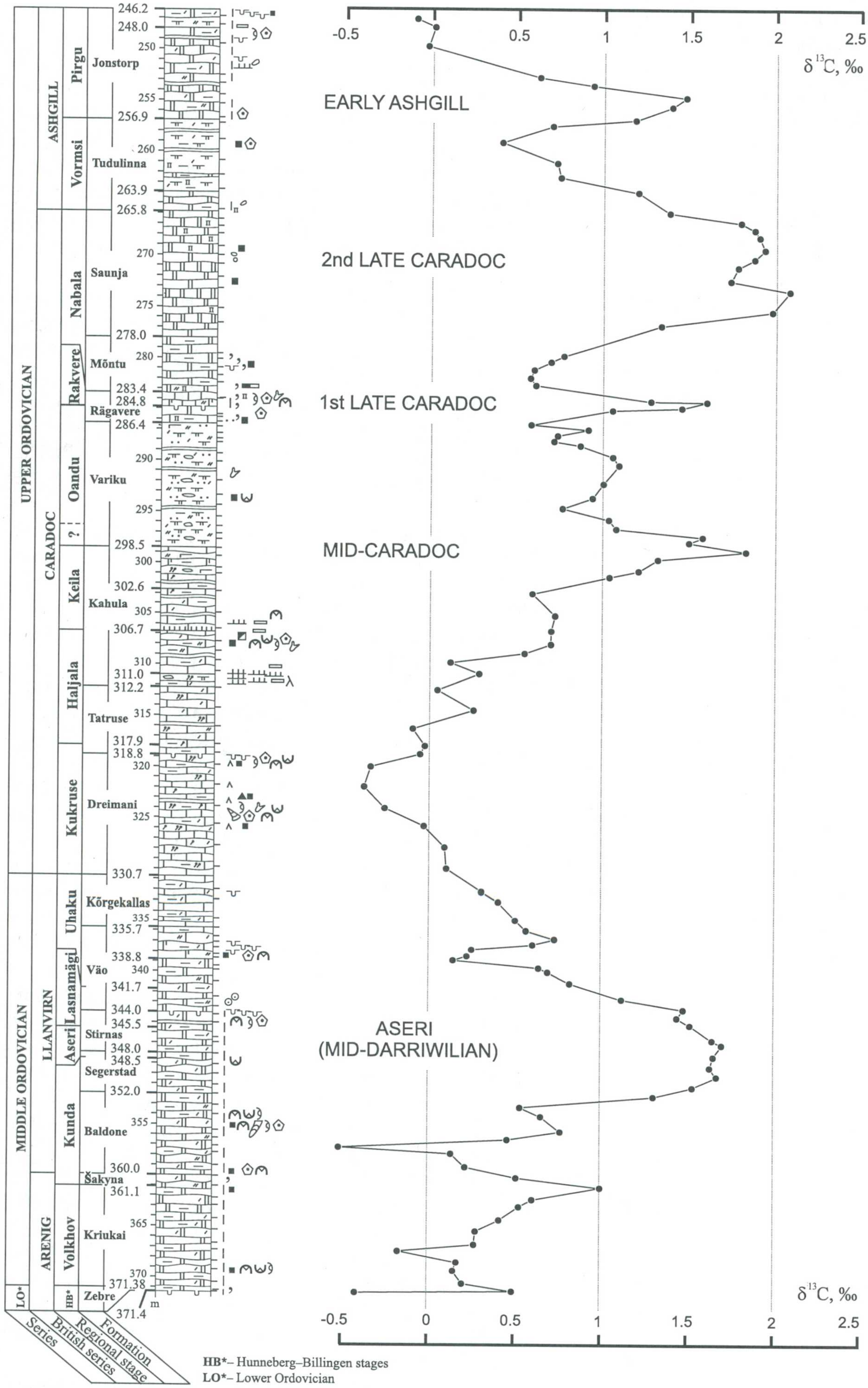

Fig. 5. The Ordovician bulk carbonate carbon stable isotope profile of the Mehikoorma (421) core. Refer to Appendix 1 for lithology and Appendix 29 for sample depths. Sampling points are marked on the right side of the column.

Formation (Rakvere Stage) in the Mehikoorma (421) core. The peak occurs just above a discontinuity surface, suggesting that the lower part of the Rakvere Stage is represented in that core (Kaljo *et al.* 1999, 2004).

(4) The 2nd late Caradoc isotopic event with $\delta^{13}C$ peak values of 2.1‰ is a rather typical excursion in the upper part of the Saunja Formation (Nabala Stage). The Mehikoorma (421) data confirm an earlier conclusion (Kaljo *et al.* 2004) that a positive shift in the lower part of the stage in the Rapla drill core (Kaljo *et al.* 1999) is a local anomaly.

(5) The early Ashgill isotopic event with $\delta^{13}C$ peak values of 1.5‰ is rather typical of the lowest part of the Jonstorp Formation (Pirgu Stage). The excursion is a good marker for carbon isotope-based correlation of the lower Pirgu sequences.

(6) Higher carbon isotope excursions like the Mid-Ashgill and Hirnantian ones are missing in the section due to a gap in the succession from the lowermost Pirgu Stage up to the lower Devonian.

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CHARACTERISTICS OF ORDOVICIAN VOLCANIC ASH BEDS

The Mehikoorma (421) section is located in the marginal area of the known occurrence of volcanic ashbeds (Fig. 6). The Kinnekulle K-bentonite bed, which in the West Estonian islands reaches a thickness of 50–70 cm, has been identified in the Mehikoorma section only as a 2 cm thick interbed at a depth of 306.7 m.

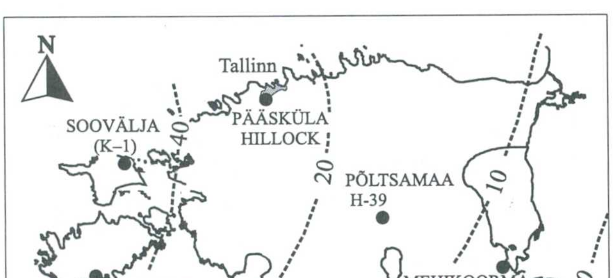

Fig. 6. Distribution of the Kinnekulle K-bentonite (isopachs given after Vingisaar 1972 and Bergström *et al.* 1995) and location of the sections mentioned in the text.

Ten argillaceous beds of suspected volcanogenic origin were sampled and analysed in the Mehikoorma (421) core. Samples were taken mostly from 0.5–15 cm thick interbeds. The chemical composition and trace elements were analysed by the XRF method (Appendix 30). Major components (in wt%) of the Kinnekulle K-bentonite bed (SiO_2 – 54.50; TiO_2 – 0.50; Al_2O_3 – 17.70; Fe_2O_3 – 3.78; MnO – 0.035; CaO – 4.62; Na_2O – 0.11; K_2O – 8.56; P_2O_5 – 0.053; LOI – 7.40) were analysed at the Laboratory of All-Union Geological Institute (VSEGEI) in St. Petersburg.

The Na content of K–Na sanidine was analysed by XRD (Table 1) as in our earlier studies (Kiipli & Kallaste 2002a, 2003). From the separated 0.04–0.1 mm fraction the $20\bar{1}$ reflection of sanidine was measured with maximum accuracy. An angular range from 23.5 to 26.0 $°2\theta$ was scanned with a step size of 0.01 $°2\theta$; the measuring time was 15 s per point. The content of $NaAlSi_3O_8$ (in mol%) in K–Na sanidine was calculated according to Orville (1967), who established that the position of the $20\bar{1}$ reflection depends almost linearly on the composition of the sanidine solid solution. The precision of the analysis of the K–Na sanidine composition is ±1% in favourable cases (low intensity of authigenic feldspar reflection, no kaolinite, high intensity of the reflection of interest), and ±2% in less favourable cases. Correlation of separate beds over a wide area is based on the properties of magmatic K–Na sanidine and trace element distribution.

XRD and XRF analyses revealed mixed volcanogenic-terrigenous material in almost all samples. High CaO contents (Fig. 7) in these samples give evidence of the occurrence of biogenic and mechanically transported material. Pure K-bentonite was found only in a 1 cm thick bed at a depth of 311.4 m (Table 1). In most cases depositional environments and diagenetic processes caused the mixing of materials from different sources. Nevertheless, pyroclastic minerals have usually a larger grain size than terrigenous material and can be separated relatively easily.

Authigenic K-feldspar, often seriously retarding separation of pyroclastic K–Na sanidine in other regions of Estonia, can be easily disintegrated in the Mehikoorma (421) core samples after dissolution of carbonate. This compositional property enables analysis of pyroclastic sanidine even in samples with a high content of authigenic feldspar. The main obstacle for the correlation of Caradocian volcanic beds arises from the lack or low concentration of K–Na sanidine in most bentonites (Kiipli & Kallaste 2002b).

Table 1. Correlation of volcanic ash beds of the Soovälja (K–1), Mehikoorma (421) and Kuressaare (K–3) cores

Complexes by Bergström *et al*. (1995)	Regional stage	Soovälja (K–1)				Mehikoorma (421)			Kuressaare (K–3)		
		Sample depth (m)	No. of bed on core box	Biotite: – absent, (+) rare, + present, ++ abundant	$NaAlSi_3O_8$ in sanidine* (mol%), shape of the $20\bar{1}$ reflection	Sample depth (m)	Biotite: – absent, (+) rare, + present, ++ abundant	$NaAlSi_3O_8$ in sanidine* (mol%), shape of the $20\bar{1}$ reflection	Sample depth (m)	Biotite: – absent, (+) rare, + present, ++ abundant	$NaAlSi_3O_8$ in sanidine* (mol%), shape of the $20\bar{1}$ reflection
	Pirgu					251.80	+	37.6			
Grimstorp	**Keila**					**306.00**	++	–	**367.70**	++	**Weak reflection**
Kinnekulle	**Keila**	**173.10**		+	**24.5****	**306.70**	++	**25.2**	**368.50**		
Sinsen	**Haljala**								**369.20**	+	**21.6**
	Haljala	177.22	1	(+)	Weak reflection						
	Haljala	**177.45**	**2**	+	**22.1****	**310.90**	+	**22.4**	**370.50**	+	**20.8**
	Haljala	**177.62**	**3**	(+)	**Wide reflection**	**311.00**	(+)	**Wide reflection**	**370.70**	(+)	**Wide reflection**
	Haljala	178.00	4	–	Wide reflection						
	Haljala	178.83	5	not examined	–	311.15	+	Weak reflection			
	Haljala	178.90	6	–	Wide reflection						
	Haljala	179.10	7	–	Wide reflection						
	Haljala	180.32	8	–	Wide reflection	311.40	(+)	Wide reflection	372.20	–	Weak reflection
	Haljala	180.90	9	–	–						
	Haljala	181.02	10	–	Wide reflection						
Grefsen	Haljala	181.40	11	+	Wide reflection						
	Haljala	181.70	12	(+)	Wide reflection						
	Haljala	181.90	13	–	Weak reflection						
	Haljala	182.60	14	–	–	311.44	not examined	–			
	Haljala	185.32	15	–	Wide reflection						
	Haljala					311.70	–	Wide reflection			
	Haljala	186.40	16	–	Wide reflection						
	Haljala	197.21	17	(+)	Wide reflection	311.90	+	Weak reflection			
	Haljala	198.25	18	+	Wide reflection						
	Haljala	203.03	19	++	–						
	Haljala	255.98	20	–	–						

* K–Na sanidine $20\bar{1}$ reflection was studied using the two-component model, only the main component is included in the table.
** in Kiipli & Kallaste (2002b) erroneously average values of the bed were published.
Well established correlations are in bold type, possible, but not proved correlations are in ordinary type. Samples collected by Tarmo Kiipli and Toivo Kallaste (Geological Survey of Estonia), analysed at the Laboratory of the Institute of Geology at Tallinn University of Technology.

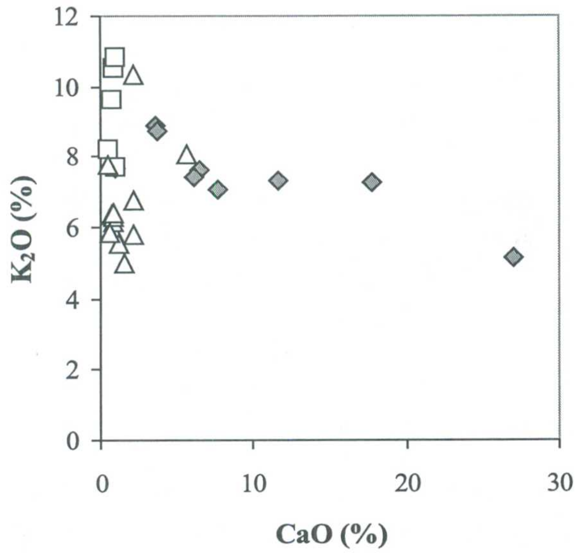

Fig. 7. CaO content versus K_2O content in studied samples.
Triangles – Soovälja (K–1), dark rhombs – Mehikoorma (421), quadrangles – Kuressaare (K–3) bentonites. High K_2O content indicates volcanic origin (typical content of terrigenous rocks is 4–5%). In the Mehikoorma (421) bentonites are rich in CaO.

