Hirnantian (latest Ordovician) bio- and chemostratigraphy of the Stirnas-18 core, western Latvia

Linda Hints, Olle Hints, Dimitri Kaljo, Tarmo Kiipli, Peep Männik,
Jaak Nõlvak, and Helje Pärnaste

Institute of Geology at Tallinn University of Technology, Ehitajate tee 5, 19086 Tallinn, Estonia; Linda.Hints@gi.ee

Abstract. Integrated study of the uppermost Ordovician Porkuni Stage in the Stirnas-18 core, western Latvia, has revealed one of the most complete Hirnantian successions in the eastern Baltic region. The interval is characterized by two shallowing upwards depositional sequences that correspond to the Kuldiga and Saldus formations. The whole-rock carbon stable isotope curve indicates a long rising segment of the Hirnantian carbon isotope excursion, with the highest peak in the upper part of the Kuldiga Formation. The bioclast carbon and oxygen curves fit well with the whole-rock carbon data. Micro- and macrofossil data enabled seven combined associations to be distinguished within the Hirnantian strata. The early Porkuni fauna of the *Spinachitina taugourdeaui* Biozone, with pre-Hirnantian affinities, is succeeded by an interval with a *Hindella–Cliftonia* brachiopod association, a specific polychaete fauna, the chitinozoan *Conochitina scabra*, and the conodont *Noixodontus girardeauensis*. The middle part of the Kuldiga Formation is characterized by a low-diversity *Dalmanella testudinaria* brachiopod association, high diversity of scolecodonts, and the occurrence of the chitinozoan *Lagenochitina prussica*. From the middle part of the Kuldiga Formation the youngest occurrence yet known of the conodont *Amorphognathus ordovicicus* is reported. Also typical of the Kuldiga Formation is the occurrence of the trilobite *Mucronaspis mucronata*. The uppermost Hirnantian Saldus Formation contains no shelly fauna, but yields redeposited conodonts and at least partly indigenous chitinozoans and scolecodonts. Palaeontological criteria and stable isotope data enable correlation of the Stirnas section with other Hirnantian successions in the Baltic region and elsewhere.

Key words: Hirnantian, Porkuni Stage, East Baltic, carbon isotopes, chitinozoans, conodonts, brachiopods, lithostratigraphy.

INTRODUCTION

The Ordovician–Silurian boundary interval, embracing the Hirnantian glaciation sequence, has been a focussed topic of research for several decades. The growth and decay of the Gondwanan ice sheet, rapid changes in sea level and ocean geochemistry, and associated mass extinction(s) make it one of the most important geo- and bioevents in Phanerozoic history.

In the East Baltic region, biotic effects of these environmental perturbations have been known for a long time. Hirnantian oxygen and carbon isotope excursions in Baltica were first documented and linked to a glaciation by Brenchley et al. (1994). Since then, numerous geo-chemical and palaeontological studies have been carried out, enabling reconstruction of facies developments, identification of sedimentary gaps in different parts of the palaeobasin, and better understanding of regional biotic effects of the Hirnantian glaciation. A state-of-the-art review and discussion of latest Ordovician event stratigraphy, correlations, carbon isotope excursion, and distribution of brachiopods, chitinozoans, and conodonts in the Baltic region was published by Kaljo et al. (2008). However, notwithstanding the considerable progress made in recent decades, the picture of Hirnantian geo- and bioevents is still far from complete and contrasting opinions exist with respect to the correlation and interpretation of the Baltic succession (Kaljo et al. 2008; Fan et al. 2009; Delabroye & Vecoli 2010; see discussion below).

The most thoroughly studied and commonly cited Hirnantian sections in the Baltic area are those from the Estonian Shelf and the periphery of the Livonian Basin (*sensu* Harris et al. 2004). Data from the central Livonian Basin, where sedimentation was generally more continuous, are less comprehensive, lacking, in particular, an integrated geochemical and palaeontological analysis. Both micro- and macrofossil data have been documented only partly from this part of the palaeobasin, which complicates any comparison and correlation with other regions. Moreover, the distribution and relationships of regional lithostratigraphical units corresponding to the Hirnantian in southern Estonia and Latvia (Kuldiga and Saldus formations, and Bernati, Edole, Piltene, and Broceni members) are still insufficiently understood.

The aim of this paper is to document environmental and biotic changes through the latest Ordovician sequence in the Stirnas-18 core, western Latvia. For this purpose

an integrated study of geochemistry, carbon and oxygen isotopes, and several groups of macro- and microfossils (brachiopods, trilobites, conodonts, chitinozoans, and scolecodonts) has been carried out. The Stirnas section is located in the type area of the above-mentioned litho-stratigraphical units (Fig. 1), providing an excellent opportunity for enhancing regional correlations. The Hirnantian strata in the section are more than 20 m thick, exceeding the thickness in most other wells in Latvia and in the cores studied previously in southern Estonia (e.g. Ruhnu, Valga, Kaugatuma, Taagepera). It is therefore of particular interest to test the completeness of the Hirnantian succession in this section and to compare the fossil assemblages with Hirnantian faunas of other regions and palaeocontinents.

GEOLOGICAL BACKGROUND

The Palaeobaltic (or Baltoscandian) epicontinental sea embracing the western part of the Baltica continent was represented by the Estonian and Lithuanian shelves (Harris et al. 2004; or the North Estonian and Lithuanian facies belts), with the more offshore Livonian Basin (or the Livonian Tongue of the Central Baltoscandian facies

belt) between them. Interpretation of the latest Ordovician succession in the Baltic area has been published recently by, e.g., Kaljo et al. (2001, 2004, 2008) and Brenchley et al. (2003).

The stratigraphical framework for this paper relies on the views summarized by Kaljo et al. (2008). Only rare graptolites are found in the latest Ordovician strata of the Baltic area, so that primarily chitinozoan bio-zones and stable isotopes are used for regional and interregional correlations. The base of the *Spinachitina taugourdeaui* chitinozoan Biozone, coinciding with the rise in carbon isotope values, is taken as the lower boundary of the Porkuni Regional Stage (Nõlvak 1999; Kaljo et al. 2007; Nõlvak et al. 2007). The latter is considered equivalent to the global Hirnantian Stage in the Baltic area.

In some recent publications (e.g. Melchin & Holmden 2006; Fan et al. 2009), however, a rather different correlation between the Porkuni Stage and the Hirnantian has been proposed. These authors suggested that the *S. taugourdeaui* Biozone, and hence the Porkuni Stage, corresponds only to the upper Hirnantian (*Normalograptus persculptus* Biozone), whereas the *Belonechitina gamachiana* Biozone is correlated with the
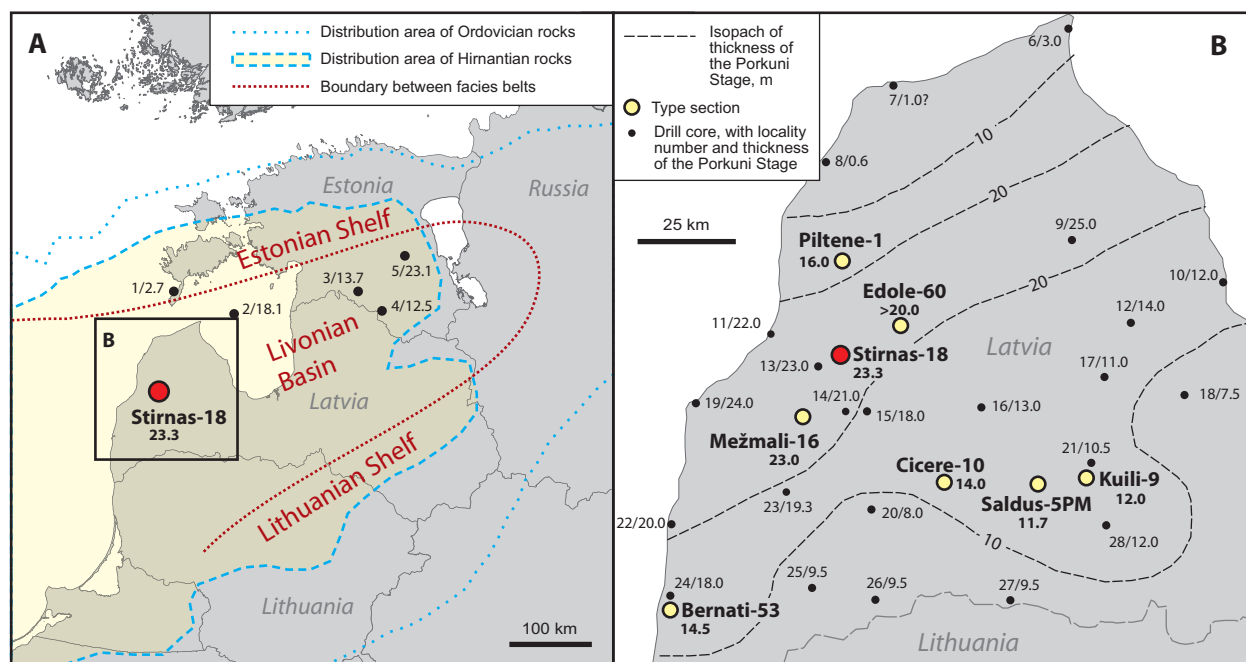

Fig. 1. (A) Location of the Stirnas-18 well and facies belts of the Palaeobaltic basin (Harris et al. 2004; modified). (B) Isopachs of the Porkuni Stage in western Latvia. Stirnas-18 and type sections of the lithostratigraphical units are shown in larger circles. Other localities are indicated by numbers: 1, Kaugatuma; 2, Ruhnu; 3, Taagepera; 4, Valga; 5, Kardla; 6, Kolka; 7, Oviši; 8, Ventspils; 9, Talsi; 10, Engure; 11, Jurkalne; 12, Kandava-28; 13, Riekstini-15; 14, Vilcini; 15, Snepele; 16, Varme; 17, Kandava-25; 18, Degole; 19, Vergale; 20, Skrunda; 21, Remte; 22, Liepaja; 23, Aizpute-41; 24, Bernati-6; 25, Priekule; 26, Vainode; 27, Ezere; 28, Sturi.

lower Hirnantian (*N. extraordinarius* Biozone). Several facts, nevertheless, disagree with such interpretation.