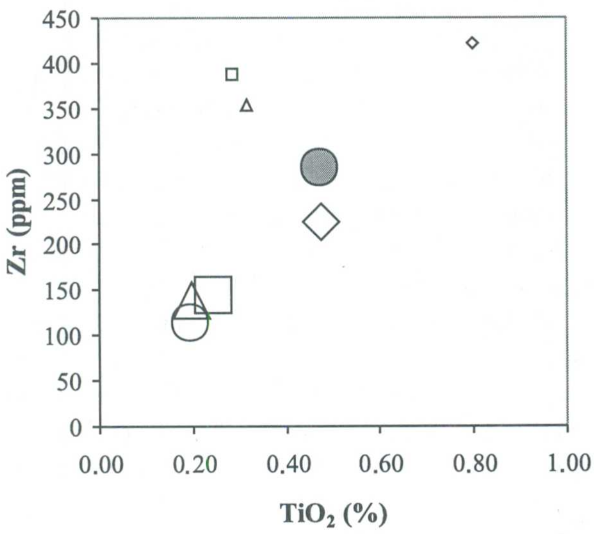

Fig. 8. TiO_2 content versus Zr content in two correlated K-bentonites.
Triangles – Soovälja (K–1), rhombs – Mehikoorma (421), quadrangles – Kuressaare (K–3), filled circle –Valga (10) cores (Kiipli & Kallaste 2001) and empty circle – Pääsküla Hillock (Hints *et al.* 1997). Large symbols – Kinnekulle K-bentonite bed. Small symbols – Grefsen K-bentonite (Soovälja (K–1) at 177.62 m, Mehikoorma (421) at 311.0 m and Kuressaare (K–3) at 370.70 m). Higher TiO_2 concentrations in Mehikoorma (421) K-bentonites are due to large amounts of terrigenous admixture. Zr content is higher in the K-bentonite from the upper part of the Grefsen complex.

Use of immobile trace elements for correlations is strongly hindered by the presence of terrigenous material in volcanogenic bentonites. For example, TiO_2 and Ba contents (Fig. 8, Appendix 30, see St. Petersburg analyses above) are much higher in the Mehikoorma (421) core bentonites than in the correlated beds of the Kuressaare (K–3) and Soovälja (K–1) cores (see Kiipli & Kallaste 2002b). This is caused by higher concentrations of Ti and Ba in terrigenous material compared to the volcanogenic material of many ash beds. Therefore concentrations of these elements should be used with some reservations. It seems that Zr and Y concentrations in the Mehikoorma bentonites (Appendix 30) are less influenced by terrigenous material, possibly due to their close concentrations in terrigenous and volcanogenic material.

A sample from the Pirgu Stage (depth 251.8 m, Appendix 1, sheet 9) was taken from a polished bedding plane, which points to an unusually soft and thin interbed lost during drilling. Provisional study showed dolomite as the main component of the sample, but the residue obtained after dissolution of carbonate material contained much authigenic K-feldspar and some biotite flakes. K–Na sanidine measurement revealed a weak, but enough distinctive $20\bar{1}$ reflex, from which we calculated the value of 37.6 mol% sodium feldspar in sanidine solid solution. Such a K–Na-feldspar composition correlates with that of the bentonite from the lower part of the Pirgu Stage, where sanidine contains 37–39 mol% Na (e.g. Põltsamaa H–39 core, depth 177.1 m; Kiipli *et al.* 2004).

Similarly to the Soovälja (K–1) section (Kiipli & Kallaste 2002b), magmatic cycles can be recognized on the basis of the Zr/TiO_2 ratio in Caradocian volcanic beds of the Mehikoorma (421) section (Appendix 1, sheet 11; Fig. 9). In the Mehikoorma (421) core we can probably distinguish the two upper cycles identified in the Soovälja (K–1) core.

The upper cycle includes some beds from the upper part of the Grefsen complex (depths 310.9 and 311.15 m), and the Kinnekulle (306.7 m) and Grimstorp (306.0 m) beds. Bentonites in this magmatic cycle are characterized by the presence or even high content of biotite (Table 1) and often also by magmatic K–Na sanidine having a specific composition. The content of sanidine containing 22.4–25.2 mol%

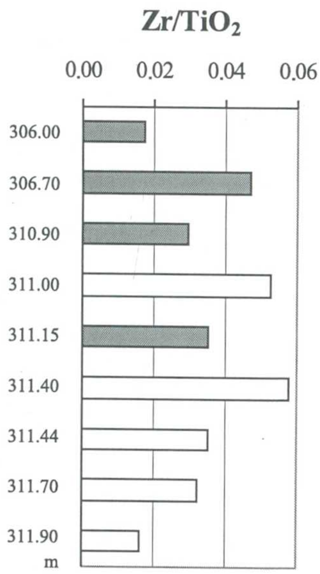

Fig. 9. Zr/TiO_2 in Mehikoorma (421) K-bentonites. Empty bars – lower cycle, filled bars – upper cycle. The upper cycle begins earlier than the lower ends, which probably indicates the formation of a new volcano.

NaAl Si_3O_8 varies largely in Caradocian volcanic ash beds, being high in Kinnekulle (depth 306.7 m) and in one of the upper Grefsen bentonites (depth 310.90 m) and only rarely detectable in other beds (Table 1). The Grimstorp bentonite can be recognized by its remarkably high biotite content and by large pyroclastic biotite and quartz grains in it. The Sinsen bentonite (not registered in the Mehikoorma (421) core) from the Kuressaare (K–3) core (depth 369.2 m; 21.6 mol% NaAl Si_3O_8) is very similar to the Upper Grefsen bentonite from the Soovälja (K–1) core (depth 177.45 m; No. 2 on the core box, Table 1) and can be identified with certainty only if both beds are present.

Bentonites of the lower magmatic cycle (medium cycle in the Soovälja (K–1) core) in the Mehikoorma (421) core are characterized by the absence or low content of biotite and a wide K–Na sanidine reflection $20\bar{1}$ (Table 1), which allows no reliable correlations on the basis of XRD patterns. In this cycle we can observe a well-developed magma differentiation cycle beginning with TiO_2-rich ash beds and ending with Zr-rich beds (see Appendix 30). An interesting new aspect is that the upper cycle begins before (volcanogenic bed at 311.15 m) the lower cycle ends (depth 311.0 m). This phenomenon needs to be checked in other sections. A preliminary conclusion is that these two cycles do not originate from the evolution of the same magma chamber inside the same volcano. Thus we can suggest the appearance of a new volcanic centre before the activity ended in the old centre.

The high content of TiO_2 (1.47%; Appendix 30) in the sample from 311.7 m indicates possible correlation with the lower part of the medium magmatic cycle in the Soovälja (K–1) section (interval 181.67–186.4 m; see Kiipli & Kallaste 2002b), but the trace element spectra of the sample do not fit with any of these beds. Thus we suppose we found a new bed, not yet registered in the Soovälja (K–1) core.

DISTRIBUTION OF FURONGIAN (UPPER CAMBRIAN) ORGANIC-WALLED MICROFOSSILS

Furongian (Upper Cambrian) sandstones and claystones, corresponding to the Petseri Formation (interval 371.4–379.4 m; Appendix 1, sheet 14), are known only in core sections of southeastern Estonia, northeastern Latvia and Pskov Region of Russia. On the basis of lithology, the complete Petseri Formation is divided into three parts. The upper and lower parts are represented by weakly cemented quartzose sandstones and coarse-grained siltstones containing rare fragments of inarticulate brachiopods and some glauconite. The middle part comprises claystones and siltstones with acritarchs and shell fragments of lingulates. The upper part of the formation is lacking in several sections due to postsedimentary denudation (Mens *et al.* 1987).

In the Mehikoorma (421) core the Petseri Formation is represented by its lower and middle parts (Appendix 1, sheet 14). Denudation of the upper part is marked by a sharp wavy yellow goethitized discontinuity surface. The lower part (374.3–379.4 m) consists of fine- and medium-grained quartz sandstones with numerous valves and fragments of lingulates. The middle part (371.4–374.3 m) of formation is composed of silty acritarch-bearing claystones. For the palynological study six samples (from 371.5, 372.3, 372.8, 374.0, 376.0 and 378.7 m) were collected from the section.

Palynologically productive samples are from the middle part of the formation (371.4–374.3 m), except for the two samples collected from the uppermost 0.9 m, which are barren (371.5 and 372.3 m) due to the weathering of rocks. The two lowermost samples from silty claystones (372.8 and 374.0 m) yielded well-preserved relatively diverse acritarch assemblages.

The acritarch assemblages of both samples are similar, containing the following species: ***Caldariola glabra*** (Martin) Molyneux, ***Cymatiogalea* aff. *cuvillieri*** (Deunff) Deunff, ***C.* aff. *multarea*** Deunff, ***C. velifera*** (Downie) Martin, ***C. virgulta*** Martin, **?*Cymatiosphaera* sp.**, ***Leiofusa stoumonensis*** Vanguestaine, ***Leiosphaeridia* sp.**, *Leiosphaeridia* div. sp., ***Micrhystridium* sp. 1**, ***Micrhystridium* sp. 2**, **?*Ovulum* sp.**, *Pirea orbicularis* Volkova, ***Poikilofusa* sp.**, ***Stelliferidium cortinulum*** (Deunff) Deunff, Gorka & Rauscher, ***Stelliferidium* sp.**,

Timofeevia lancarae (Cramer & Diez) Vanguestaine, ***T. phosphoritica*** Vanguestaine, ***Timofeevia* sp. 1**, ***Veryhachium incus*** Paalits, ***V. setuensis*** Paalits, *Vulcanisphaera* aff. *africana* Deunff, ***V. turbata*** Martin, **?*Vulcanisphaera* sp. 1**, **Gen. *et* sp. indet. 1** and **Gen. *et* sp. indet. 2**. Acritarchs in the bold font are illustrated in appendix 30.

The first results of the palynological study of the Petseri Formation in the Mehikoorma (421) core are briefly described in Volkova *et al.* (1981). The above acritarch assemblage is similar to those described from the Petseri Formation in the neighbouring cores of southeastern Estonia, for example Hino (452), Põlva (423), Luutsniku (451) (Volkova *et al.* 1981; Volkova 1983, 1990), Tartu (453) (by I. Paalits in Põldvere & Paalits 1998) and from the western part of the Pskov Region in Russia (Panikoviči core, Paalits 1992; Petseri (330) core, Volkova 1983). According to the acritarch zonation proposed for Estonia and neighbouring areas, the assemblage of the Petseri Formation in the Mehikoorma (421) core can be tentatively correlated with the *Olenus* trilobite Zone (Mens *et al.* 1993).

The samples (376.0 and 378.7 m) collected from the sandstones of the lower part (374.3–379.4 m) of the Petseri Formation were barren. For this reason the biostratigraphical conclusions about the Mehikoorma (421) section concern the middle part (371.4–374.3 m) of the formation only.

CHEMICAL COMPOSITION AND PHYSICAL PROPERTIES OF THE ROCK

A total of 133 rock samples from the Ediacaran (Upper Vendian), Cambrian, Ordovician and Devonian parts of the Mehikoorma (421) drill core were studied by geochemical methods (Appendix 14). Of those, 104 samples were additionally studied by petrophysical methods (Appendix 32). Thin sections were made from 43 samples to determine relationships between minerals, skeletal and nonskeletal rock-forming grains, cements, fabric, porosity and diagenetic alteration of rocks (Appendix 2). The investigated core section (Appendix 1, sheets 2–18) is represented by primary sedimentary rocks (limestones, calcitic marlstones, sand-, silt- and claystones) and by dolomitized rocks (dolostones, dolomitic marlstones, dolomite-cemented sand- and siltstones).

Methods

The bulk chemical composition of the rocks was determined by XRF spectrometry in the Laboratory of the All-Russian Geological Institute (VSEGEI), St. Petersburg. The insoluble residue (IR), MgO, CaO and FeO contents were additionally measured by wet chemical analysis in the Institute of Geology at Tallinn University of Technology (IG TUT).

Thin sections were prepared in the IG TUT. Chemical and physical parameters were interpreted together using correlation analysis.

Physical properties of the rock were analysed on 24 × 24 mm cubes at room temperature and pressure in the petrophysical laboratory of the Research Institute of Earth's Crust of St. Petersburg University (Prijatkin & Poljakov 1983; Shogenova & Puura 1998).

For density measurements samples were dried at a temperature of 100–110 °C and the weight of dry samples (P_d) was determined. Then the samples were saturated with water for 7 days and after that weighed in air (P_w) and water (P_{ww}). Dry density $\delta_d = P_d/V = P_d/(P_w - P_{ww})$, where V represents sample volume; wet density $\delta_w = P_w/(P_w - P_{ww})$; grain density $\delta_g = P_d/(P_d - P_{ww})$; and effective porosity $\varphi = (P_w - P_d)/(P_w - P_{ww})$ were calculated.

Relative dielectric permittivity E (dielectric constant) was determined on dry samples using the apparatus Q-meter E9–4. The dielectric permittivity measured with a high frequency depends on the mineral composition of rock. The samples were measured in three directions (X, Y, Z). The dielectric constant in the X direction (E_X) was calculated by the formula $E_X = 4.75C_X + 1$, where C_X is the measured capacitivity of the sample in the X direction. The dielectric constant (E) was calculated as the mean of the values measured for three directions: $E = (E_X + E_Y + E_Z)/3$.

The magnetic susceptibility of rock samples was determined with an IMV–2 susceptibility meter using a field of 64 A/m, and the isothermal remanent magnetization was measured on an ION–1 instrument (rock generator type) after the samples had been magnetized in a strong magnetic field of 300 kA/m.

Composition of rock samples

The IR, MgO and CaO contents found by wet chemical analysis, and other chemical parameters measured by XRF analysis were used to determine the rock lithology (Appendix 14). The rocks were subdivided into six lithological groups based on the following limits of the calculated and measured chemical components (Fig. 10): (1) limestone (IR < 25%, $CaCO_3 > CaMg(CO_3)_2$), (2) calcitic marlstone (25 < IR < 50%, $CaCO_3 > CaMg(CO_3)_2$), (3) dolostone (IR < 25%, $CaMg(CO_3)_2 > CaCO_3$), (4) dolomitic marlstone (25 < IR < 50%, $CaCO_3 < CaMg(CO_3)_2$), (5) mixed carbonate-siliciclastic rock (50 < IR < 70%) and (6) siliciclastic rock

(IR > 70%). The first group (12 samples) includes pure (IR < 10%) and variously argillaceous (10 < IR < 25%) Ordovician limestones (Appendix 2, T-27, T-32). Calcitic marlstones (Appendix 2, T-30, T-31) are represented by 8 Ordovician samples. The third group (45 samples) of Devonian and Ordovician dolostones is also represented by pure (Appendix 2, T-9, T-11, T-12, T-24, T-25) and variously argillaceous rocks (Appendix 2, T-21, T-22, T-34, T-36…38, T-41). Two samples (from depths of 301.6 and 337.2 m) in cluded into the dolostone group consist of calcitic dolostone. Dolomitic marlstones (Appendix 2, T-1, T-4, T-5, T-7, T-23, T-35, T-39) are represented by 18 samples. Two samples of calcitic dolomitic marlstones (depths 286.7 and 338.6 m) were also included in this group. The fifth group (14 samples) includes highly argillaceous Devonian and Ordovician silt- and sand-containing dolomitic marlstones (Appendix 2, T-6, T-10). Silt-, sand- and claystones (sixth group) are represented by 23 Devonian (Appendix 2, T-2, T-3, T-8, T-13…18, T-20), 1 Ordovician (Appendix 2, T-29), 10 Cambrian (Appendix 2, T-42) and 2 Ediacaran (Upper Vendian) samples (Appendix 2, T-43).

Dolostones together with dolomitic marlstones and siliciclastic rocks showed a high negative MgO–IR correlation (correlation coefficient $R = -0.79$) and a significant positive MgO–CaO correlation (0.45). The MgO–IR correlation for limestones together with calcitic marlstones was not significant and their MgO–CaO correlation was –0.57. The negative MgO–IR correlation for siliciclastic and mixed carbonate-siliciclastic rocks was very high (–0.95); their CaO–MgO correlation was 0.85 (Figs 10, 11).

The chemical constituents SiO_2, Al_2O_3, TiO_2 and K_2O had a high positive correlation with the IR (respectively 0.99, 0.91, 0.91 and 0.96) and with each other in carbonate rocks (Figs 11, 12). The correlation coefficient of Na_2O with the IR for carbonate rocks was 0.75, which is lower than that for other components entering the IR. SiO_2 enters the IR in the form of clastic and clay minerals. K_2O and Na_2O may enter also clastic (feldspar) and clay minerals. The higher K_2O content in siliciclastic and mixed carbonate-siliciclastic rocks compared to carbonate rocks shows the abundance of K-feldspar in siliciclastic rocks (Fig. 12B). Potassium feldspar is a typical clastic component in Cambrian rocks (Shogenova *et al.* 2001). Na_2O occurs usually in very low (< 0.3%) amounts and is nearly absent (below the detection limit of XRF spectrometry of 0.02%) in several Devonian and Ordovician rocks (see Appendix 14). Al_2O_3 and TiO_2 may serve as indicators of the clay content because they enter only clay minerals (Figs 11, 13).

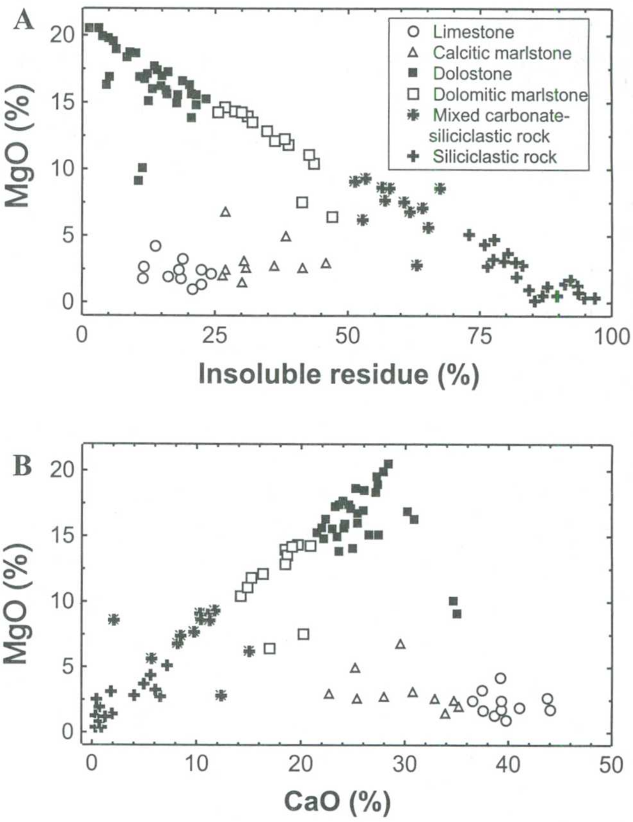

Fig.10. (A) MgO content versus insoluble residue content measured by wet chemical analyses. Correlation coefficient $R = -0.79$ for dolostones and dolomitic marlstones, $R = 0.32$ for limestones and calcitic limestones, $R = -0.95$ for siliciclastic and mixed carbonate-siliciclastic rocks.

(B) MgO content versus CaO content measured by wet chemical analyses. $R = 0.45$ for dolostones and dolomitic marlstones, $R = -0.57$ for limestones and calcitic limestones, $R = 0.85$ for siliciclastic and mixed carbonate-siliciclastic rocks.

The total iron (Fe_2O_3 total) content of most of the studied rocks correlated with the clay content. The correlation of Fe_2O_3 total with Al_2O_3 as an indicator of clay content was the highest (0.92) in limestones and calcitic marlstones (Fig. 13), and higher than in the Ruhnu (500) core (0.78; Shogenova *et al.* 2003a). The correlation was lower in siliciclastic and mixed carbonate-siliciclastic rocks (0.71) and the lowest in dolostones and dolomitic marlstones (0.58). The lowest correlation in dolomitized rocks is due to the absence of the Fe_2O_3 total–Al_2O_3 correlation in the Middle Ordovician rocks except in the rocks of the Väo Formation (Fig. 11, Appendix 14). This may be explained by secondary nature of iron-bearing minerals in Middle Ordovician rocks (Kriukai to Stirnas formations). These rocks had the highest total iron content for the given Al_2O_3 content. Limestones had a lower iron content for the given Al_2O_3 content, and it was the lowest in siliciclastic rocks. In general, lime-

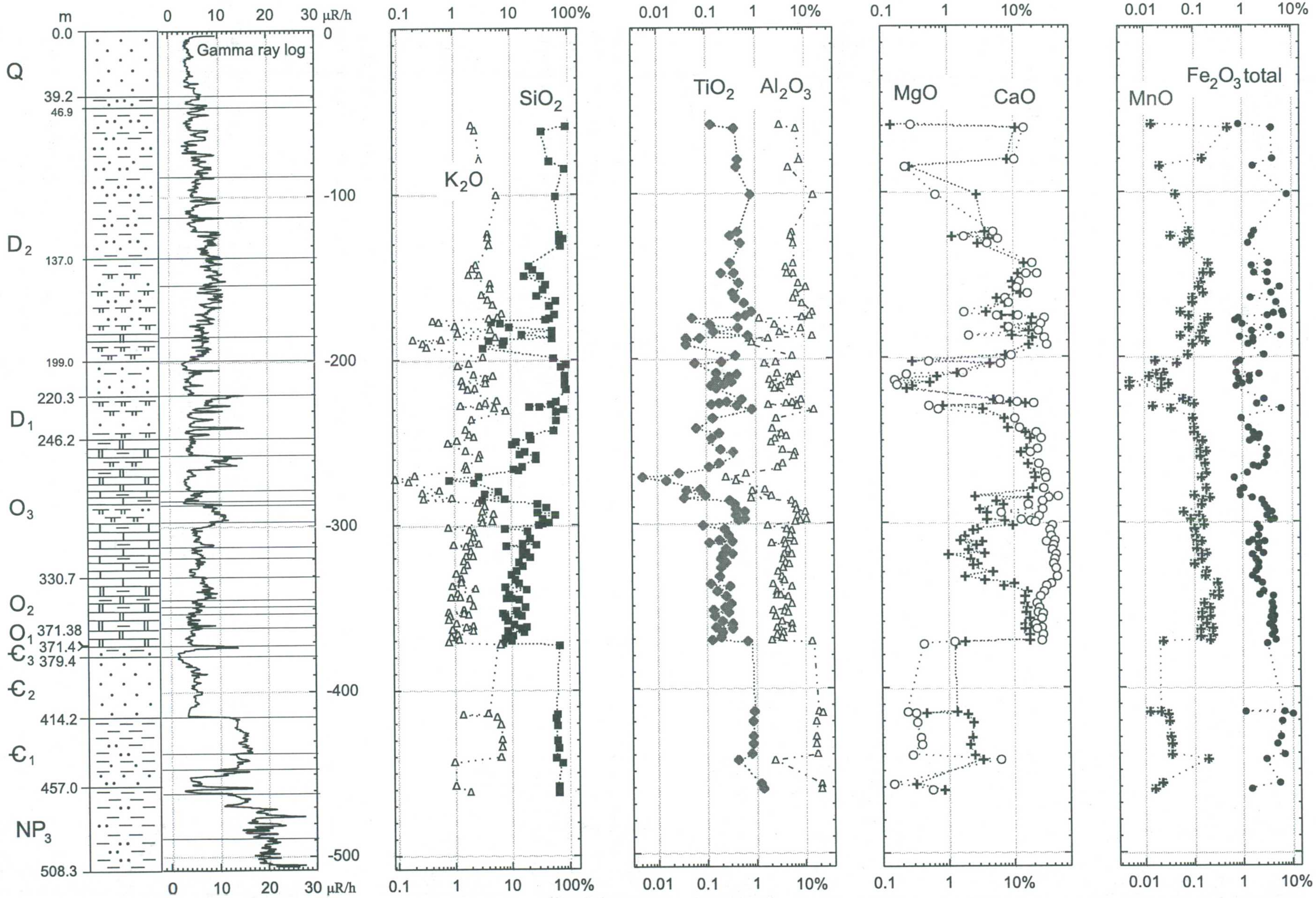

Fig. 11. Chemical composition of the Mehikoorma (421) core determined by XRF spectrometry. NP_3 – Ediacaran (Upper Vendian); $Є_1$ – Lower Cambrian; $Є_2$ – Middle Cambrian; $Є_3$ – Furongian (Upper Cambrian); O_1 – Lower Ordovician; O_2 – Middle Ordovician; O_3 – Upper Ordovician; D_1 – Lower Devonian; D_2 – Middle Devonian. Refer to Appendix 1 for lithology.

stones had the lowest total iron content (0.86–2.23%); it was higher in calcitic marlstones (2.07–3.11%), dolostones (0.64–4.53%) and dolomitic marlstones (1.53–4.57%). The total iron content was the highest in the mixed carbonate-siliciclastic (0.95–6.7%) and siliciclastic rocks (0.65–10.2%; Figs 11, 13).

The MnO content was generally higher in the carbonate rocks than in siliciclastic rocks in the studied core (Fig. 14). Unlike the Ruhnu (500) core (Shogenova *et al.* 2003a), MnO had no significant correlation with total iron content, but it correlated with MgO (R = 0.71 for all rocks and R = 0.78 for mixed carbonate-siliciclastic and siliciclastic rocks). In reality, only Devonian rocks taken together had a high MnO–MgO correlation. Except for one dolomitic marlstone sample with 0.5% of Mn (depth 61.5 m), the correlation was 0.85 for Devonian rocks. The MnO–MgO correlation coefficient for upper Ordovician rocks was lower (0.42) and the increase in MnO was not great. Upper Ordovician dolostones had the same MnO (0.12–0.21%) range as Devonian dolostones (0.094–0.220%). Middle Ordovician rocks had a higher MnO content (0.13–0.31%) with the highest values in the Väo and Stirnas formations (Fig. 11, Appendix 14).

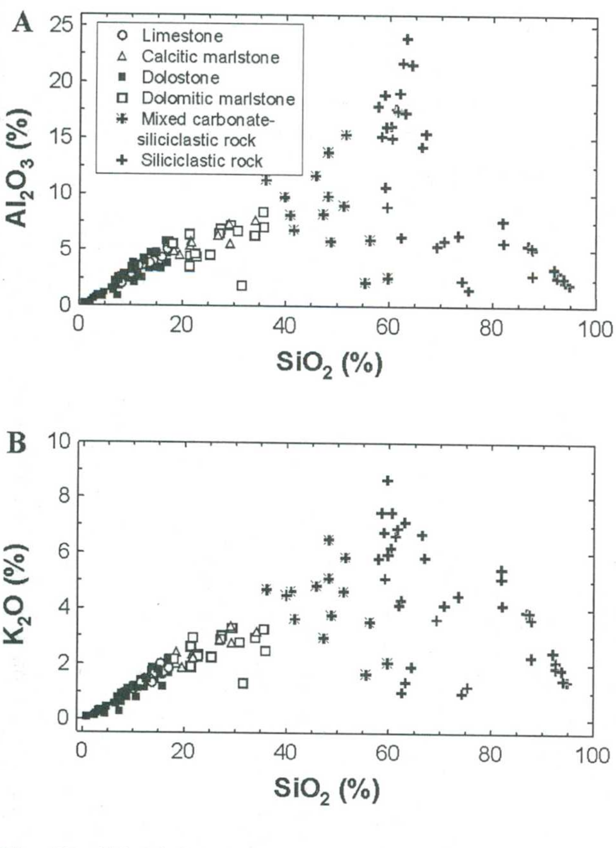

Fig. 12. (A) Al_2O_3 content versus SiO_2 content. Correlation coefficient R = 0.88 for carbonate rocks.
(B) K_2O content versus SiO_2 content. R = 0.92 for carbonate rocks.