(1) The Hirnantian carbon isotope excursion (HICE) in the stratotype area of the Hirnantian Stage (Chen et al. 2006), supplemented with Katian data from the Honghuayuan section (Zhang et al. 2009), fits well with that in the Baltic area (Ainsaar et al. 2010 and references therein). A fundamental change in the carbon isotope trend (both $\delta^{13}C_{carb}$ and $\delta^{13}C_{org}$ data) occurs at the Katian–Hirnantian boundary (Finney et al. 1999; Brenchley et al. 2003; Melchin & Holmden 2006; Zhang et al. 2009). The pre-Hirnantian part of the curve is rather smooth, with two minor excursions in the Baltic Pirgu Stage (Kaljo et al. 2007) and Richmondian of the Cincinnati Area (Bergström et al. 2007), and even smoother in Anticosti (Long 1993). The Hirnantian part, corresponding to the HICE, is marked by much higher values and considerable variation that reflect strong environmental perturbations. This difference can serve as an efficient chemostratigraphic marker for the Katian–Hirnantian boundary as well as the Pirgu–Porkuni boundary.

(2) Graptolites found together with *S. taugourdeaui* from the upper Lousy Cove Member of the Ellis Bay Formation in Anticosti were recently re-examined by J. Riva (Achab et al. in press). He did not recognize any specific late Hirnantian graptolite species there, which indicates that the *S. taugourdeaui* Biozone in Anticosti may well be of early Hirnantian age. This is in good agreement with chemostratigraphic and brachiopod data (for further details see Kaljo et al. 2008).

(3) The zonal chitinozoan *Belonechitina gamachiana* co-occurs in the uppermost Pirgu Taučionys Formation (Kaljo & Hints 1996; Paškevičius 2000) with the *Holorhynchus* brachiopod fauna (Nõlvak et al. 1989; Brenchley et al. 1997), which, in turn, has been shown to be late Katian in age (Rong & Harper 1988; Sutcliffe et al. 2001; Rong et al. 2004; Koren′ & Sobolevskaya 2008; Suzuki et al. 2009). Moreover, the *B. gamachiana*

Biozone in the Baltic area, as well as in Anticosti, is characterized by low values of the carbon isotope curve, predating the HICE.

These data show that the correlation advocated by Kaljo et al. (2008) is more likely correct than those by, e.g., Melchin (2008) and Fan et al. (2009).

The lithostratigraphy of the topmost Ordovician of the central East Baltic, including western Latvia, has been compiled mainly by Latvian geologists (Ulst et al. 1982, 1984; Brangulis et al. 1989). In the 1960s many wells were drilled in western Latvia in the course of oil exploration, and the rocks of the Porkuni Stage, partially rich in hydrocarbons, were a subject of special interest (Bahtin et al. 1969; Kursh 1977). The Porkuni Stage encompasses the Kuldiga and Saldus formations, both of which are subdivided into two members: Bernati and Edole, and Piltene and Broceni, respectively (Fig. 2; Kaljo et al. 2001). The type sections of these units were initially chosen from drill core sections located in western-most Latvia (Figs 1 and 3). However, because most of these cores were not preserved, neostratotypes were defined in the Saldus-5PM core (Brangulis et al. 1989), which is located east of the holostratotypes, in an area of restricted thickness of the Porkuni Stage (Fig. 1). Data on the properties of Hirnantian rocks (mineral composition, grain size, distribution and morphology of bio- and lithoclasts) in Latvian and southern Estonian sections are recorded in a number of earlier publications (e.g. Bahtin et al. 1969; Volkolakov & Springis 1969; Oraspõld 1982, 1986; Ulst et al. 1982, 1984; Kaljo et al. 2001).

The Stirnas well is within the limits of a relatively narrow, SW–NE aligned belt in westernmost Latvia, where the thickness of the Porkuni Stage is over 20 m (Fig. 1; see also Ulst et al. 1984), exceeding that in most other parts of Baltica (an exception is the Kardla core in SE Estonia; Kaljo et al. 2001, 2008). The Pirgu–Porkuni boundary is represented by a gap in the study area: the
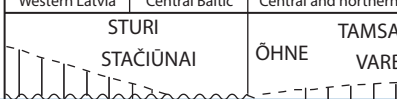
GLOBAL UNIT		REGIONAL STAGE	GRAPTOLITE BIOZONE	CHITINOZOAN BIOZONE	CONODONT BIOZONE	FORMATIONS and members		
						Western Latvia	Central Baltic	Central and northern Estonia
SIL.	Rhudda-nian	Juuru	*P. acuminatus - A. ascensus*	*Belonechitina postrobusta* *Ancyrochitina laevaensis*	*Distomodus kentuckyensis*	STURI STAČIŪNAI	ÕHNE	TAMSALU VARBOLA
U. ORD.	Hirnan-tian	Porkuni	*N. persculptus* *N. extraordinarius*	*Conochitina scabra* *Spinachitina taugourdeaui*	? *Noixodontus* Fauna	Broceni Piltene SALDUS Edole Bernati KULDIGA		ÄRINA
U. ORD.	Katian	Pirgu	*D. anceps* *D. complanatus*	*Belonechitina gamachiana* *Tanuchitina anticostiensis* *Conochitina rugata* *Tanuchitina bergstroemi*	*Amorphognathus ordovicicus*	KUILI PAROVEJA JONSTORP	HALLIKU JELGAVA MOE	ADILA

Fig. 2. Stratigraphical framework (modified from Nõlvak et al. 2006 and Harris et al. 2004). U. ORD. – Upper Ordovician, SIL. – Silurian; vertical ruling – hiatus.
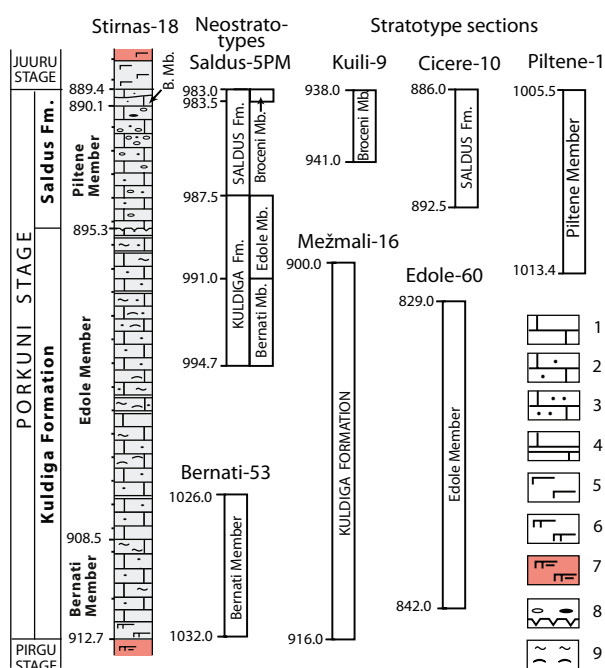
Fig. 3. Comparison of the thicknesses of formations and members of the Porkuni Stage in the Stirnas-18 core and in the stratotype and neostratotype sections after Brangulis et al. (1989). For locations of the sections see Fig. 1B. Legend for the log of the core: 1, limestone; 2, silty limestone; 3, sandy limestone; 4, limestone with interbeds of calcareous marlstone; 5, calcareous marlstone; 6, calcareous dolomitic marlstone; 7, greyish-brown argillaceous dolomitic marlstone; 8, calcareous ooids (empty oval) and clayey pebbles (black oval) (above), discontinuity surface (below); 9, burrows (above), layer with numerous brachiopods (below). B. Mb. – Broceni Member.

topmost part of the Pirgu Stage, corresponding to the Paroveja and Kuili formations, is missing in the Stirnas section, as well as in some neighbouring localities (Ulst et al. 1982; Hints et al. 2005). The upper boundary of the Porkuni Stage, corresponding to the Ordovician–Silurian boundary, is marked by an erosional surface (Ulst & Gailite 1970), which is overlain by lower Llandovery marlstones.

MATERIAL AND METHODS

The Stirnas-18 well was drilled in the Kuldiga district (geographical coordinates 56°55.0′N, 21°33.6′E) in 1965. The core was sampled by Estonian geologists in 1966. More than 100 samples were collected from the Porkuni Stage and adjacent strata, which form the basis of this study. Sampling depths are given in accordance with the depths of drilling intervals (i.e., not corrected using geophysical logging data).

In 2008–09, the samples were restudied and analysed in the Institute of Geology at Tallinn University of Technology by applying different palaeontological, lithological, and geochemical methods. The lithological description (units 1–10) is based largely on the collected samples (among them 31 samples with polished surfaces), data from chemical analyses, and an earlier lithological description of the core by L. Põlma (Figs 4, 5).

The chemical composition of 50 samples was analysed by standard X-ray fluorescence (XRF) methods from pressed powders in a Bruker S4 spectrometer. Major element concentrations were calculated to rock-forming components after Kaljo et al. (1997) with modifications. The dolomite was derived as $(MgO - x) \times 4.57$, where $x = MgO$ in terrigenous matter $= 0.139 \times Al_2O_3$. Calcite was calculated as $(CaO - y) \times 1.79$, where $y = $ dolomite $\times 0.304$. Terrigenous matter was calculated as: $SiO_2 + TiO_2 + 1.139 \times Al_2O_3 + Fe_2O_3 + Na_2O + K_2O + (LOI - z)$, where $z = CO_2$ in calcite $+ CO_2$ in dolomite $+ CO_2$ in Mn-carbonate. Dolomite, calcite, and terrigenous matter were then normalized to 100%.

Data on carbon isotopes were obtained by whole-rock analyses of 45 samples using conventional methods in a Delta E mass spectrometer. Analytical work was undertaken by T. Martma. Additionally, bioclast $\delta^{13}C$ and $\delta^{18}O$ measured from 19 brachiopod shells from five different levels (unpublished data by J. D. Marshall, Liverpool) were included in the study.

Species-level identification of brachiopods was often complicated because of weak lithification of shells and valves, which could not be detached from the harder rock without breaking them. Microfossils (conodonts, chitinozoans, and scolecodonts) were extracted by standard treatment with buffered acetic acid from 93 samples, some of which were barren. Quantitative data on scolecodonts were obtained by counting the most common element of each species in the sample.

The studied samples, fossil specimens, and preparations are deposited at the Institute of Geology at Tallinn University of Technology, for which the institutional abbreviation GIT is used.