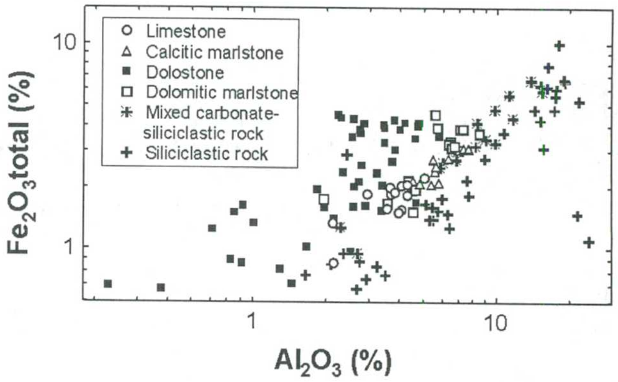

Fig. 13. Total iron content versus Al_2O_3 content as the indicator of clay content. Correlation coefficient R = 0.92 for limestones and calcitic marlstones, R = 0.58 for dolostones and dolomitic marlstones, R = 0.71 for mixed carbonate-siliciclastic and siliciclastic rocks.

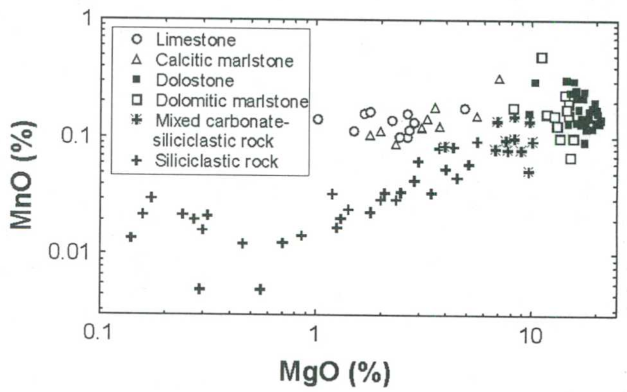

Fig. 14. MnO content versus MgO content determined by XRF spectrometry. Correlation coefficient R = 0.78 for mixed carbonate-siliciclastic and siliciclastic rocks, R = 0.71 for all rocks except for the limestones and calcitic marlstones, R = 0.61 for all rocks together.

Porosity and density

Middle Ordovician dolomitized rocks had the highest wet density for the given porosity (Figs 15, 16A). The studied rocks are lithologically well discriminated on the porosity–wet density plot (Fig. 16A). The correlation lines for dolomitized carbonate rocks, primary carbonates and clastic rocks are parallel and their correlation coefficients are –0.95, –0.84 and –0.93, respectively.

The porosity of the carbonate rocks from the Mehikoorma (421) core, as of most of Estonian carbonate rocks (Shogenova 1998; Shogenova *et al.* 2003a; Shogenova *et al.* 2003b), has a significant positive correlation with the Al_2O_3 content, an indicator of the presence of clay (Figs 15, 16B).

Among the studied rocks, Upper Ordovician dolostones and limestones (Fig. 15, 265.8–284.8 m; Fig. 16B) had the lowest porosity for the given Al_2O_3 content (1.24–11.00%). The porosity of Devonian (5.0–14.8%) and Middle Ordovician dolostones (6.5–14.0%) was higher for the given Al_2O_3 content and had lower correlation with it. This is explained

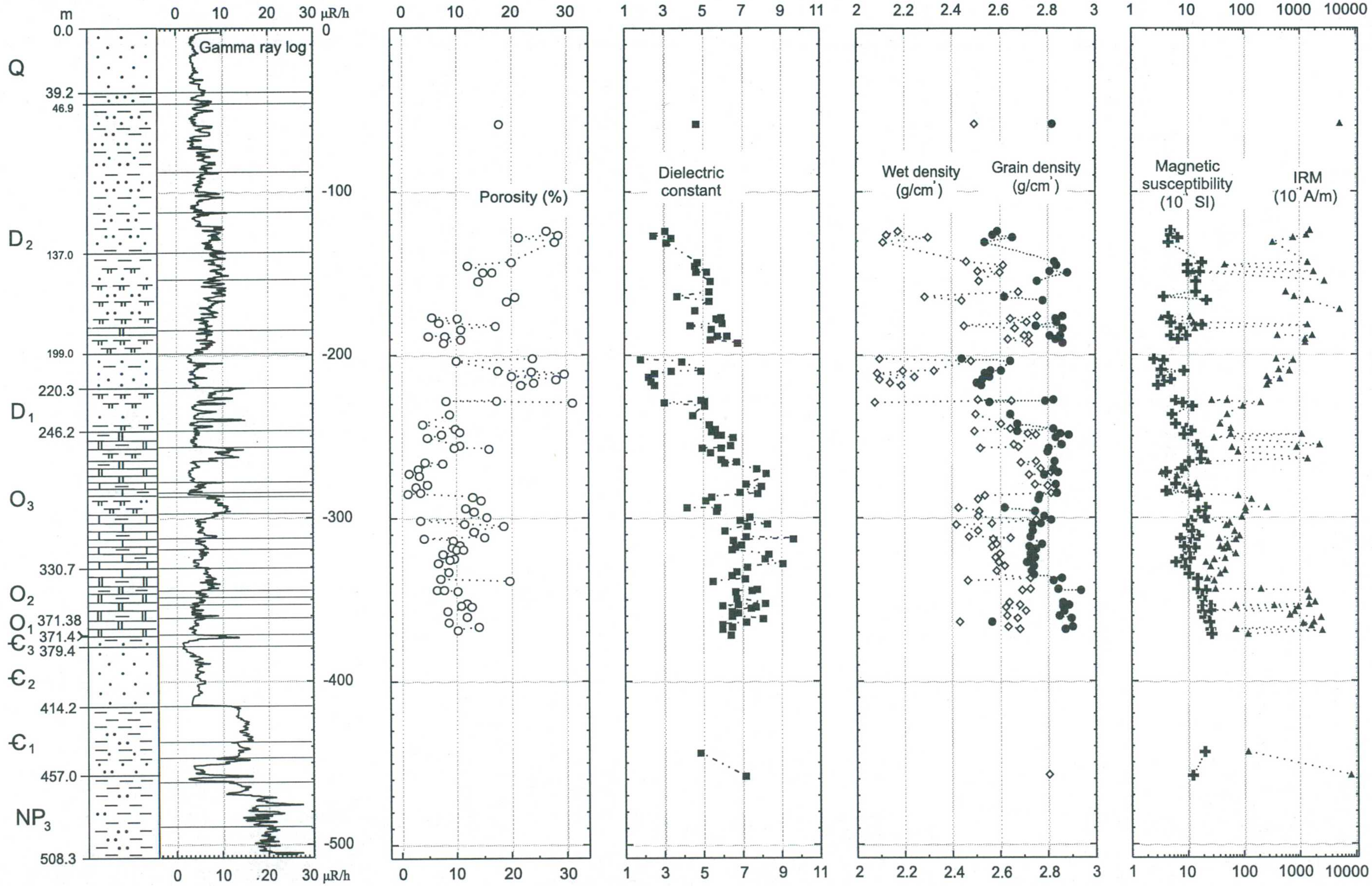

Fig. 15. Physical properties of the Mehikoorma (421) core. NP_3 – Ediacaran (Upper Vendian); $Є_1$ – Lower Cambrian; $Є_2$ – Middle Cambrian; $Є_3$ – Furongian (Upper Cambrian); O_1 – Lower Ordovician; O_2 – Middle Ordovician; O_3 – Upper Ordovician; D_1 – Lower Devonian; D_2 – Middle Devonian; IRM – isothermal remanent magnetization. Refer to Appendix 1 for lithology.

by secondary nature of porosity of Devonian and Middle Ordovician dolostones. The porosity of limestones was lower than that of dolostones for the given IR. Siliciclastic Devonian rocks had the highest porosity (10–31%) and the lowest wet (2.08–2.81 g/cm^3) and grain density (2.54–2.82 g/cm^3), whereas porosity was the highest and density the lowest in the siliciclastic rocks of the Vadja, Leivu, Gorodenka and Aruküla formations (Fig. 15, Appendix 32). The Cambrian and Ediacaran (Upper Vendian) rocks were not measured for porosity, because they were loosely cemented.

The porosity of mixed carbonate-siliciclastic rocks (3.95–19.3%) was lower than that of siliciclastic rocks and nearly in the same range as the porosity of marlstones (8–20%), although mixed rocks had a higher Al_2O_3 content (Fig. 16B). Wet density (2.44–2.61 g/cm^3) of the mixed carbonate-siliciclastic rocks was close to that of marlstones (Fig.16A), but grain density (2.65–2.78 g/cm^3) of mixed rocks was lower.

Dolostones had the highest wet (2.43–2.82 g/cm^3) and grain densities (2.78–2.94 g/cm^3) among the studied rocks and the highest wet density for the given porosity (Fig. 16A).

Dielectric constant

The dielectric constant of rocks depends on their porosity, density and mineral composition (Figs 15, 17). The dielectric constant of all studied rocks had a negative correlation with porosity and SiO_2 content and a positive correlation with wet and grain density and CaO content. A high positive correlation was also recorded for all dolomitized rocks taken together, mixed carbonate-siliciclastic and siliciclastic rocks included (0.8). The dielectric constant of limestones (6.5–9.6) was the highest among the studied rocks. Calcitic marlstones (5.4–8.2), dolostones (4.6–8.4) and dolomitic marlstones (4.6–8.0) had a lower, and siliciclastic (1.8–7.1) and mixed rocks (4.3–5.4) the lowest dielectric constant. The dielectric constant was the lowest in the porous siliciclastic rocks of the Devonian Pärnu Formation (Fig. 15).

Magnetic properties

Low-field magnetic susceptibility in the studied rock sequence correlated with the total iron content (Fig. 18) and increased from diamagnetic and paramagnetic to ferromagnetic minerals. The correlation coefficient was 0.94–0.95 for different rock groups.

Magnetic susceptibility for the given iron content was higher in carbonate rocks and lower in siliciclastic rocks. Among carbonate rocks Upper Ordovician dolostones had a higher magnetic susceptibility (Fig. 15). Variability in magnetic properties of the rocks with similar iron contents is explained by different mineral compositions of rocks. Paramagnetic properties of the studied rocks are mainly explained by illite but also by chlorite content in the clay fraction (Appendix 19). Ferromagnetic properties of the rock groups depend on magnetic carriers. The carriers may be represented by antiferromagnetic goethite and haematite, sometimes by siderite in the clay fraction, and also by ferrimagnetic magnetite, paramagnetic pyrite and ilmenite in the heavy mineral fraction (Appendixes 17, 19) of Devonian siliciclastic rocks. Goethite and haematite pigmentation of Devonian siliciclastic rocks (Appendix 2, T-2, T-3, T-8, T-13, T-15, T-17, T-18) and dolomitic marlstones (Appendix 2, T-1, T-4...7, T-10) may be seen visually and in thin sections. Pyrite is a paramagnetic mineral, which could enter, together with goethite and haematite, Devonian dolostones (Appendix 2, T-12, T-19) and Ordovician carbonate rocks (Appendix 2, T-21...34, T37...T41). Ilmenite, magnetite, goethite and pyrite were also found in the studied heavy mineral fraction of Upper Ordovician rocks (Appendix 17).

Twelve of the studied Ordovician limestone samples had the total iron content of 0.86–2.2% and low magnetic susceptibility (4.2×10^{-5} to 12.6×10^{-5} SI). Calcitic marlstones had a higher total iron content (2.1–3.1%) and higher magnetic susceptibility (10×10^{-5} to 16.7×10^{-5} SI). Dolostones had a wider range of total iron content (0.64–4.53%) and magnetic susceptibility (3.3×10^{-5} to 27.6×10^{-5} SI), whereas the highest values were recorded in Middle and Lower Ordovician dolostones (Figs 11, 15). The highest magnetic susceptibility of 27.6×10^{-5} SI was measured in two brownish-red samples (depths 366.6 and 371.3 m) from the Kriukai Formation (Appendix 32).

Siliciclastic rocks had the widest limits of total iron content (0.65–10.2%), but their magnetic susceptibility range (2.5×10^{-5} to 21.3×10^{-5} SI) was less than that of dolostones. Mixed carbonate-siliciclastic rocks had also a wider range of total iron content (0.95–6.7%) but a narrower range of magnetic susceptibility (5.4×10^{-5} to 21.8×10^{-5} SI) than dolostones. Among samples where magnetic properties were measured, Ediacaran (Upper Vendian) goethite pigmented claystones of the Voronka Formation had the highest total iron content (5.4%) (Appendix 2, T-43), but their magnetic susceptibility was only 12.1×10^{-5} SI.

High values of isothermal remanent magnetization are an evidence of ferromagnetic properties in part of the siliciclastic and mixed carbonate-siliciclastic rocks, dolostones and dolomitic marlstones

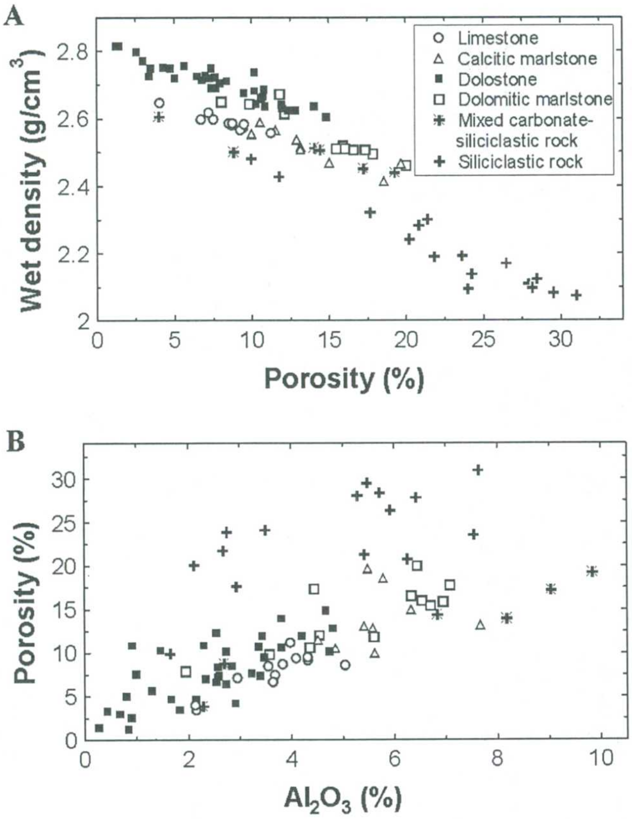

Fig. 16. (A) Wet density versus porosity. Correlation coefficient $R = -0.95$ for limestones and calcitic marlstones, $R = -0.84$ for dolostones and dolomitic marlstones, $R = -0.95$ for siliciclastic rocks, $R = -0.93$ for siliciclastic and mixed carbonate-siliciclastic rocks together.
(B) Porosity versus Al_2O_3 content as indicator of clay content. $R = 0.74$ for limestones and calcitic marlstones, $R = 0.84$ for dolostones and dolomitic marlstones, $R = 0.82$ for carbonate and mixed carbonate-siliciclastic rocks together.