LITHOLOGICAL DESCRIPTION

Rocks of the Porkuni Stage comprise two main lithologies in the Stirnas core – mixed carbonate-siliciclastics in the lower and middle part of the section, and relatively silty to sandy oolitic limestones in the upper part, corresponding to the Kuldiga and Saldus formations, respectively. The studied succession is subdivided from oldest to youngest strata into 10 intervals. Characteristic rock types are illustrated in Fig. 5.
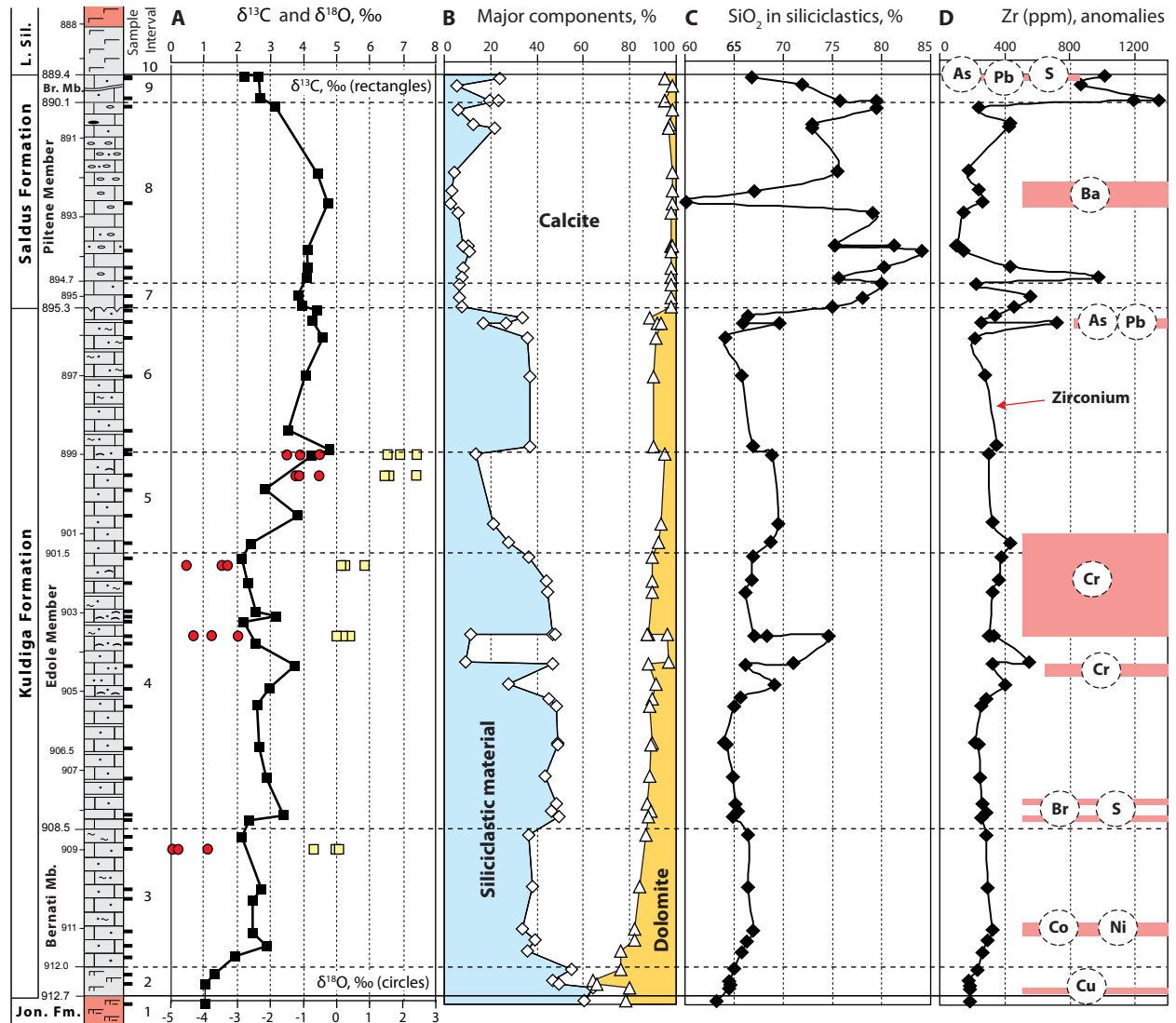
Fig. 4. Lithology and geochemistry of the Porkuni Stage and adjacent strata in the Stirnas-18 core. (A) Whole-rock carbon isotopes (black rectangles), and carbon (empty rectangles) and oxygen (circles) isotopes of brachiopod shells; (B) content of rock-forming components; (C) content of SiO₂ in the composition of siliciclastic material; (D) content of Zr (ppm) and levels of high concentration of selected elements. Numbers 1–10 denote descriptive units of the section. Br. Mb. – Broceni Member; Jon. Fm. – Jonstorp Formation; L. Sil. – Lower Silurian. For lithological legend see Fig. 3.
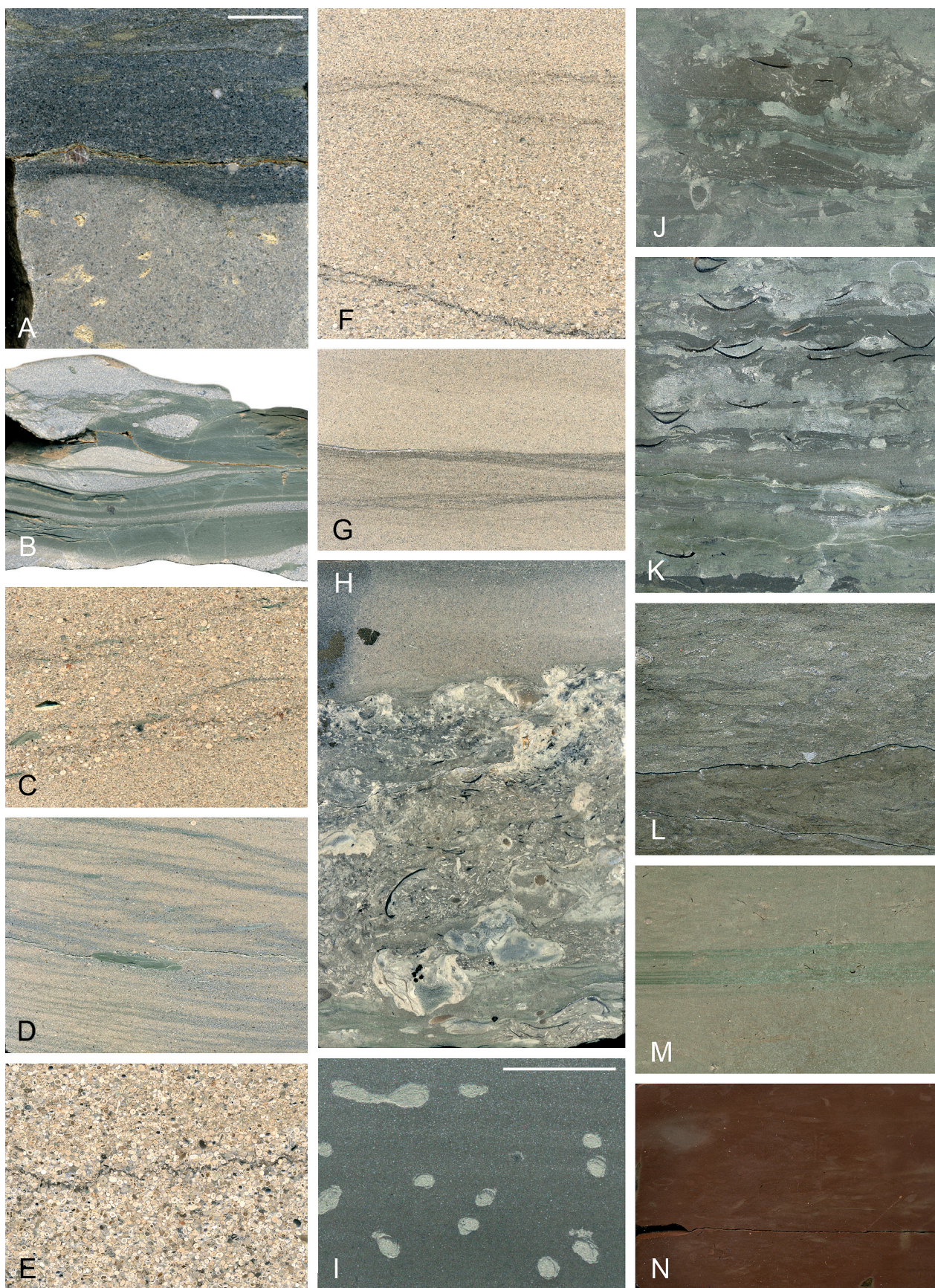
PIRGU STAGE, **Jonstorp Formation** (interval 929.0–912.7 m, thickness 16.3 m).

    1. Reddish-brown argillaceous dolomitic marlstone with greenish-grey interbeds (Fig. 5N).

PORKUNI STAGE, **Kuldiga Formation** (interval 912.7–895.3 m, thickness 17.4 m).

Bernati Member (interval 912.7–908.5 m, thickness 4.2 m).

    2. 912.7–912.0 m (0.7 m). Greenish-grey massive dolomitic marlstone with a faint purple and greyish-brown tinge and spots in the lower part of the interval. Recrystallized bioclasts and horizontal burrows (diameter about 1 mm) occur.

    3. 912.0–908.5 m (3.5 m). Olive grey, predominantly massive weakly dolomitized silty limestone with sporadic bioclasts and rare burrows. A thin bed (about 7 mm) of dusky yellowish-green marlstone occurs at 911.5 m depth (Fig. 5M).

Edole Member (interval 908.5–895.3 m; thickness 13.2 m).

    4. 908.5–901.5 m (7 m). Lithologically most variable interval in the formation. Greenish-grey to grey silty calcareous marlstone intercalating with silty brownish-grey to grey limestone (Fig. 5J–L). Stromatolite-like microlaminated structures occur at some levels (Fig. 5K); strong bioturbation with burrows up to 2–3 mm in diameter is characteristic.

    5. 901.5–899.0 m (2.5 m). Silty limestone, more calcareous and fossiliferous than the underlying strata. Burrows are distinct, filled with light green argillaceous material (Fig. 5I).

    6. 899.0–895.3 m (3.7 m). Yellowish-grey silty calcareous marlstone with interbeds of grey to brownish-grey fine-crystalline limestone. The uppermost 7 cm are highly bioturbated with small (1–2 mm) bioclasts and intraclasts (diameter up to 2 cm). Dispersed fine pyrite crystals and their aggregates are common. The discontinuity surface at the top of the interval is unevenly eroded and penetrated by burrows (Fig. 5H). It corresponds to the boundary between the Kuldiga and Saldus formations.

Saldus Formation (interval 895.3–889.4 m, thickness 5.9 m).

Piltene Member (interval 895.3–890.1 m, thickness 5.2 m).