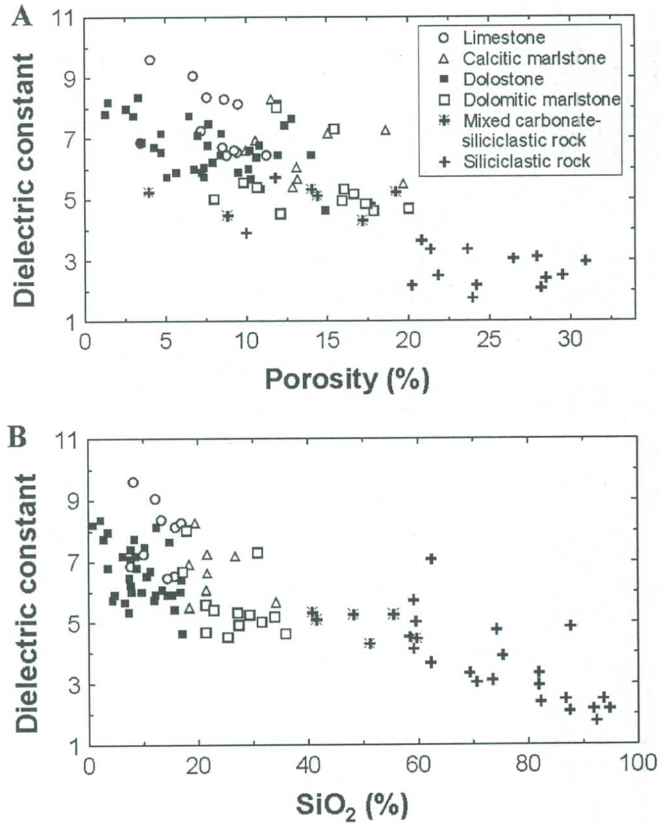

Fig. 17. (A) Dielectric constant versus porosity. (B) Dielectric constant versus SiO_2 content.

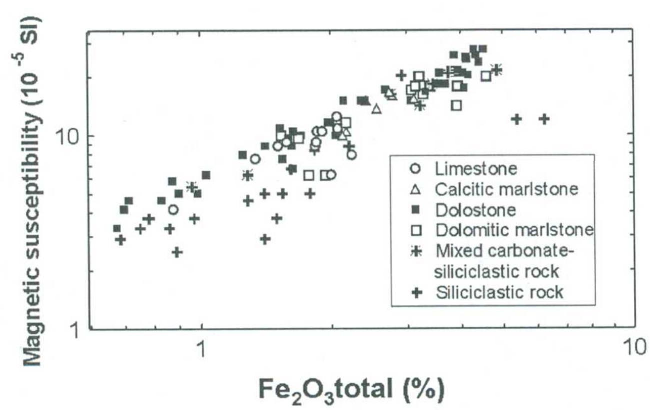

Fig. 18. Magnetic susceptibility versus total iron content. Correlation coefficient $R = 0.94$ for carbonate rocks, $R = 0.96$ for siliciclastic and mixed carbonate-siliciclastic rocks.

(Fig. 15). These rocks from all stratigraphic levels include ferrimagnetic (magnetite) and antiferromagnetic (goethite and haematite) minerals. The high total iron content and ferromagnetic properties of Lower and Middle Ordovician dolostones could be of secondary origin.

Conclusions

The studied Devonian and Ordovician sequence of the Mehikoorma (421) core has undergone several diagenetic processes that have influenced the composition and physical properties of the rock. The increase in total iron content is associated with dolomitization of Middle Ordovician rocks (Kriukai–Stirnas formations), and early–late dolomitization and secondary mineralization of Devonian rocks. The increase in total iron content during dolomitization has already been reported in Ordovician and Silurian sequences of Estonia (Shogenova 1999).

Iron minerals in Upper Ordovician and Devonian rocks are associated with argillaceous minerals. The content of titanium-bearing minerals is higher in siliciclastic rocks than in carbonate rocks. During dolomitization, replacement of clay minerals in carbonate rock took place.

Unlike in the Ruhnu (500) core (Shogenova *et al.* 2003a), the MnO content is not associated with iron minerals but is controlled by dolomitization in the Mehikoorma (421) core.

The density–porosity plot reveals a good discrimination of primary carbonate, dolomitized carbonate and siliciclastic rocks.

The dielectric constant has a good negative correlation with porosity and SiO_2 content and a positive correlation with wet density and CaO content.

The magnetic susceptibility versus total iron content plot shows both lithological and stratigraphic discrimination. Magnetic susceptibility for the given total iron content is the highest in Ordovician dolomitized rocks, lower in Devonian dolostones and the lowest in Devonian siliciclastic rocks. This may be explained by the variability of the mineral composition of rocks with paramagnetic and ferromagnetic properties.

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APPENDIX 1

DESCRIPTION OF THE MEHIKOORMA (421) CORE

The description is given in a standardized form. The table is divided into nine columns based on the type of information.

STANDARD UNITS — Chronostratigraphic and geological time units.

LOCAL STRATIGRAPHIC UNITS — Stages, substages, formations, members, beds and block.

CORE BOX NO./FIGURES — Numbers of boxes, location of the intervals of core illustrated on compact disc in the read-only memory (detailed core photos marked as D-1…23, thin sections as T-1…43, and Ordovician photo-log in Appendix 4).

DEPTH/SAMPLES — Depth of the boundaries and sample levels: A, acritarchs; B, brachiopods; C, conodonts; Ch, chitinozoans; F, X-ray fluorescence samples; G, granulometric samples; Is, stable isotope analyses ($\delta^{13}C$); K, chemical samples; M, mineralogical samples; Ms, miospores; O, ostracods; Ph, physical properties; Ps, fishes (agnathans, gnathostomes, otoliths); T, thin sections; X, X-ray diffractometry.

LITHOLOGY — For legend see the next page. The core section is given at a scale of 1:200.

SEDIMENTARY STRUCTURES — According to thickness of beds: micro- (< 0.2 cm), thin- (0.2–2.0 cm), medium- (2–10 cm) and thick-bedded (10–50 cm); massive – visible bedding is missing. According to size of nodules: thin-nodular (vertical diameter of nodules < 0.2 cm), medium-nodular (2–5 cm) and thick-nodular (> 5 cm).

MARLSTONE BEDS — The most frequent thicknesses of the marlstone beds; in parentheses infrequent thicknesses. Contacts between marlstone and other types of rock may be distinct (D) or indistinct (IND). Colours were identified on damp core.

MARLSTONE PERCENTAGE — The content of marlstone beds in the described interval was estimated visually.

SHORT DESCRIPTION — Main types of rocks are in bold. The colour of rocks was identified on damp core; the dominant size of limestone crystals (in italics) was estimated visually: cryptocrystalline (< 0.005 mm), microcrystalline (0.005–0.01 mm), very finely crystalline (0.01–0.05 mm), finely crystalline (0.05–0.1 mm), medium-crystalline (0.1–1.0 mm) and coarsely crystalline (> 1.0 mm). The percentage of allochems (mainly bioclasts and clastic material) is also indicated. Clastic fractions (size of particles; in italics) are described as follows: clay (< 0.005 mm), fine silt (0.005–0.01 mm), coarse silt (0.01–0.05 mm), very fine sand (0.05–0.1 mm), fine sand (0.1–0.25 mm), medium sand (0.25–0.5 mm), coarse sand (0.5–1.0 mm), very coarse sand (> 1 mm) and gravel (2–10 mm). Grain size limits for igneous rocks (in italics) are: fine-grained (< 1 mm) and medium-grained (1–5 mm).

Appendix 1 continued

LEGEND

- limestone (in general)
- argillaceous limestone
- dolomitized limestone
- silty limestone
- dolostone
- argillaceous dolostone
- calcitic marlstone
- dolomitic marlstone (in general)
- claystone
- silty claystone
- siltstone
- argillaceous siltstone
- sandstone
- migmatite granite (a), weathered migmatite granite (b)
- K-bentonite bed, on the boundary
- skeletal limestones: grains 10–25% (a) and grains 25–50% (b)
- crypto- and microcrystalline limestone (a) and dolostone (b)
- fine bioclasts, pyritized (0.05–1.0 mm)
- coarse bioclasts, pyritized (> 1 mm)
- horizontal bedding: thin (a), medium (b) and thick bedding (c)
- wavy bedding
- nodular
- thin intercalation
- breccia
- nodules
- discontinuity surface
- mud crack
- slickenside
- veins
- caverns (vugs)
- porous
- burrows
- pyritic mottles
- mottled, red-coloured and yellow streaks
- ooliths
- clastic material
- feldspar grains
- glauconite grains
- quartz grains
- kerogen
- barytes
- chlorite
- dolomite
- limonite
- micas (in general)
- pyrite
- siderite
- brachiopods
- trilobites
- ostracods
- cephalopods
- gastropods
- lingulates
- calcareous algae
- siliceous sponges
- bryozoans
- echinoderms (crinoids)
- tabulate corals
- fishes

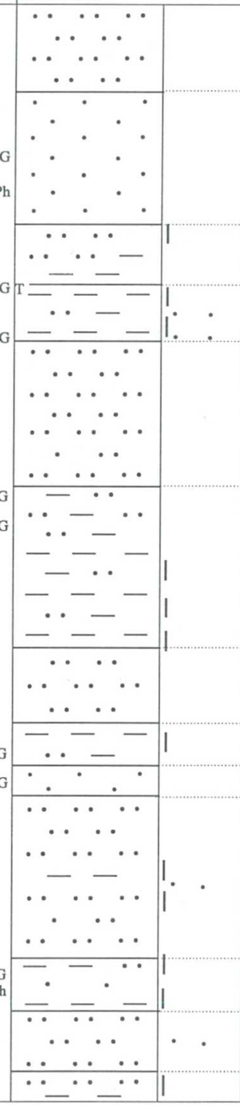
DESCRIPTION OF THE MEHIKOORMA (421) CORE

Location: 58° 14' 26" N, 27° 27' 09" E. Length of the core 548.5 m. Elevation of the top 32.5 m above sea level.

STANDARD UNITS	LOCAL STRATIGRAPHIC UNITS	CORE BOX NO.	FIGURES	DEPTH (m)	SAMPLES	LITHOLOGY	SEDIMENTARY STRUCTURES	MARLSTONE BEDS	MARLSTONE PERCENTAGE	SHORT DESCRIPTION
Upper Pleistocene, Quaternary		1, 2		0.0			! (Core yield 60%)			Brownish-grey, with yellow tinge, dirty, silty, mostly *medium-* to *fine-grained* **sand**
		2, 3, 4, 5, 6		7.7			! (Core yield 85%)			Whitish-grey, dirty, silty, *medium-* to *fine-grained* **sand**

STANDARD UNITS	LOCAL STRATIGRAPHIC UNITS	CORE BOX NO.	FIGURES	DEPTH (m)	SAMPLES	LITHOLOGY	SEDIMENTARY STRUCTURES	MARLSTONE BEDS	MARLSTONE PERCENTAGE	SHORT DESCRIPTION
Upper Pleistocene, Quaternary		6, 7, 8								☞ follow up
				39.2						**Till**, represented by pebbles of diorite (diameter up to 10 cm) and porphyrite (up to 5 cm)
Middle Devonian, Givetian	Burtnieki Stage, Burtnieki Formation, Härma Member	9		39.5	XG		Rubbly			Reddish-brown, with brown patches (in the upper part mostly grey), silty **claystone** interbedded with greenish-grey, medium-cemented, clayey **siltstone** (thickness 5–20 cm)
				42.4			(Core yield 30%)			Brownish-grey, weakly cemented, silty, *fine-grained* **sandstone**
Middle Devonian, Eifelian	Aruküla Stage, Aruküla Formation, Tarvastu Member	10		46.9	G		Rubbly (core yield 55%)			Reddish-brown, with violetish-grey patches (lower 0.2 m rusty-brown), silty **claystone** interbedded with greenish-grey, medium-cemented, sandy **siltstone** (thickness 3–10 cm)
				51.7			Rubbly (core yield 50%)			Yellowish-grey **siltstone**