    7. 895.3–894.7 m (0.60 m). Light grey to grey, indistinctly wavy laminated (cross-bedded?) fine-grained massive limestone rich in dispersed pyrite crystals and irregular aggregates. The amount of siliciclastic material is decreased four-fold by comparison with unit 6 (Fig. 4).

    8. 894.7–890.1 m (4.6 m). Highly variable succession of semihorizontal to cross-bedded silty and sandy limestone with oolitic limestones in the middle part (Fig. 5C–G). Ooids of variable size (commonly less than 1 mm in diameter) are circular or irregularly oval (Fig. 5E), in the upper part of the interval they are arranged in small lens-like interbeds and on distinct cross-bedding planes (Fig. 5D).

Broceni Member (890.1–889.4 m; thickness 0.7 m).

    9. Greenish-grey, wavy laminated marlstone with carbonate and sandy lenses (ripple marks?) and interbeds (Fig. 5B) occurring in the lower part of the member. The top of the member (sample 889.5–889.4 m) (Fig. 5A) is a sandy peloidal(?) limestone rich in fragments of shelly fossils. Clayey pebbles occur above the contact between light and darker grey carbonate sandstones.

Silurian, JUURU STAGE (lowermost part).

    Stačiunai Formation (interval 889.4–887.5 m).

    10. Light grey argillaceous marlstone (889.4–888.1 m) overlain by red argillaceous marlstone (888.1–887.5 m).

Fig. 5. Polished rock samples of the Stirnas-18 core; all at natural size, except A and I where 1 cm scale is shown. **A, B**, Broceni Member; **C–H**, Piltene Member; **I–L**, Edole Member; **M**, Bernati Member; **N**, Jonstorp Formation (Pirgu Stage). **A**, light grey silty limestone (below) and dark grey calcareous sandstone (above), depth 889.5–889.4 m, sample GIT 581-35. **B**, laminated marlstone with light grey interbeds and lenses of calcareous sandstone (ripple marks?), depth 889.85 m, sample GIT 581-34. **C**, limestone with unevenly distributed fine fragments of shelly fossils and ooids, weakly developed cross-beddings, depth 890.2 m, sample GIT 581-1. **D**, cross-bedded sandy limestone with a lenticular intraclast of clay, depth 890.7 m, sample GIT 581-2. **E**, oolitic limestone with a stylolite, part of the sample at a depth of 892.48–892.3 m, sample GIT 581-4. **F**, cross-bedded limestone with unevenly distributed fine ooids and bioclasts, depth 892.96–892.85 m, sample GIT 581-8. **G**, limestone with a finely cross-bedded sandy interbed, depth 894.67–894.6 m, sample GIT 581-10. **H**, burrowed limestone with litho- and bioclasts below the discontinuity surface (depth 895.30 m) on the boundary between the Kuldiga and Saldus formations, and fine-grained silty limestone above it, depth 895.38–895.28 m, sample GIT 581-12. **I**, massive burrowed silty limestone, depth 900.75 m, sample GIT 581-14; scale bar corresponds to 1 cm. **J**, burrowed silty limestone, depth 902.95–902.85 m, sample GIT 581-32. **K**, silty limestone with brachiopod valves and stromatolite-like structures, depth 904.94–904.85 m, sample GIT 581-19. **L**, massive burrowed silty limestone, depth 908.01–907.9 m, sample GIT 581-21. **M**, massive dolomitic silty limestone with horizontal interbeds, depth 911.78–911.65 m, sample GIT 581-27. **N**, massive argillaceous dolomitic marlstone, depth 913.05–912.9 m, sample GIT 581-31.

GEOCHEMISTRY

Major components. X-ray fluorescence analyses reveal that the studied rocks in the Stirnas core mostly comprise more than 50% carbonate material (Fig. 4B, Appendix 1). Only the two samples in the Pirgu–Porkuni boundary interval (in units 1 and 2) contain 60–64% siliciclastic material. A shorter interval of pure limestone occurs between 904.3 and 903.6 m (upper half of unit 4). The content of siliciclastic material decreases considerably between 901.5 and 899.0 m (unit 5). Relatively pure limestones with less than 10% siliciclastic material dominate in the Piltene Member (units 7 and 8) of the Saldus Formation above the 895.3 m depth. The content of dolomite decreases upwards through the Porkuni Stage (Fig. 4), being mostly less than 20%, except below 911.5 m (unit 2 and the lowermost part of unit 3) where it is up to 36%.

Substantial variations in the grain size of non-carbonate material are indicated by variations in the SiO_2 content of siliciclastic material (Fig. 4C). Initial rise in SiO_2 at the base of the Kuldiga Formation may indicate decrease in water depth. Variations between 65% and 70% in the upper half of the Kuldiga Formation (units 5 and 6) indicate that silty material is dominant. Higher values in the upper part of the section reflect a substantial input of fine sand, suggesting further lowering of sea level. A lower decrease in sea depth also occurs between 904.3 and 903.6 m (unit 4).

Several components and elements established in the Porkuni Stage of the Stirnas section earn attention by their high values (Appendix 1), indicating changes in the depositional regime.

Zirconium. Sandy limestones in the upper part of the section show high Zr concentrations, reaching 1300 ppm in siliciclastic material (Fig. 4). This is 6.5 times higher than the average for sandstones (Turekian & Wedepohl 1961). High Zr contents possibly indicate terrigenous influx from the weathering of alkaline rocks.

Chromium. In the interval between 904.3 and 901.0 m, the chromium content of siliciclastic material reaches more than 300 ppm, exceeding three times the average of claystones. Elevated Cr contents in the Upper Ordovician of the Aizpute section were explained by terrigenous influx from the weathering of ultramafic rocks obducted during the Caledonian orogenesis in Central Europe and Norway (Kiipli et al. 2009).

Barium. Ba concentrations in oolitic limestones at 892.6–892.3 m reach 533 ppm (Appendix 1), exceeding 50 times the average for limestones and possibly indicating freshwater influx into the basin as hydrocarbonate of Ba dissolved in fresh water precipitates in the form of barite reacting with sulphate in seawater.

Bromine and sulphur. The concentrations of both elements are increased at depths of 908.23 and 907.9 m. This may suggest preservation of some organic matter in sediments at these levels. This suggestion is supported by a slightly darker colour of rocks in this interval. Anoxic decomposition of organic matter in sediments causes fixation of sulphur in the form of pyrite. Elevated concentrations of bromine also in organic-rich sediments like Ordovician *Dictyonema* shale have previously been observed (T. Kiipli et al. 2000).

Arsenic, lead, and sulphur. Higher contents of these elements are recorded at 895.65 and 889.4 m. These levels lie just below discontinuity surfaces and coincide with the occurrences of pyrite.

Cobalt, nickel, and copper. In greenish dolomitic marlstones in the lower part of the studied interval (912.6–911.0 m; in units 2 and 3) the contents of these elements are high, exceeding the world shale average 4–5 times for Cu and Ni. The Co content of 315 ppm at 911.1 m exceeds the shale average ca 15 times. The high Cu content directly above red-coloured mudstones may have a genetic analogy with chalcopyrite occurrences in Silurian red volcanic ash beds in Estonia and Latvia (E. Kiipli et al. 2000). The concentration of Co, Ni, and Cu may have been caused by favourable redox conditions between oxic and sulphate-reducing environments, indicated by the green colour of sediments.

CARBON AND OXYGEN ISOTOPES

Carbon isotopes of the Porkuni Stage have been studied up to now in three sections of western Latvia: the Vilcini and Aizpute (only the lower half of the stage; Carden 1995), and Jurmala (Ainsaar et al. 2004) cores. In the Stirnas section the $\delta^{13}C_{bulk}$ trend in the interval 912.7–889.4 m reveals a plateau-like Hirnantian excursion (HICE) with relatively low absolute peak values. From bottom to top it can be described as follows.

(1) A short, rapidly rising limb of the HICE ends with a first peak value of 2.9‰ at 911.5 m, within the basal part of the *Conochitina scabra* chitinozoan Biozone and above the first occurrence of the conodont *Noixodontus girardeauensis*.

(2) The initial limb is followed by a slow, stepwise increase in values of $\delta^{13}C_{bulk}$, with two intermediate peaks at 908.1 m (3.4‰) and 904.3 m (3.8‰). The interval from 903.5 to 901.0 m is characterized by relatively lower values.

(3) At ca 901.0 m a further stepped rise of the excursion is recorded, reaching 4.9‰ at 898.8 m. This peak, located in the upper half of the Edole Member close to the boundary between beds 4 and 5, is the highest

observed in the Stirnas core. Biostratigraphically it occurs just above the last identifiable specimens of *Amorphognathus ordovicicus,* but still within the range of *Noixodontus girardeauensis*. This peak is followed by relatively high $\delta^{13}C_{bulk}$ values, varying between 3.6‰ and 4.7‰ up to ca 892.0 m in the middle part of the Piltene Member.

(4) Subsequently the falling limb of the HICE is identified with $\delta^{13}C_{bulk}$ values decreasing to ca 3‰ in the top of the Piltene Member, and to nearly 2‰ in the top of the Broceni Member. Since no samples suitable for isotope study were available from the Silurian strata, completeness of the isotope record across the Ordovician–Silurian boundary cannot be estimated.

This pattern is complemented and supported by $\delta^{13}C_{brach}$ data, which show slight increase in mean values from 4.8‰ at 909.0 m to 5.6‰ at 901.8 m and a peak of 7.0‰ at 899.0 m (Fig. 4). However, it should be noted that the difference between absolute values of bioclast and bulk rock analyses is 2.3–3.3‰ in the Stirnas core. Previous authors have usually reported a less than 2‰ difference in the HICE interval of the Baltic area (e.g. Heath et al. 1998; Kaljo et al. 2001; Brenchley et al. 2003).

On a broad scale, the $\delta^{13}C$ data from the Stirnas core conform well with earlier data and the model proposed by Brenchley et al. (2003). The rising and falling limbs of the HICE curve are very similar in shape as well as stratigraphical position to those observed in other eastern Baltic sections. The main part of the HICE in the Stirnas core is plateau-like, differing in that respect from excursions typically recorded in condensed successions and from those that are truncated by erosion. A specific feature of the Stirnas excursion is the rather long rising limb formed mainly by its stepped upper part. This may be due to particular depositional settings in this part of the Livonian basin during the early Hirnantian or/and more complete succession as suggested also by the occurrence of certain chitinozoans that are missing in other sections.

Detailed comparisons with numerous Baltic sections published by Brenchley et al. (2003) and Kaljo et al. (2008) reveal aspects of similarity, but also with some differences. Geographically closest to the Stirnas section is the Riekstini borehole (Fig. 1; Brenchley et al. 2003, fig. 8). In that section the $\delta^{13}C$ curve from the Kuldiga Formation matches the Stirnas excursion both in shape and absolute values (but no data are available from the Saldus Formation). In the southern Estonian sections (e.g. Ruhnu, Taagepera, Valga) the highest $\delta^{13}C_{bulk}$ values are encountered commonly in the lower or middle part of the Kuldiga Formation. In the Stirnas section the peak values occur in the upper part of the Kuldiga Formation and continue in the Saldus Formation, resembling the Kardla curve in this respect (Brenchley et al. 2003, fig. 10). It is also noteworthy that the peak in $\delta^{13}C_{brach}$

curve in the Ruhnu section, post-dating the $\delta^{13}C_{bulk}$ peak, occurs in the topmost part of the Kuldiga Formation according to recent reinterpretation of the section (see discussion below). Taking this into account, the HICE excursion of the Stirnas core also fits well with the Ruhnu data. The relatively high values in the oolitic limestone of the Piltene Member may reflect the fact that this interval contains a great deal of reworked material (cf. discussion on conodonts) and the recorded isotope values could derive from pre-Piltene strata.

Considering $\delta^{13}C_{bulk}$ data, it is possible that small variations in plateau-like excursions are to some extent facies-dependent and therefore the position of the main peak may vary. For example, oolitic and bahamitic carbonates show often elevated $\delta^{13}C$ values (Wigforss-Lange 1999). Another plausible interpretation is that the middle part of the Kuldiga Formation in the Stirnas core is more complete or less condensed, and the slowly increasing part of the $\delta^{13}C_{bulk}$ curve (interval 2 above) is stretched by comparison with the above-mentioned Baltic sections.

On a global scale, the Stirnas $\delta^{13}C$ curve shows strong similarities to the HICE in Wangjiawan Riverside, South China, where the rising limb is interrupted by a low, so that two peaks are recorded: a smaller one in the lower Hirnantian (*Normalograptus extraordinarius* Biozone) and a main peak in the middle Hirnantian (*Normalograptus persculptus* Biozone; Chen et al. 2006; see also Kaljo et al. 2008, fig. 12). We suggest that the stepped rising interval from 911.5 to 903.5 m in the Stirnas section corresponds to the first Wangjiawan peak. Similarly, the main peak in Stirnas probably matches the second Wangjiawan peak, suggesting correlation with the basal *N. persculptus* Biozone.

Oxygen isotope analyses ($\delta^{18}O_{brach}$) from the Stirnas core are limited in number and come only from the Kuldiga Formation (Appendix 2). They display a total shift of ca 4‰ and peak values of ca −1‰ at 899.0 m, closely following the $\delta^{13}C$ curve (Fig. 4). Although the trend is similar to those in the Riekstini (Brenchley et al. 2003) and Ruhnu cores (Heath et al. 1998), the absolute values are somewhat lower in the Stirnas core. However, it cannot be excluded that higher $\delta^{18}O$ values are to be found above the currently recorded peak. Leaving local details aside, data from the Stirnas core indicate the occurrence of a strong $\delta^{18}O$ excursion, coinciding with the HICE.

MACROFOSSILS
Brachiopods

Macrofossils, including rhynchonelliformean brachiopods, have been identified in the Stirnas core only in the Kuldiga Formation. There, intervals with a diverse or

rich brachiopod fauna (lingulates are not discussed here) intercalate with those of low-diverse or sparse faunas (Fig. 6). The lowermost highly argillaceous strata of the Porkuni Stage mostly contain only the brachiopod *Eoplectodonta*, and in some other sections (Aizpute-41) also *Kinnella* sp., a widely distributed brachiopod within the Hirnantian fauna (Temple 1965; Rong & Harper 1988; Chen et al. 2000). *Eoplectodonta*, which is one of the brachiopods surviving the end-Ordovician crisis, occurs in some Hirnantian associations but is more common among pre-Hirnantian faunas (Brenchley &

Cocks 1982). Besides brachiopods, the early Porkuni sparse shelly fauna also comprises cephalopods, trilobites, bryozoans, and microgastropods that have been documented elsewhere in the lowermost Kuldiga Formation (Brenchley et al. 2003). This early Porkuni fauna presumably represents a transitional association from a pre-Hirnantian to Hirnantian Fauna.

The earliest Hirnantian brachiopods *Cliftonia* and *Dalmanella* appear about 1 m above the lower boundary of the Porkuni Stage and Kuldiga Formation. The most diverse association, with the characteristic Hirnantian
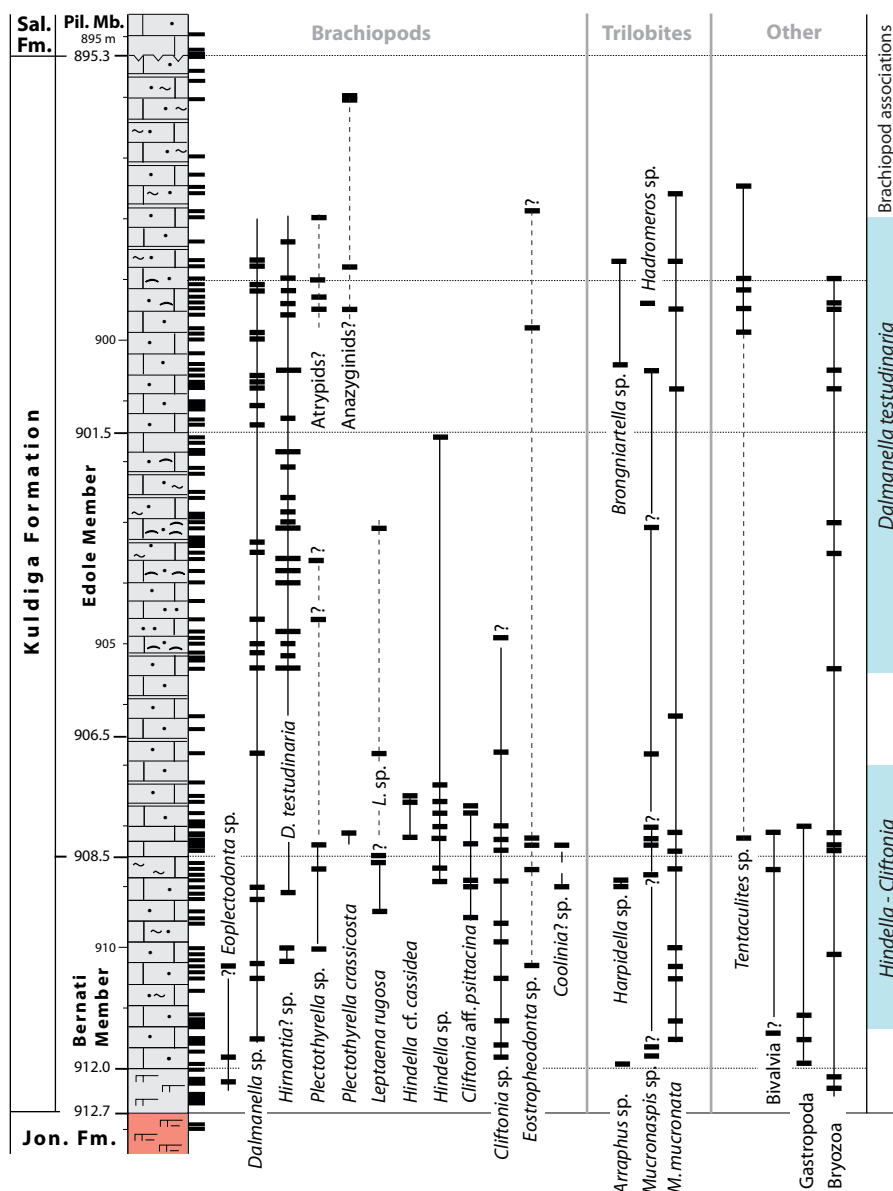

Fig. 6. Distribution of shelly fossils in the Sirnas-18 core. Jon. Fm. – Jonstorp Formation; Sal. Fm. – Saldus Formation; Pil. Mb. – Piltene Member. Dotted horizontal lines mark the boundaries of described lithological units, here and in Figs 7–10. For lithological legend see Fig. 3.

brachiopods *Leptaena rugosa*, *Cliftonia* aff. *psittacina*, *Hindella* cf. *cassidea*, *Plectothyrella crassicosta*, *Dalmanella* sp., and *Eostropheodonta* sp., occurs up to a depth of about 907 m.

The brachiopod assemblage shows similarity with the *Hindella–Cliftonia* Association identified in Norway (Brenchley & Cocks 1982) and with the Hirnantian fauna of southern Sweden (Bergström 1968). The Norwegian associations differ from the Swedish and Estonian faunas in the absence of *P. crassicosta*, which is common in the last two regions. At the same time, *Eospirigerina* and some other brachiopods occur in the Norwegian association, but their occurrence in the easternmost areas is not certain. Because of these differences the Norwegian Hirnantian brachiopod fauna has been considered as atypical (Rong & Harper 1988; Zhan & Jin 2007). The Baltic *Hindella–Cliftonia* Association does not reach the diversity observed in Norway, particularly because of limited material available from the drill core. The disappearance of the *Hindella–Cliftonia* Association at a depth about 907 m coincides with a change in the composition of sediments, marked by continuous increase in siliciclastic material in the upper part of the Bernati Member. The overlying silty limestones (interval 907–905.5 m) contain only few brachiopods. However, the scarcity of shelly fossils in this part of the section could be partly artificial as a result of lower sampling density.

The association dominated by *Dalmanella testudinaria* occurs in the middle part of the Kuldiga Formation (905.5–898.0 m). Brachiopod shells are commonly accumulated at the contacts between silty limestone and more argillaceous strata. Beds with disarticulated or rare brachiopod valves intercalate with beds comprising complete shells of different size. The distribution of brachiopods suggests variable water energy, but the shells seem to be mostly autochthonous. *Dalmanella testudinaria* is associated with few other taxa that are found in older strata (Fig. 6). Only in the uppermost part of the section atrypids and small anazyginids appear, associated with tentaculitids. Most probably the strata with *D. testudinaria* correspond to the Norwegian *Dalmanella* Association, which is a low-diversity fauna occupying a slightly shallower environment than the *Hindella–Cliftonia* Association (Brenchley & Cocks 1982). The succession of these two brachiopod associations in the shallowing upwards sequence supports the opinion of Rong & Harper (1988) that the *Dalmanella testudinaria* Association inhabits more onshore environments than the *Hindella–Cliftonia* Association.