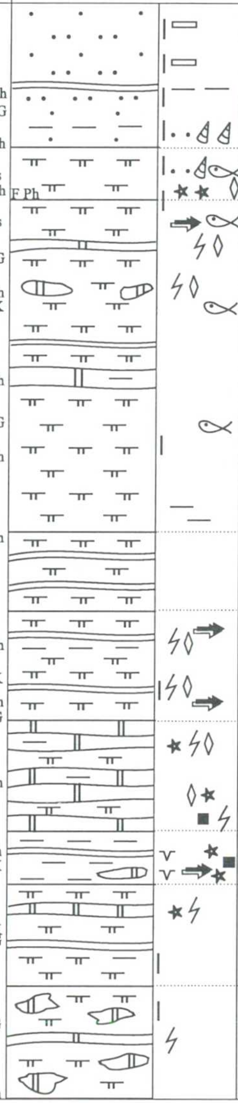
MEHIKOORMA (421) DRILL CORE

STANDARD UNITS	LOCAL STRATIGRAPHIC UNITS	CORE BOX NO.	FIGURES	DEPTH (m)	SAMPLES	LITHOLOGY	SEDIMENTARY STRUCTURES	MARLSTONE BEDS	MARLSTONE PERCENTAGE	SHORT DESCRIPTION
Middle Devonian Eifelian	Aruküla Stage Aruküla Formation Tarvastu Member	10								☞ follow up
				56.3						
		11			MG F Ph		Cross-bedded (core yield 75%)			Reddish-brown (lower 0.1 m pinkish-grey), medium-cemented, *fine-grained* **sandstone**
				59.8			Horizontal, in places platy			Intercalation of medium-cemented **silt-** and **claystone** (dominating in the lower part). The complex is brownish-grey, in places siltstone layers are greenish- and bluish-grey
			T-1	61.4	F Ph X G T					
					MG		Horizontal, in the upper part platy	2–5 cm; IND reddish-brown	5	Reddish-brown, with greenish- and yellowish-grey patches, silty **claystone** interbedded with grey silty **sandstone** and, in the upper part, reddish-brown silty **dolomitic marlstone** (thickness 1–2 cm)
				62.9			(Core yield 65%)			Brownish-grey sandy **siltstone**
		12		66.7	G MG		Indistinctly bedded, rubbly (core yield 35%)			Reddish-brown, with greyish-brown interlayers (in the lower part with violetish-grey patches), silty **claystone** interbedded with grey **siltstone** and bluish-grey **sandstone**
				71.0			(Core yield 50%)			Brownish-grey **siltstone**
				73.0	XG		Rubbly (core yield 55%)			Reddish-brown **claystone** with greenish-grey sandy **siltstone** interbeds
				74.1	MG		(Core yield 50%)			Yellowish- to reddish-brown (lower and upper parts are rusty-brown), weakly cemented, *fine-grained* **sandstone**
		13		74.9			(Core yield 65%)			Brownish-grey, sandy **siltstone**, with a reddish-brown and greenish-grey mottled **claystone** interbed (thickness 10 cm) and a brownish-grey, thin, strongly cemented **sandstone** interbed in the middle part
				79.2	MG F Ph		Indistinctly wavy, siltstone platy (core yield 65%)			Thin intercalation of reddish-brown, medium-cemented **siltstone**, violetish-brown (in the lower part greenish-grey mottled) silty **claystone** and greyish-brown, strongly to medium-cemented (in places with globules) silty **sandstone** (thickness of beds 5–7 cm)
		14		80.6			(Core yield 75%)			Brownish-grey, sandy **siltstone**
				82.2			Indistinctly bedded, siltstone platy			Intercalation of brownish-grey **siltstone** and reddish-brown (with greenish-grey patches and interbeds) **claystone** (in the lower part sandy)

STANDARD UNITS	LOCAL STRATIGRAPHIC UNITS	CORE BOX NO.	FIGURES	DEPTH (m)	SAMPLES	LITHOLOGY	SEDIMENTARY STRUCTURES	MARLSTONE BEDS	MARLSTONE PERCENTAGE	SHORT DESCRIPTION
Middle Devonian, Eifelian	Aruküla Stage, Aruküla Formation, Tarvastu Member									☞ follow up
		14		84.1–88.5	F Ph, MG		Cross-bedded, claystone platy (core yield 30%)			Purplish-brown, medium-cemented, silty **sandstone** with reddish-brown and violetish-grey **claystone** and bluish-grey **siltstone** interbeds. Bedding surfaces are covered with mica flakes
	Kureküla Member			88.5–89.8	MG, G		Thin-bedded, in places thick-bedded (core yield 60%)			Reddish-brown (in some layers with greenish-grey patches and pockets), medium-cemented **siltstone** with reddish-brown **claystone** and reddish-brown *fine-* to *very fine-grained* **sandstone** interbeds. Grain-size and content of sandy material change vertically
		15		89.8–96.8			Rubbly (core yield 55%)			Brownish-grey, sandy **siltstone** with a reddish-brown silty **claystone** interbed in the upper part
				96.8–98.1	MG		Thin-bedded			Reddish-brown, weakly cemented, *very fine-grained* **sandstone**. Bedding surfaces are covered with mica flakes
				98.1–100.5			(Core yield 65%)			Brownish-grey, sandy **siltstone**
		16		100.5–102.6	F Ph, MG, G		Thin-bedded (core yield 70%)			Reddish-brown, in places light bluish-grey, medium-cemented **siltstone** with rare clayey interbeds. The upper part includes greyish-brown, weakly cemented, *very fine-grained* **sandstone** interbeds (thickness 2–20 cm). At 101.0–101.1 m yellowish-grey **sandstone** bedding surfaces are covered with mica flakes
				102.6–104.1			Rubbly (core yield 65%)			Brownish-grey, sandy **siltstone**
				104.1–104.3	MG		(Core yield 50%)			Greyish-brown, weakly cemented, *fine-* to *very fine-grained* **sandstone**
				104.3–107.3	B Ps		Indistinctly thin-bedded, in places rubbly (core yield 75%)			Yellowish-, reddish- and purplish-brown (in places with bluish-grey pockets and interbeds), medium-cemented **siltstone** with rare clayey interbeds. Clay content increases downwards
		17		107.3–112.0	MG		Cross-bedded (core yield 10%)			Yellowish-brown (in places purplish-brown), weakly cemented, *fine-* to *medium-grained* **sandstone**

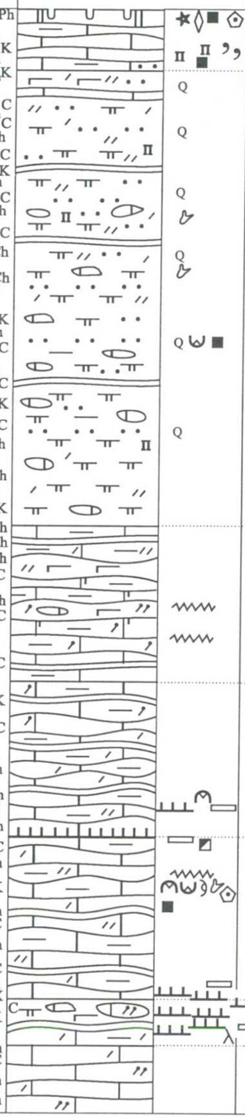
STANDARD UNITS	LOCAL STRATIGRAPHIC UNITS	CORE BOX NO.	FIGURES	DEPTH (m)	SAMPLES
Middle Devonian, Eifelian	Narva Stage; Leivu Substage, Leivu Formation	24	T-8		TF Ph, X MG
				173.8	F Ph
				175.2	Ps, F Ph, F Ph
					Ps
					F Ph, X MG
			T-9		T F Ph, K
					F Ph
					Ps Ps X MG
			T-10		T F Ph
		25		184.0	F Ph
				186.1	
					F Ph
			T-11		TF Ph K, F Ph
	Narva Stage; Vadja Substage, Vadja Formation			189.0	X MG
			T-12		TF Ph
				191.9	F Ph, K X
				193.3	
		26			K X X MG K
				196.0	Ms X MG
				199.0	F Ph

LITHOLOGY



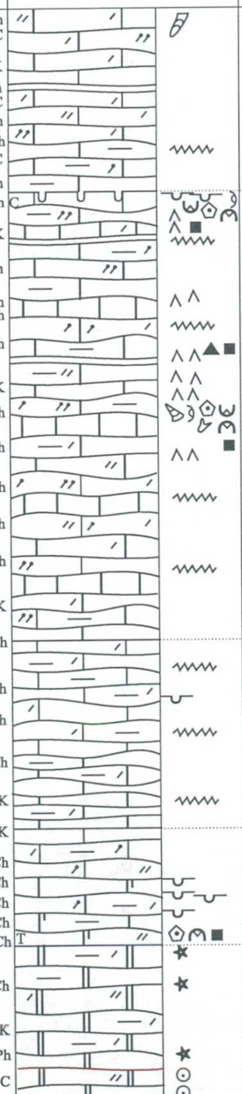
DEPTH (m)	SEDIMENTARY STRUCTURES	MARLSTONE BEDS	MARLSTONE PERCENTAGE	SHORT DESCRIPTION
–173.8	Horizontally medium- to thin-bedded, claystone platy (core yield 40%)			Intercalation of reddish-brown, medium-cemented, clayey **siltstone**, brownish-grey and light grey, weakly to medium-cemented, *fine-* to *very fine-grained* **sandstone** and reddish-brown (with greenish-grey patches), silty **claystone**. Bedding surfaces are covered with mica flakes
173.8–175.2	Indistinctly bedded, in places rubbly	2–5 cm; IND mainly reddish-brown	100	Greyish-, reddish- and violetish-brown mottled (with greenish-grey patches), in places silty **dolomitic marlstone**. The lowermost 0.1 m is light grey, strongly cemented and with vugs (filled with authigenic dolomite crystals)
175.2–184.0	Indistinctly thin- to thick-bedded, in places rubbly; dolostone wavy-bedded, in places breccia-like (core yield 70%)	Up to 3 (5) cm; IND mainly greenish-grey	85	Greenish-grey (in the upper and lower parts with rare reddish-brown and violet patches), argillaceous and silty **dolomitic marlstone** with light grey, mainly up to 20 cm thick **dolostone** and rare thin grey **claystone** interbeds. Vugs in dolostone are filled with authigenic dolomite crystals. Clay content changes vertically and increases downwards. At 179.7–180.3 m occurs greenish-grey, strongly cemented *aphanocrystalline* **dolostone**
184.0–186.1	Indistinctly bedded, dolostone thick-bedded (core yield 50%)	Up to 4 cm; IND mainly greenish-grey	80	Greenish-grey, argillaceous, **dolomitic marlstone** with light grey **dolostone** interbeds in the upper and lower parts
186.1–189.0	Indistinctly bedded, dolostone medium-bedded (core yield 70%)	Up to 4 cm; IND grey	50	Intercalation of grey **dolomitic marlstone**, dark grey dolomitic **claystone** (30%) and grey aphanocrystalline **dolostone** (20%). Dolostone contains authigenic dolomite crystals and clay-filled fractures
189.0–191.9	Indistinctly thin- to thick-bedded (core yield 55%)	Up to 3 cm; IND greenish-grey	5	Light grey, in places yellowish-grey, *aphanocrystalline* **dolostone** with rare **dolomitic marlstone** interbeds. Dolostone fractures are filled with grey and brownish-grey clay. Vugs are partially filled with authigenic dolomite crystals
191.9–193.3	Indistinctly thin-bedded, nodular (core yield 90%)			Dark grey dolomitic **claystone** with light and yellowish-grey, fractured *aphanocrystalline* **dolostone** interbeds (thickness 3–5 cm)
193.3–196.0	Indistinctly bedded, in places nodular (core yield 75%)	Up to 5 (10) cm; IND grey	50–60	Grey (lowermost 1.4 m dark grey), argillaceous **dolomitic marlstone** with light and yellowish-grey *aphanocrystalline* **dolostone** interbeds. Dolostone contains fractures (filled with dark and brownish-grey clay) and vugs
196.0–199.0	Breccia-like, nodular (core yield 20%)	Up to 3 (10) cm; IND grey	20–30	Grey, in the lowermost part silty **dolomitic marlstone** with light grey **dolostone** clasts (diameter 0.1–3.5 cm). In the middle yellowish-grey fractured dolostone and brownish-grey dolomitic **claystone** interbeds occur

STANDARD UNITS	LOCAL STRATIGRAPHIC UNITS	CORE BOX NO.	FIGURES	DEPTH (m)	SAMPLES	LITHOLOGY	SEDIMENTARY STRUCTURES	MARLSTONE BEDS	MARLSTONE PERCENTAGE	SHORT DESCRIPTION
Middle Devonian Eifelian	Pärnu Stage Pärnu Formation	26, 27	T-13, T-14	199.0	X MG; T F Ph G; T F Ph; Ps X MG; Ps X MG		Cross-bedded, in places horizontal (core yield 40%)			Yellowish-, violetish- and pinkish-grey, in places dark grey, weakly to medium-cemented, *fine-* and *medium-grained* **sandstone**. At 205.0–206.0 m claystone was revealed by gamma logging. In places carbonate ooliths and *Trochiliscus* are found
				209.4	X MG F Ph		Indistinctly thin-bedded, platy			Grey, cemented by dolomite, sandy **siltstone** with **claystone** interbeds
			T-15	209.7	T F Ph; F Ph; X MG		Cross-bedded, lowermost 0.4 m in places conglomeratic			Greyish-yellow and pinkish-grey, in places dark grey, medium-cemented, *very fine-grained* and *fine-grained* **sandstone**. Bedding surfaces are covered with mica flakes. The lowermost 0.4 m is violetish-grey, strongly cemented by dolomite and with bluish-grey **claystone** pebbles
		28	T-16	213.0	T F Ph		Cross-bedded, in places horizontally thin-bedded (core yield 85%)			Greyish-brown and pinkish-grey, medium-cemented, *fine-* to *medium-grained* **sandstone**. In places *Trochiliscus* is found. The lowermost 0.2 m is bluish-grey dolomitic **claystone**
			T-17, T-18	214.0	X G; F Ph; T F Ph X MG; F Ph; T F Ph; Ps; X MG		Cross-bedded, (core yield 70%)			Violetish-, yellowish- and pinkish-grey (in places grey and dark grey), medium-cemented, *medium-* to *fine-grained* **sandstone** (uppermost 0.8 m greyish-brown and *very fine-grained*). Bedding surfaces are covered with mica flakes. At 214.8 m occurs a grey platy inclined **claystone** interbed (thickness 1 cm)
Lower Devonian Emsian	Rēzekne Stage Mehikoorma Formation	29		220.3	K; Ms X MG		Horizontally thin-bedded, rubbly (core yield 65%)	Up to 3 cm; IND greenish-grey	100	Greenish-grey, silty **dolomitic marlstone**. Clay and dolomite content changes vertically. Burrows and siltstone-filled pockets are present Intercalation of greyish-brown **sand-** and **siltstone** with greenish-grey **dolomitic marlstone** and thin grey **claystone** interbeds
				224.0			Rubbly, not observable		?	
		30	T-19	224.5	F Ph Ps; X MG; F Ph X MG; T F Ph F Ph		Horizontal and wavy, medium- to thin-bedded, dolostone horizontally thin-bedded	Up to 3 cm; IND mainly bluish- and greenish-grey	90	Bluish- and greenish-grey (lowermost 1.3 m light grey), sandy and silty **dolomitic marlstone** with **siltstone** interbeds, clay films and burrows filled with sand and silt. Clay, silt and sand content changes vertically. At 225.3–225.45 and 228.0–228.3 m are yellowish-grey, silty and sandy **dolostone** interbeds with rare vugs (up to 1 cm in size)
				228.3						