The diversity and frequency of the shelly fauna decreased essentially before the sedimentation gap on the top of the Kuldiga Formation. Samples from the topmost 3 m of the formation yield virtually no brachiopod remains. Silty and oolitic limestones of the Saldus Formation do not contain macroscopic shelly fauna. The *Brevilamnulella* Association, typical of ooidal tidal shoal, and the *Thebesia* Association characteristic of tidal channels, as described by Brenchley & Cocks (1982), have not been identified in the East Baltic sequences.

Trilobites

The first Hirnantian trilobite, *Arraphus* Angelin, appears at a depth of 911.9 m in the Bernati Member. It is a very rare trilobite known previously only at Ålleberg in Västergötland (Angelin 1854) from the *Dalmanitina* Beds. A trilobite fragment (librigena) co-occurring with *Arraphus* belongs to an odontopleurid taxon, probably *Primaspis* Richter & Richter.

Dalmanitids are the most common trilobites in the Kuldiga Formation (Fig. 6). Twenty-seven specimens were counted from the 14.2 m thick interval (911.8–897.6 m), which also includes strata (906.2–900.8 m) with very rare trilobites. All identified dalmanitid specimens are *Mucronaspis mucronata kiaeri* (Troedsson), in which some trends of morphological variation occur throughout its range. Previous morphometric studies of *Mucronaspis* Destombes or its senior synonym *Dalmanitina* Reed (partly) in Baltoscandia have been made by Temple (1952, 1957) and Owen (1981a). However, these authors based their analyses on few specimens collected from different localities and lacking precise temporal constraints. The collection discussed here is of great interest for the observation of gradual morphometric changes through time, and also in the context of environmental changes. However, such analysis is beyond the scope of this paper and will be presented elsewhere. In the Oslo-Asker region, *M. mucronata kiaeri* is associated with both the *Hindella–Cliftonia* and *Dalmanella testudinaria* brachiopod associations (Brenchley & Cocks 1982). The trilobite *Harpidella* sp. indet. also co-occurs with the first brachiopod association at 909.0 and 908.9 m in the lower part of the distribution range of *M. mucronata kiaeri*. *Harpidella s.l.* (revision in progress, see Pärnaste et al. 2009) is a common element in the latest Ordovician trilobite fauna. This group survived the end-Ordovician Extinction and became widely distributed in the Silurian, similar to the dalmanitids.

Two thoracic segments of a homalonotine trilobite were recorded at depths of 898.7 and 900.4 m, most probably belonging to *Brongniartella platynota* (Dalman), but bearing resemblance also to *Platycoryphe* sp. (see Owen 1981b). *Brongniartella platynota* together with *M. mucronata* (Brongniart) are present in the Porkuni Stage of westernmost Latvia (Männil 1966; Ulst et al. 1982). In the Stirnas core *B. platynota* occurs in the upper half of the Kuldiga Formation. This trilobite also occurs at a high stratigraphical position (close to the

lower boundary of the oolitic limestones of the Saldus Formation) in the Ikla core in SW Estonia (Männil et al. 1968). Besides *B. platynota*, a fragment (thoracic pleura) of a new odontopleurid, and a small fragmentary cranidium of a cheirurid, possibly *Hadromeros* sp. indet., were found in the uppermost fossiliferous part of the Kuldiga Formation (Fig. 6).

It should be mentioned that the trilobite genera *Brongniartella* Reed and *Platycoryphe* Foerste, unless congeneric, are strikingly similar, differing mainly in the strength or effacement of certain furrows. However, specimens assigned to *Platycoryphe* sp. in the Oslo Region (Owen 1981b) occur in the interval coinciding also with the upper part of the range of *M. mucronata kiaeri*. Similarly, *B. platynota* and *M.* cf. *mucronata* appear together in the uppermost Ordovician Foel-y-Ddinas Mudstone in the Bala area, Wales (Whittington 1968, p. 121). *Brongniartella platynota* is rather large and is one of the most common trilobites in the *Dalmanitina* Beds of Västergötland, Sweden, but rare occurrences are known from equivalent beds in Scania, Sweden, and the Holy Cross Mountains, Poland, where it is much smaller (Kielan 1960).

MICROFOSSILS

Conodonts

In total, 76 samples were processed, all but three yielding conodonts (Fig. 7). The specimens are generally well preserved and thermally unaltered (CAI = 1), apart from at a depth of 895.30–889.40 m where almost all elements were broken and strongly worn.

Based on ranges and abundance data, three main assemblage intervals can be recognized in the Ordovician part of the studied section. Samples from the Pirgu Stage (three samples studied lie below the interval indicated in Fig. 7) and from the basal part of the Porkuni Stage yielded rich faunas including *Amorphognathus ordovicicus*, *Scabbardella altipes*, *Drepanoistodus suberectus*, *Hamarodus* cf. *europaeus*, *Dapsilodus* spp., *Gamachignathus* sp., and several others. Above 912 m, the abundance of conodont elements drops considerably and all taxa listed above become rare higher up in the section. Up to the top of the Kuldiga Formation most taxa occur sporadically and, as a rule, are represented by only a few specimens per sample. Some taxa are totally absent (*Icriodella* spp., *Oslodus* cf. *semisymmetricus*, *Protopanderodus* sp., *Pseudooneotodus* ex gr. *beckmanni*) or occur in only a single sample (e.g. *Gamachignathus* sp., *Drepanoistodus suberectus*), but reappear in the Saldus Formation. This interval of the Kuldiga Formation is characterized by the occurrence of *Noixodontus girardeauensis*, which appears just above the level of

disappearance of rich Pirgu-type faunas (Fig. 7). Together with the lowermost *Noixodontus girardeauensis*, the oldest specimens of *Ozarkodina* ex gr. *oldhamensis* were found in the studied section. *Amorphognathus* is also one of the most common taxa in this interval, but in most samples the elements are too poorly preserved to allow reliable identification. Nevertheless, in some samples *A. ordovicicus* can be recognized and it is evident that in the Stirnas section this species extends at least to the upper part of the Kuldiga Formation.

The abundance of conodonts increases considerably at the boundary between the Kuldiga and Saldus formations, and most of the taxa that were rare or missing in the interval below then become rather common. Conodonts are most abundant in the uppermost Saldus Formation. However, as the majority of specimens in this formation are represented by strongly worn (rounded) fragments, most of them are certainly redeposited. It is difficult to assess which of them are autochthonous and which are reworked from older strata from elsewhere. Only some well-preserved specimens of *Dapsilodus* spp. from the upper Saldus Formation can be considered autochthonous.

The most distinct change in the conodont faunas takes place at 889.4 m, at the Ordovician–Silurian boundary (Fig. 7). Just below this level the abundance of conodonts reaches its maximum – 1260 specimens per kilogram of rock. In the sample just above this level only a single unidentifiable specimen was found. Representatives of only three lineages (*Pseudooneotodus* ex gr. *beckmanni*, *Dapsilodus*, and *Panderodus*) cross the Ordovician–Silurian boundary in the Stirnas section. At 888.70 m *Decoriconus* sp. n. 1, *Distomodus* sp., and *Walliserodus* cf. *curvatus* are present, and somewhat higher in the section *Dapsilodus praecipuus* and *Decoriconus fragilis* characteristic of the lower Silurian appear.

According to earlier data, identifiable *A. ordovicicus* disappears together with Pirgu-type rich faunas in the lower Bernati Member (e.g. Männik 2001, 2003). The strata above this level have been assigned to an informal unit named the *Noixodontus* Fauna (Nõlvak et al. 2006). The conodont succession in the Stirnas core is as follows: disappearance of the Pirgu fauna in the lower Bernati Member, followed immediately by the appearance of *Noixodontus girardeauensis*; this species is particularly common in the lower part of its range. *Amorphognathus ordovicicus* ranges at least to the upper half of the Edole Member (middle part of the Porkuni Stage), the youngest specimens in the Kuldiga Formation coming from the sample at 900.50 m. The upper boundary of the *A. ordovicicus* conodont Biozone is drawn tentatively just above this level (Fig. 7). The true LAD of *A. ordovicicus* still remains problematic. Although *Amorphognathus* is fairly common also in the upper Kuldiga Formation, its
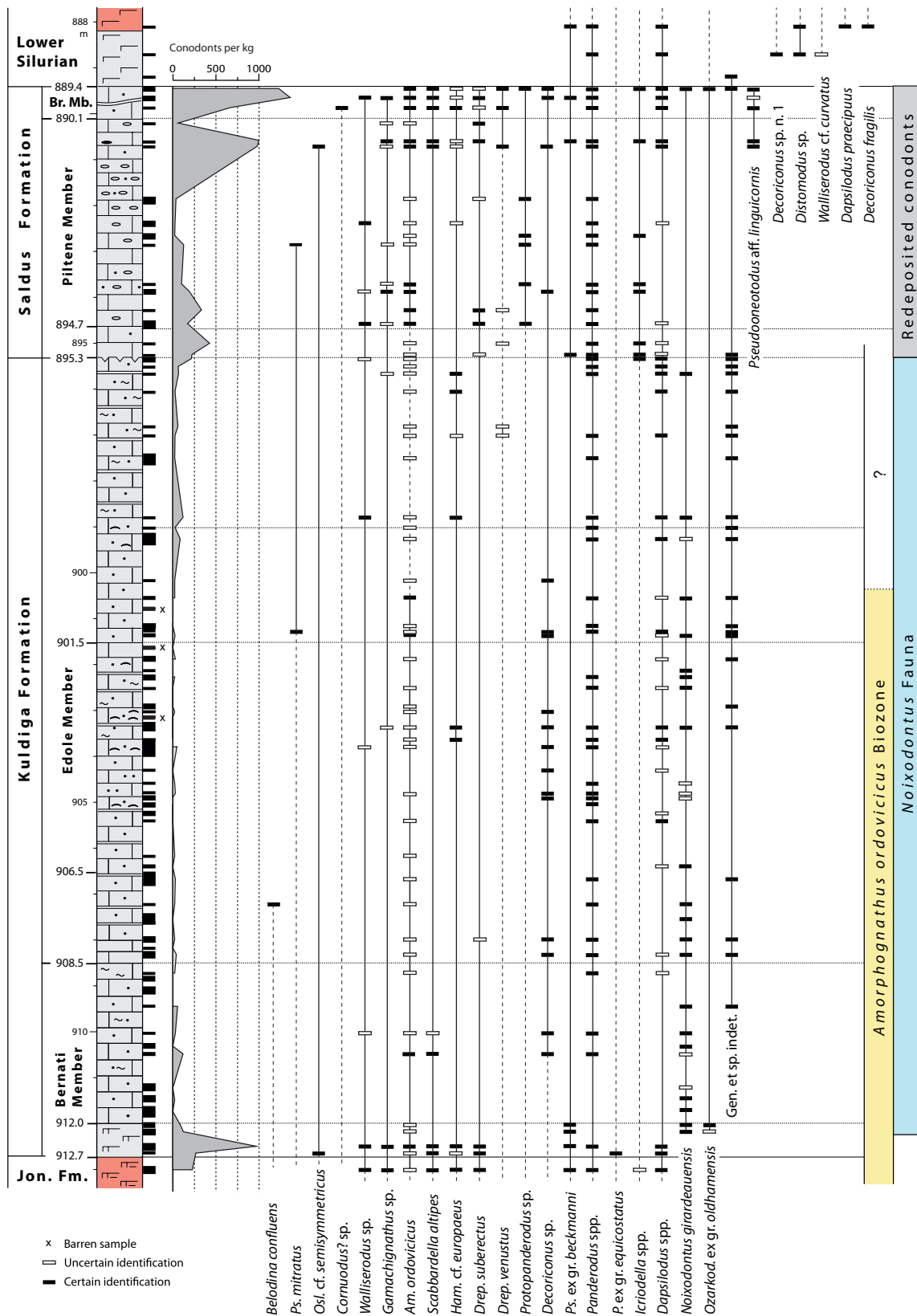
Fig. 7. Distribution and abundance of conodonts in the Stirnas-18 core. *Ps. − Pseoudooneotodus*; *Osl. − Oslodus*; *Am. − Amorpho-gnathus*; *Ham. − Hamarodus*; *Drep. − Drepanoistodus*; *P. − Panderodus*; *Ozarkod. − Ozarkodina*; Br. Mb. − Broceni Member; Jon. Fm. − Jonstorp Formation. For lithological legend see Fig. 3.

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elements are generally too poorly preserved for reliable identification. Identifiable *A. ordovicicus* occurs also in the Saldus Formation, but all specimens there are evidently redeposited and their origin cannot be determined. Except for the longer range of *A. ordovicicus*, the conodont distribution in the Stirnas core agrees well with the data summarized by Kaljo et al. (2008).

Chitinozoans

A total of 66 samples weighing 100–600 g were studied for chitinozoans (Fig. 8), but 24 samples were barren; most specimens are generally poorly preserved and rare. Their abundance remains well below 1 specimen per gram of rock in most of the productive samples. Higher abundance was recorded only in the lowermost Bernati Member (at 912.6–912.4 m) and in one sample from the Broceni Member (at 889.8 m).

The lowermost part of the Bernati Member (912.6–912.4 m) contains *Spinachitina taugourdeaui* and *Acanthochitina* sp. (Brenchley et al. 2003; Kaljo et al. 2008), the indicative species of the *S. taugourdeaui* chitinozoan Biozone and lowermost Porkuni Stage. The same species in the Harelbeke well of the Brabant Massif has enabled correlation of the lowermost Hirnantian strata of Baltica and Avalonia (Vanmeirhaeghe 2006).

In the next interval, overlying the *S. taugourdeaui* Biozone, the diversity and abundance of chitinozoans decrease, similar to many other previously studied sections of the Baltic region (Nõlvak & Grahn 1993; Brenchley et al. 2003; Nõlvak et al. 2006). Rare representatives of *Conochitina scabra* Grahn & Nõlvak, the index species of the topmost Ordovician chitinozoan biozone in Baltica, appear at a depth of 911.7 m.

The interval between 910.0 and 906.5 m contains only long-ranging *Rhabdochitina gracilis*. Similar and probably contemporaneous strata with *R. gracilis* have been documented in the Kętrzyn IG-1 core, NE Poland (Modliński et al. 2002, fig. 3), where the brachiopod *Eostropheodonta hirnantensis* occurs in the same interval.

At 906.4 m *Lagenochitina prussica* appears. This species has been identified previously only in the interval from the Rakvere Stage to the lower part of the Pirgu Stage in numerous East Baltic sections. Its occurrence in the Porkuni Stage of the Stirnas core may indicate that these strata are unknown or absent in the Ruhnu, Valga, Taagepera, and other sections discussed earlier by Brenchley et al. (2003) and Kaljo et al. (2008). Alternatively, it may be argued that more than 100 specimens observed from 15 samples (906.4–896.0 m) are reworked specimens from older strata, within the previously recorded range of *L. prussica*. On other palaeocontinents *L. prussica* has been identified from Hirnantian strata. In Wales rare specimens of
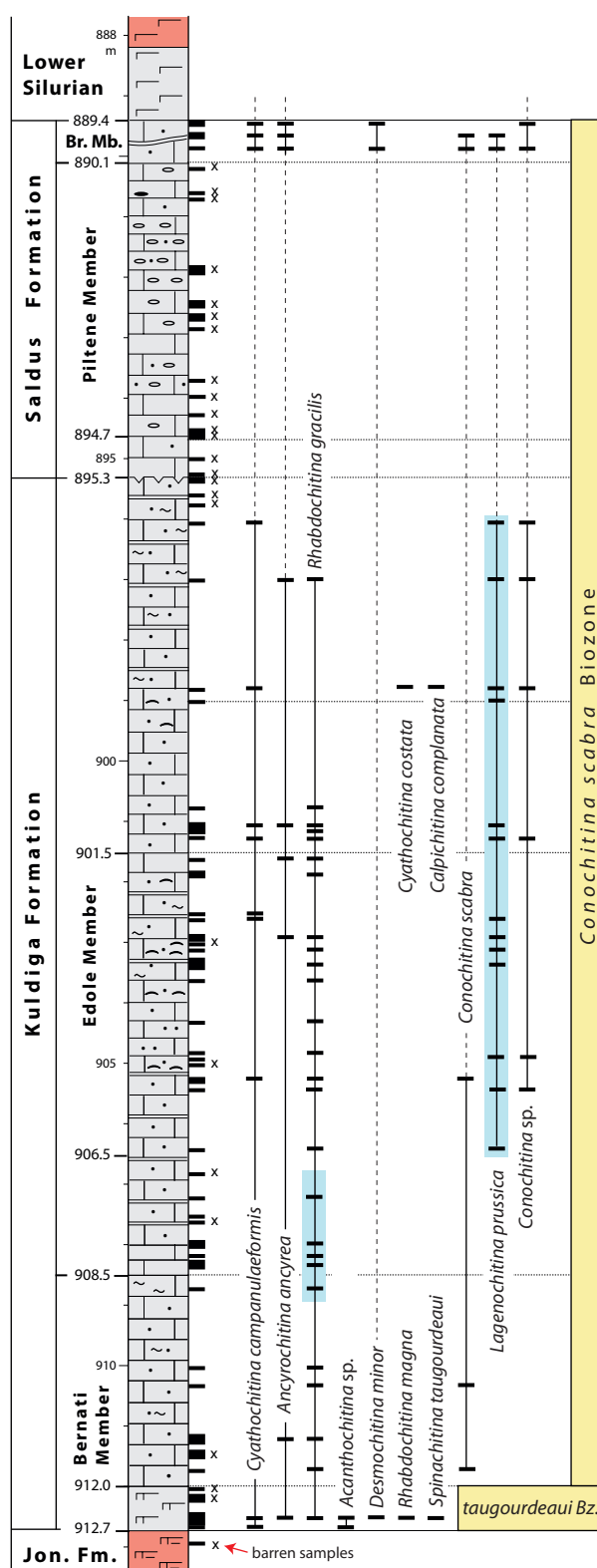

Fig. 8. Distribution and biozones of chitinozoans in the Stirnas-18 core. Br. Mb. – Broceni Member; Jon. Fm. – Jonstorp Formation. For lithological legend see Fig. 3.

Lagenochitina occur in the *S. taugourdeaui* Biozone, but because of its presence, aberrantly high in the stratigraphy, was interpreted as reworked material (Vandenbroucke et al. 2008). The same has been concluded for the occurrence of *Lagenochitina* in the Parc de Sart-Eustache section, Belgium, in presumed Hirnantian strata (Vanmeirhaeghe et al. 2005). Paris et al. (2000) reported *Lagenochitina* also from the Hirnantian of NE Algeria. Thus the highest Ordovician occurrences of *Lagenochitina* are geographically somewhat widespread, ranging through Baltica, Avalonia, and North Gondwana. It is unlikely that all these finds represent reworked material. Instead, the uppermost range of *Lagenochitina* may be useful for interregional correlation. Only a few specimens of *Cyathochitina costata* and *Calpichitina complanata* have been found from the Stirnas core (Fig. 8). The previously known range of these species was limited to pre-Porkuni strata.

All samples from the Piltene Member of the Saldus Formation (895.3–890.1 m) are barren of chitinozoans and other organic-walled microfossils. However, in the Broceni Member (889.8–889.45 m; Figs 5, 8) an unusually rich assemblage of chitinozoans was found, containing *C. scabra*, *L. prussica*, *Cyathochitina campanulaeformis*, and *Desmochitina minor*. All specimens in this interval are strongly flattened, even *Lagenochitina prussica* which possesses some of the thickest vesicle walls known in chitinozoans. This type of preservation may suggest reworking (see discussion on reworked cono- donts above). Therefore it is not clear whether the records of *L. prussica* and *D. minor* in the Broceni Member can be considered as the youngest occurrences of these species from the East Baltic Ordovician.

Consequently, on the basis of chitinozoan distribution the *S. taugourdeaui* and *C. scabra* biozones can be distinguished in the Porkuni Stage in the Stirnas core. The latter biozone can be subdivided further into the following intervals: (1) beds with only *R. gracilis* (910.2–906.5 m); (2) beds with *L. prussica* (906.5–895.3 m); (3) a barren interval of the Piltene Member (895.3–890.1 m), and (4) the Broceni Member embracing a possibly reworked assemblage of chitinozoans, unknown in sections studied earlier (Nõlvak 2001, 2003). If the latter assemblage is not reworked, its close similarity to the assemblage below the barren interval may indicate that the possible maximum shallowing during the glaciation had little effect on chitinozoans. Similarily, Paris et al. (2000, p. 99), studying North Gondwanan material, concluded that the 'glacial event cannot be regarded as being responsible for the extinction of Ordovician chitinozoan fauna'. On a larger scale, however, the extinction of chitinozoans in Baltica began in pre-Hirnantian times (Kaljo et al. 1996; Hints et al. 2009) and continued during Porkuni time as observed in the Stirnas section.