STANDARD UNITS	LOCAL STRATIGRAPHIC UNITS	CORE BOX NO.	FIGURES	DEPTH (m)	SEDIMENTARY STRUCTURES	MARLSTONE BEDS	MARLSTONE PERCENTAGE	SHORT DESCRIPTION
Upper Ordovician (Caradoc)	Räga*	40	T-28, D-10	284.8, 286.4	Wavy, indistinctly thin- to medium-bedded, in places nodular	< 0.2, 0.2–3 cm; D (IND) grey, dark and greenish-grey	10–20	Light grey (in places with brownish-grey tinge), *very finely crystalline* **limestone** with **marlstone** and rare **dolostone** interbeds. The uppermost 25 cm is dark grey *medium-crystalline* **calcitic dolostone**. Clay content changes vertically
	Oandu Stage, Variku Formation; ?	41, 42	D-11, T-29; Appendix 4		Massive, in places indistinctly medium- to thick-bedded, nodular, lens-shaped and bioturbated	5–80 cm; IND greenish-grey	50–90	Greenish-grey, in places silty and sandy, **calcitic** and **dolomitic marlstone** with light greenish-grey argillaceous **limestone**, **dolostone** and clayey **siltstone** interbeds. The content of bioclasts (< 10%, in some layers up to 30%), silt, sand, clay and carbonate changes vertically. In places fine pyritized lines (bioturbation?) occur. At 292.7–296.8 m pyritized crusts (thickness 0.1 cm) are found along the bedding surfaces. Rare indistinct dark grey stripes occur below 295.1 m
	Keila Stage, Kahula Formation	43		298.5	Medium-bedded and indistinctly nodular	Up to 10 cm; IND, lowermost 2 m D (dark) greenish-grey	20–30	Light greenish-grey, medium- to strongly argillaceous, in places dolomitized *very finely crystalline* **limestone** (grains < 10%, in places 10–20%) with **calcitic** and **dolomitic marlstone** interbeds. The lowermost 2 m includes slightly to medium-argillaceous **limestone** interbeds (nodules) with burrows and fine pyritized bioclasts
		44		302.6	Wavy, medium-bedded, indistinctly thin- to medium-nodular, in places bioturbated	Up to 3 (4) cm; IND dark greenish-grey	< 10; < 20	Light greenish-grey, in places slightly to strongly argillaceous, *very finely crystalline* **limestone** (grains < 10%) with **calcitic marlstone** (grains in places 10–20%) interbeds. In the upper part fine pyritized bioclasts occur. **K-bentonite claystone** (thickness 2 cm) lies on the lower boundary
	Haljala Stage, Jõhvi Substage		T-30	306.7	Medium-bedded indistinctly thin- to thick-nodular	Up to 3 cm; IND dark greenish-grey	5	Light (greenish-) grey, in places slightly to strongly argillaceous, *very finely crystalline* **limestone** (grains < 10%; in places 30%, rarely 50%) with dolomitized **calcitic marlstone** interbeds. Fine bioclasts are often pyritized. A marlstone interbed with biotite flakes and limonitized pockets lies at 306.7 m
	Idavere Substage, Tatruse Fm.	45		311.0, 312.2	Medium-bedded, medium-nodular	Up to 3 (7) cm; D dark greenish-grey	60	Light (greenish-) grey, in places slightly to medium-argillaceous, *very finely crystalline* and *finely crystalline* **limestone** (grains < 10%, in places 20%) with argillaceous **dolomitic marlstone** and **calcitic marlstone** interbeds.
							5	☞ follow down

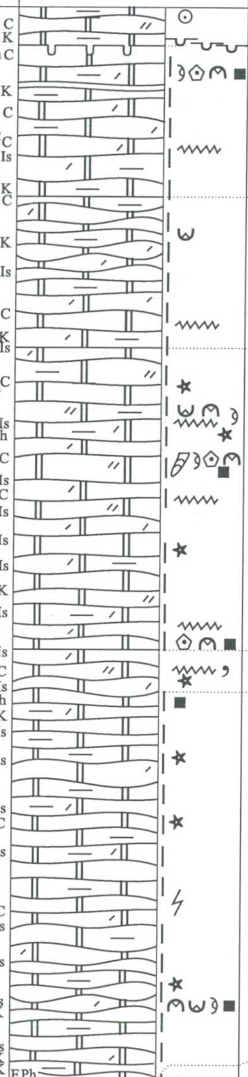
Räga*– Rägavere Formation

STANDARD UNITS	LOCAL STRATIGRAPHIC UNITS	CORE BOX NO.	FIGURES	DEPTH (m)	SAMPLES	LITHOLOGY	SEDIMENTARY STRUCTURES	MARLSTONE BEDS	MARLSTONE PERCENTAGE	SHORT DESCRIPTION
Upper Ordovician (Caradoc)	Haljala Stage, Idavere Substage, Tatruse Formation	45, 46	T-31	317.9, 318.8	CCh, C, CCh, CK, FPh CCh, C, CCh, CCh FPh C, CCh		Wavy, indistinctly thin- to medium-bedded, in places nodular or with microbedded intervals (thickness up to 3 cm)	< 0.2, up to 1 cm; D dark grey	5, < 10	Light grey, in places argillaceous, *medium-* to *finely crystalline*, in places *very finely crystalline* **limestone** (grains 10–25%, in places 30%) with **calcitic marlstone** interbeds. Clay content increases downwards. The discontinuity surface is limonitized
Upper Ordovician (Caradoc)	Kukruse Stage, Dreimani Formation	47, 48	D-12, D-13, T-32, Appendix 4	330.7	TFPh CCh C, FPh CCh K, CCh, CCh FPh, CCh, CCh K, TFPh CCh, FPh CCh, CCh, FPh CCh, CCh, FPh CChK, CCh		Wavy thin- to medium-bedded, in places indictinctly nodular, marlstone indistinctly microbedded	< 0.2, up to 3 cm; D (IND) dark greenish-grey, in places brownish-grey	10–30	Light grey, in places argillaceous *very finely crystalline* and *micro-crystalline* **limestone** (grains 10–25%, in places 30–40%, often pyritized) with brownish-grey **kerogenous limestone** and **calcitic marlstone** interbeds and pockets. Pinkish-white baryte concretion (diameter 2 cm) is surrounded by pyrite aggregates (diameter up to 0.3 cm)
Middle Ordovician (Llanvirn)	Uhaku Stage, Kõrgekallas Fm.	49		335.7	CCh, FPh CCh, CCh, FPh CCh K, CChK		Wavy, thin- to medium-bedded, in places microbedded and indistinctly nodular or lens-shaped	< 0.2, up to 2 (3) cm; IND and D greenish-grey	10–20	Light grey, in places argillaceous, *very finely crystalline* **limestone** (grains < 10%, in places > 10%, rarely pyritized) with **marlstone** interbeds
Middle Ordovician (Llanvirn)	Uhaku Stage, Väo Formation	50	D-14, T-33	338.8	CCh, FPh CCh, CCh, CCh, FPh CCh T		Wavy, thin- to medium-bedded,	< 0.2, 0.3–2 (3) cm; IND (D) greenish-grey and dark grey	< 5	Light grey, *very finely crystalline* **limestone** (grains < 10%, in places 20%, often pyritized) with **dolomitic** and **calcitic marlstone** interbeds. Indistinct ring-shaped pyritic mottles occur at 334.9–337.4 m. The lowermost 0.8 m is dolomitized, argillaceous and comprises coarse bioclasts. Discontinuity surfaces are pyritized
Middle Ordovician (Llanvirn)	Väo Formation; Las*	51	D-15	341.7	Is CCh, Is CCh K, FPh, C		Wavy, thin- to medium-bedded, in places visually massive	< 0.2, 0.3 cm; IND greenish-grey	< 5	Grey (in the upper part with green tinge), in places argillaceous, *finely crystalline* **dolostone** with **dolomitic** and **calcitic marlstone** interbeds. At 342.3–343.2 m lie interbeds containing iron ooliths (diameter up to 0.5 mm). Discontinuity surfaces are pyritized and haematized (lower one)

Las*– Lasnamägi Stage

MEHIKOORMA (421) DRILL CORE

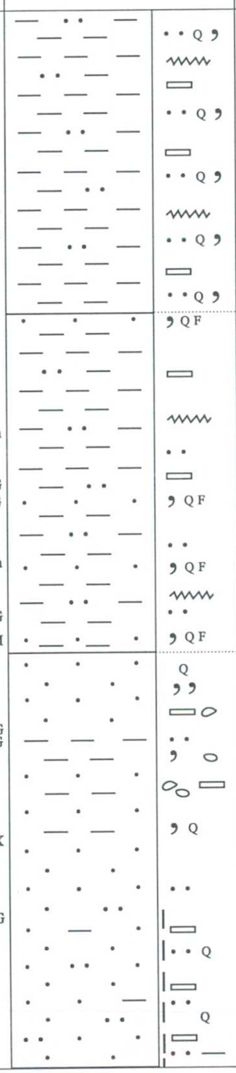
STANDARD UNITS	LOCAL STRATIGRAPHIC UNITS	CORE BOX NO. / FIGURES	DEPTH (m) / SAMPLES	LITHOLOGY	SEDIMENTARY STRUCTURES	MARLSTONE BEDS	MARLSTONE PERCENTAGE	SHORT DESCRIPTION
Middle Ordovician (Llanvirn)	Lasnamägi* Väo Fm.	51, T-34	344.0 C, FPhK, FPhC					☞ follow up
Middle Ordovician (Llanvirn)	Aseri Stage, Lasnamägi*; Stirnas Fm.	52	TFPhC CK, 345.5 C, C, FPhC C Is, 348.0 CK FPhC		Wavy, indistinctly thick- to medium-bedded, in places visually massive	< 0.2, up to 1 cm; IND dark brownish-red	5–10	Brownish-red (at 344.0–345.9 m with greenish-grey interbeds and pockets), *finely* to *very finely crystalline* **calcitic dolostone** (grains < 10%) with **dolomitic marlstone** interbeds. Clay and carbonate content changes vertically
Middle Ordovician (Llanvirn)	Kunda Stage, Segerstad Fm.	T-35, D-16, 53	348.5 TFPhIs CK, Is, Is C, 352.0 FPhCK Is		Wavy, indictinctly thin-nodular, in places medium-bedded	< 0.2, up to 1 cm; IND dark brownish-red	5	Brownish-red (with rare greenish-grey pockets), *very finely crystalline* to *medium-crystalline* **dolostone** (grains < 10%, in places > 10%) with **dolomitic marlstone** and **argillaceous dolostone** interbeds. Clay and carbonate content changes vertically
Middle Ordovician (Llanvirn)	Kunda Stage, Baldone Formation	T-36, T-37, D-17, 54, D-18; Appendix 4	FPhIsC, TFPhIs FPh, TFPhC, Is, C, Is, FPhIs, FPhIs, CK, Is		Wavy, indistictly thin- to medium-bedded	< 0.2, up to 1 cm; IND and D greenish-grey	< 5	Intercalation of brownish-red, light greenish-grey and yellowish-grey *microcrystalline* to *very finely crystalline*, in places *very finely crystalline* to *medium-crystalline* **calcitic dolostone** (grains < 10%, in places 10–25%) with **dolomitic marlstone** interbeds. Rock is in places recrystallized. Characteristic are uneven goethitized or limonitized surfaces
Middle Ordovician (Llanvirn)	Kunda Stage, Š*	T-38, D-19	360.0 TFPhIs, C, Is, 361.1 TFPh		Wavy, indistinctly thin- to medium-bedded	< 0.2, up to 1 cm; IND dark grey	< 5	Grey (in places with green tinge), *very finely crystalline* and *finely crystalline* **dolostone** (grains < 10%, in places 10–25%) with **dolomitic marlstone** interbeds. Clay and carbonate content changes vertically. Glauconite grains are up to 0.3 mm in diameter
Middle Ordovician (Arenig)	Volkhov Stage, Kriukai Formation	T-39, 55, T-40, D-20, 56, T-41	FPhCK Is, Is, FPhIs, TFPhC, Is, FPh IsC Is, FPhIs, TFPhIs CK		Wavy, medium-bedded and indistinctly thin-nodular, rarely thin-bedded	< 0.2, up to 1 cm; IND and D dark grey	< 5	Brownish-red (with rare beigish- and yellowish-grey pockets and interbeds), *finely* to *very finely crystalline*, in places *finely* to *medium-crystalline* **calcitic dolostone** (grains < 5%, in places 10–25%, at 369.00–371.38 m > 10%) with **dolomitic marlstone** and **argillaceous dolostone** interbeds. Clay and carbonate content changes vertically. Below 370.4 m occur wavy and horizontal, yellowish-grey goethitized discontinuity surfaces
Middle Ordovician (Arenig)	Volkhov Stage, Kriukai Formation	57, D-21	Is C K, 371.38 IsCK, 371.4 FPh		Not observable			Greenish-grey (with violet and yellow pockets), *finely*? *crystalline* **glauconitic dolostone**. Glauconite grains (up to 20%) are up to 0.5 mm in diameter. The sharp wavy yellow discontinuity surface is goethitized
L. O*	HB*, Z*							



L. O*– Lower Ordovician; HB*– Hunneberg–Billingen stages; Z*– Zebre Formation; Š*– Šakyna Formation; Lasnamägi*– Lasnamägi Stage

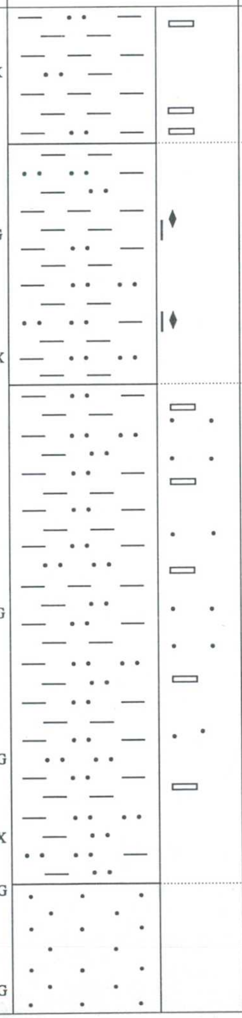
STANDARD UNITS	LOCAL STRATIGRAPHIC UNITS	CORE BOX NO.	FIGURES	DEPTH (m)	SAMPLES	LITHOLOGY	SEDIMENTARY STRUCTURES	MARLSTONE BEDS	MARLSTONE PERCENTAGE	SHORT DESCRIPTION
Furongian (Upper Cambrian) Paibian Stage	Petseri Formation	57	Appendix 4	371.4 – 374.3	A C G X; A F Ph; A X G; F Ph C Ac; A X MG		Horizontally micro-bedded, uppermost part in places cross-bedded (core yield 65%)			Grey (uppermost 0.9 m (ochre-) yellowish-grey and violetish-grey), weathered, silty and sandy **claystone**. In the upper part uneven weathered surfaces are marked by iron oxides
				374.3 – 379.4	A; A MG		Uppermost 0.2 m cross-bedded and thin-bedded (core yield 35%)			Light grey, with beige tinge, in places silty, *fine-* and *medium-grained* **quartz sandstone**. Grains are sub-rounded. The uppermost 0.2 m is medium-cemented *medium-grained* **sandstone**. Numerous valves and fragments of lingulates are present
Middle Cambrian	Paala Formation	58; 59		379.4	MG; MG; MG; MG		Not observable (core yield 50%)			White, *fine-* and *medium-grained* **quartz sandstone**. Grains are sub-rounded. A clayey medium-cemented *fine-grained* **sandstone** interbed (thickness 10 cm; grains angular) lies at 388.0? m. Below 401.5 m in places *coarse* **siltstone** lumps and whitish-grey *fine-grained* **sandstone** interbeds with muscovite flakes are found

STANDARD UNITS	LOCAL STRATIGRAPHIC UNITS	CORE BOX NO.	FIGURES	DEPTH (m)	SAMPLES	SEDIMENTARY STRUCTURES	MARLSTONE BEDS	MARLSTONE PERCENTAGE	SHORT DESCRIPTION
Middle Cambrian	Paala Formation	59			MG, G				☞ follow up
		60			MG	Horizontally (indistinctly) microbedded			White (in the upper part in places violet), clayey (illite and kaolinite) strongly cemented **siltstone** (grains sub-angular). Bedding surfaces are covered with muscovite flakes
				413.0	X	Cross- and thin-bedded (core yield 60%)			White, medium-cemented, *fine-grained* (in places *medium-grained*) **sandstone** (dull grains are well rounded). Feldspar grains are missing
				413.7	XG, GM				
Lower Cambrian	Lontova Stage Lontova Formation Mahu Member	61		414.2	GX, XG, FPh, X, FPh, XG	Indistinctly microbedded (core yield 70%)			Intercalation of dirty ochre-yellow and violetish-brown, silty **claystone** (mainly illite) containing distinct pyritized fine burrows with red contours. At 416.55 and 416.7 m occur brownish-grey layers (thickness 1–3 cm). The uppermost 0.2 m (whitish-grey with *Platysolenites antiquissimus*, muscovite flakes and **siltstone** films) and following 0.3 m (yellowish-grey with light grey spots 1–2 mm in diameter) is weathered, silty **claystone** (illite and kaolinite)
		62, 63		416.7	XG, FPh, XG, FPh, XG	Indistinctly microbedded (core yield 90%)			Greenish-grey **claystone** (mainly illite), in the upper part with dark yellowish-grey pyritized burrows (diameter up to 1 and 1.5–2 mm) and violetish-grey spots (diameter 3–4 mm). Silt content changes vertically and increases at 423.0–428.0 m. In the lower part bedding surfaces are covered with mica flakes and films of organic matter are present. At 418.5–421.0 m hornlike chitinous? sclerites (*Viitnaella*) and throughout the interval *Platysolenites antiquissimus* and *Yanichevskyites petropolitanus* fragments are found

STANDARD UNITS	LOCAL STRATIGRAPHIC UNITS		CORE BOX NO.	FIGURES	DEPTH (m)	SAMPLES	LITHOLOGY	SEDIMENTARY STRUCTURES	MARLSTONE BEDS	MARLSTONE PERCENTAGE	SHORT DESCRIPTION
Lower Cambrian	Lontova Stage Lontova Formation	Mahu Member	63 64		437.0	FPh XG FPh XG					☞ follow up
		Sämi Member	65	T-42	446.0	FPh XG MG TFPh XG M		Indistinctly microbedded (core yield 70%)			Greenish-grey, silty (particularly in the lower part) **claystone** (mainly illite) with dark yellowish-grey pyritized fine burrows and scattered muscovite flakes. In the upper part bedding surfaces are covered with gravel and glauconite grains. At 437.0 (*coarse-grained*), 442.0 (*very fine-grained*), 443.6 (well-rounded, *medium-* and *coarse-grained*) and 445.6 m (*fine-grained*) occur greenish-grey, weakly to strongly cemented **sandstone** (quartz, feldspar) interbeds (thickness 4–10 cm) with glauconite grains. At 440.2 and 441.7 m *Yanichevskyites petropolitanus* fragments are found
			66		457.0	MG XG X MG		Inclined or horizontal, thin-bedded, rarely thick-bedded, at 449.0–450.0 m micro- and thin-bedded, at 452.8–457.0 lens-like (core yield 10%)			Light grey (with green, in the lower part with yellow and brown shade), medium- and strongly cemented by clay and carbonate, *medium-* and *fine-grained* **quartz sandstone** (grains sub-rounded) containing rare glauconite grains (light and dark coloured, particularly in the upper part) and phosphate? pebbles. Microinterbeds of greenish-grey clayey (kaolinite-illite) **siltstone** occur. The interval 448.0–451.0 m (depths specified by gamma logging) shows indistinct thin intercalation of **claystone** and **sandstone** (with mica and rare gravel; feldspar and glauconite are missing) interbeds. Contacts of layers are burrowed. Trace fossil excavations are 2–3 (horizontal direction) and 3–6 mm (vertical direction) wide and filled with sandy and silty sediments. The lower boundary of the interval is sharp. At 452.8–457.0 m lens-like intercalation of mottled **siltstone** and **claystone** with *Platysolenites antiquissimus*, *Sabellites cambriensis* and rare *Yanichevskyites petropolitanus* fragments is observed

STANDARD UNITS	LOCAL STRATIGRAPHIC UNITS	CORE BOX NO.	FIGURES	DEPTH (m)	SAMPLES	LITHOLOGY	SEDIMENTARY STRUCTURES	MARLSTONE BEDS	MARLSTONE PERCENTAGE	SHORT DESCRIPTION
Ediacaran (Upper Vendian)	Voronka Fm. Sirgala Member	66	T-43 D-22, 23	457.0	XG X TFPh X XG FPh		Horizontally thin- to microbedded (core yield 30%)			Whitish-grey (in places violet and with yellow or green shade), clayey (illite-kaolinite), strongly cemented **siltstone** with rare *fine-grained* **sandstone** and **claystone** interbeds. Bedding surfaces are covered with mica flakes. The lower part includes **quartz sandstone** (*coarse* to *fine* grains are sub-angular) and sandy **siltstone** interbeds
	Kotlin Stage Kotlin Formation Meriküla Member			461.4	XG X		Indistinctly microbedded (core yield 30%)			Beige (ochreous), weathered, silty **claystone**, in the lower part with violet spots, at 464.9–465.1 m grey
		67 68 69		465.4	XG X XG XG		Indistinctly micro- and thin-bedded, in places lenticular (core yield 85%)			Greenish-grey, in the upper part dark grey, silty **claystone**. The uppermost 2 m is rich in films of dark brown organic matter and contains vendotaenids. At 470.9, 474.0, 477.5 and 478.5 m occur siderite-containing interbeds (1–10 cm thick) with silt and rare sand grains, often yellowish-grey in colour. The lower part includes a grey, strongly cemented (by clay) **quartz sandstone** interbed (thickness 1.5 cm, grains sub-angular) containing pyrite crystals
				483.4			Indistinctly micro- and thin-bedded, in places lenticular (core yield 85%)			Greenish-grey **claystone** (similar to the upper complex). At 484.0, 487.8 and 488.3 m occur yellowish-grey sandy and silty interbeds (thickness 3–10 cm) showing pressure features. Bedding surfaces are covered with mica flakes. The lower part includes a strongly cemented (by clay) fine-grained **quartz sandstone** interbed (thickness 4 cm, grains sub-angular)

STANDARD UNITS	LOCAL STRATIGRAPHIC UNITS	CORE BOX NO.	FIGURES	DEPTH (m)	SAMPLES	LITHOLOGY	SEDIMENTARY STRUCTURES	MARLSTONE BEDS	MARLSTONE PERCENTAGE	SHORT DESCRIPTION
Ediacaran (Upper Vendian)	Meriküla Mb.	70		488.6	X					☞ follow up
	Kotlin Stage Kotlin Formation Jaama Member	71		495.0	XG X		Indistinctly wavy, microbedded and lenticular (core yield 90%)			Grey, in places with yellow shade, silty **claystone** (mainly kaolinite and illite)
		72		508.2	G G X MG		Indistinctly wavy, microbedded and lenticular, in the uppermost 2 m sandstone layers are cross-bedded (core yield 40%)			Grey, silty **claystone** (similar to the upper complex). In the upper part *very fine-grained* **sandstone** interbeds (thickness 0.5–1.0 mm) occur. Bedding surfaces are covered with mica flakes. The lowermost 10 cm is yellowish-grey, silty, strongly cemented, *very fine-grained* (sub-angular) **feldspar-quartz sandstone**
	Gdov Fm.				MG		Not observable (core yield 40%)			Yellow and beige, with brown pockets, clayey, *very fine-grained* (sub-rounded, in some layers *fine-* to *coarse-grained*) **feldspar-quartz sandstone**



STANDARD UNITS	LOCAL STRATIGRAPHIC UNITS	CORE BOX NO.	FIGURES	DEPTH (m)	SAMPLES	LITHOLOGY	SEDIMENTARY STRUCTURES	MARLSTONE BEDS	MARLSTONE PERCENTAGE	SHORT DESCRIPTION
Ediacaran (Upper Vendian)		72				core is missing				☞ follow up
Palaeoproterozoic	South Estonian block	73		514.2	T		Oriented in parallel, in places cataclastic and gneissic (core yield 20%)			Greenish-red, in places crushed and quartzose, in the upper part weathered, *fine-* to *medium-grained* **migmatite granite** with relicts of dark green **gneiss**. Indistinct patches, lens-like and disjointed layers (thickness 1–3 cm) of relicts (15–20%) consist mainly of chloritized biotite. Granite contains quartz (60%), microcline (30%), chloritized biotite, sericite, iron hydroxides and kaolinite
		74		520.2	T T T		Cataclastic, oriented in parallel, gneissic (core yield 45%)			Greyish-red, quartzose, in the upper part slightly weathered, *fine-* to *medium-grained* **migmatite granite** with indistinct biotite bands. Similar to the upper complex, but with less dark-coloured minerals. Granite consists of quartz (60–70%), microcline (15–30%), plagioclase (up to 5%), chloritized biotite (5–10%), muscovite, iron hydroxides and sericite (up to 5%)

STANDARD UNITS	LOCAL STRATIGRAPHIC UNITS	CORE BOX NO.	FIGURES	DEPTH (m)	SAMPLES	LITHOLOGY	SEDIMENTARY STRUCTURES	MARLSTONE BEDS	MARLSTONE PERCENTAGE	SHORT DESCRIPTION
Palaeoproterozoic	South Estonian block	74								☞ follow up
				541.2	T		Cataclastic, gneissic (core yield 45%)			Reddish-grey, quartzose, *fine-* to *medium-grained* **migmatite granite** with bands (thickness 2–5 mm) and lens-shaped relicts (15–20%) containing chloritized biotite (60–70%), quartz and feldspar. Granitoids consist of quartz, microcline, plagioclase and biotite (chlorite). In places thin veins of secondary carbonate occur
					T					
				548.5						

Other issues in the series
Estonian Geological Sections:

Tartu (453) drill core (Bulletin 1; 1998)
Taga-Roostoja (25A) drill core (Bulletin 2; 1999)
Valga (10) drill core (Bulletin 3; 2001)
Soovälja (K-1) drill core (Bulletin 4; 2002)
Ruhnu (500) drill core (Bulletin 5; 2003)

Forthcoming issue:

Kerguta (565) drill core