Scolecodonts

Scolecodonts (polychaete jaws) are abundant and diverse microfossils in the Ordovician and Silurian deposits of Baltica (Hints & Eriksson 2007). They are common in the Hirnantian strata of the Stirnas core and, in many cases, the most diverse group of acid-resistant microfossils. The entire assemblage contains poorly to excellently preserved specimens of 35–40 apparatus-based species, including about ten new taxa that are referred to here under open nomenclature. The abundance (specimens/kg) varies considerably from zero to about 1200/kg. The succession of scolecodonts can be divided into five more or less distinct associations (see Figs 9, 10).

Association 1 (about 912.6–912.4 m, basal part of the Kuldiga Formation) is dominated strongly by *Rakvereprion balticus* and *Ramphoprion* cf. *gotlandensis*, but species of *Oenonites*, *Leptoprion*, *Atraktoprion, Protarabellites*, and *Mochtyella* ex gr. *fragilis* are also common. The absolute frequency may reach to ca 200 specimens/kg in this interval. According to Hints (2001), *Rakvereprion balticus* is a long-ranging species, pre- vailing in the Pirgu Stage of the Valga core, southern Estonia. A bloom of *Ramphoprion gotlandensis* has also been observed in the Valga section, but in somewhat older strata. It seems therefore that Association 1 represents continuation of the Pirgu-type fauna and/or environments. This is also supported by the overall composition of the organic-walled microfossil assemblage, where the planktic component (chitinozoans) is strongly prevalent.

Association 2 (about 910.0–906.3 m, lower part of the Kuldiga Formation) is characterized by the appearance of *Pistoprion transitans*, *Mochtyella* cf. *cristata*, xanioprionids, and two distinct species of *Oenonites*, referred to as sp. A and sp. B (as designated by Hints 2001). The absolute frequency is somewhat low in the lower part of this interval, but increases rapidly at 910.0 m and achieves a peak with 1200 specimens/kg at 908.23 m (Fig. 9). This bloom of scole- codonts is due to the high abundance of *Oenonites*, *Mochtyella, Xanioprion, Conjungaspis* and, in particular, *Pistoprion transitans*. This last species may account for up to 60% of all scolecodonts. At ca 907.2 m, the abundance drops rapidly to less than 100 specimens/kg. A largely similar distribution pattern was observed in the Kuldiga Formation in the Valga core, with the abundance peak in the lower part of the formation and the successive predominance of *Oenonites*, *Xanioprion*, *Mochtyella*, and *Pistoprion* (Hints 2001).

Association 3 (about 905.4–900.75 m, middle part of the Kuldiga Formation) is characterized by increased diversity, a predominance of *Kettnerites* and *Oenonites*, and the absence or low percentage of *Pistoprion*.
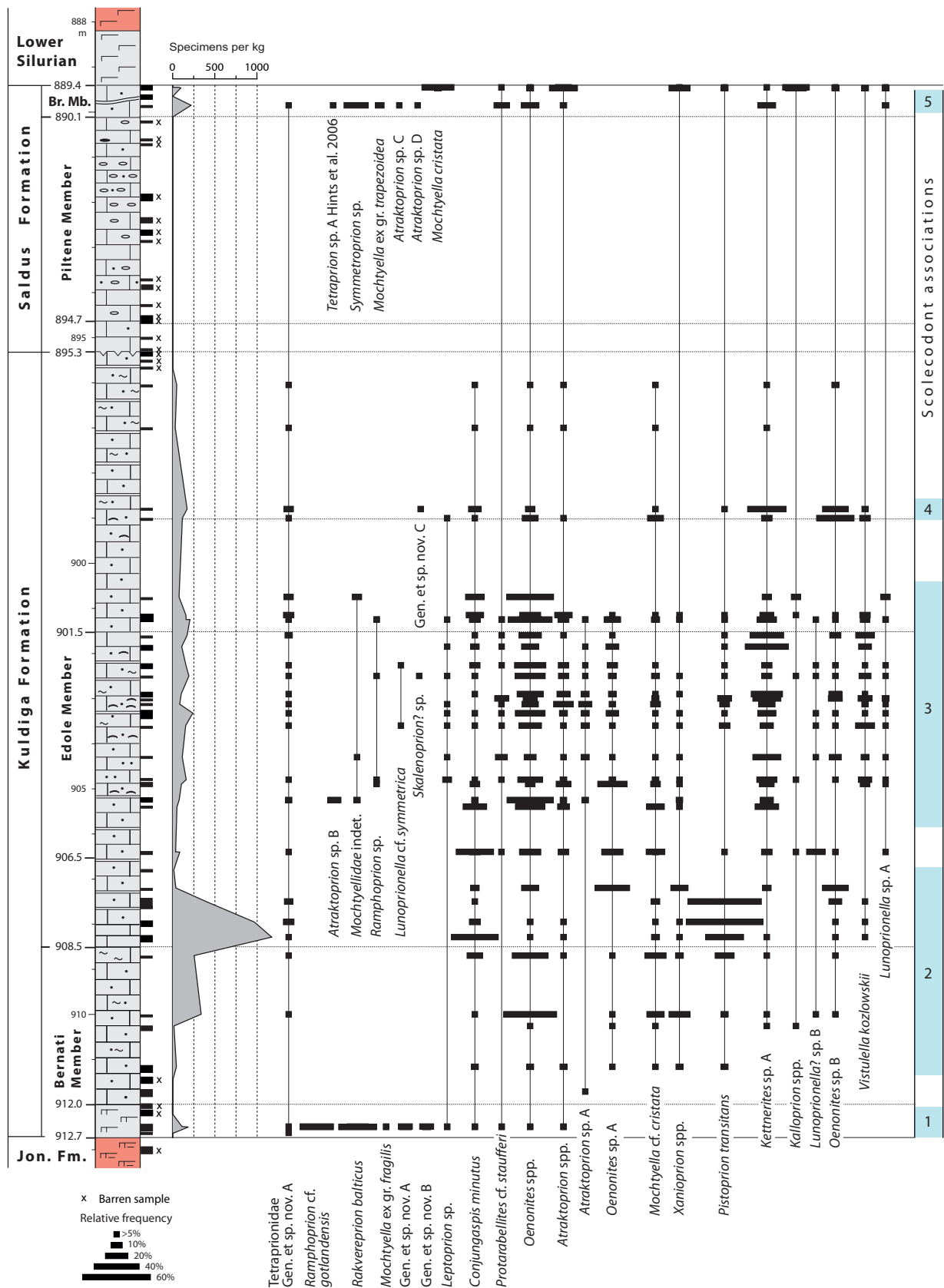
Fig. 9. Distribution, abundance, and relative frequency of scolecodonts in the Stirnas-18 core. For lithological legend see Fig. 3.
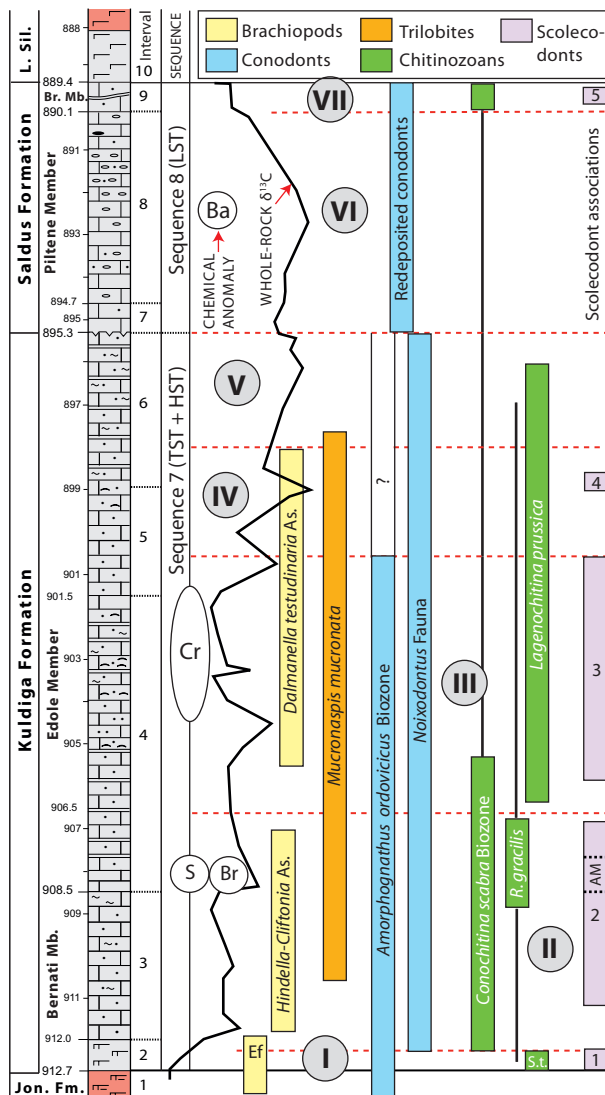
Fig. 10. Integration of sedimentological, geochemical, and palaeontological data of the Porkuni Stage of the Stirnas-18 core. For details see Figs 4, 6–9 and Appendixes 1, 2. Ef – brachiopod fauna with *Eoplectodonta*; I–VII – intervals characterized by distinct faunal associations. S.t. – *Spinachitina taugourdeaui* Biozone; AM – abundance maximum of scolecodonts, Br. Mb. – Broceni Member; Jon. Fm. – Jonstorp Formation. For lithological legend see Fig. 3.

Specimens of *Vistulella*, *Atraktoprion*, *Conjungaspis*, *Mochtyella*, *Leptoprion*, *Protarabellites*, *Lunoprionella*, and Tetraprionidae gen. nov. are also common. The absolute frequency varies between 100 and 200 specimens/kg in this interval. A closely similar association has not been observed in the Valga core. Notably, *Kettnerites* sp. A is only a minor element in the Valga section, whilst *Oenonites* sp. B, predominant in Valga, is much less common in the Stirnas core. This dis-

crepancy may be explained by facies differences or, possibly, by the fact that the succession in Stirnas is more complete than in Valga.

Association 4 (899.0 to about 898.8 m, upper part of the Kuldiga Formation) is represented by two samples that yielded a relatively high proportion of *Oenonites* sp. B and *Mochtyella* cf. *cristata*, similar to the upper Kuldiga interval in the Valga core (Hints 2001). However, unlike in Valga, *Kettnerites* also is common in this interval of the Stirnas core.

Association 5 (about 889.85–889.4 m, Broceni Member) is represented by three samples, one of which contains several species that are not recorded in underlying strata. Most significant is the occurrence of *Symmetroprion* – a typical element of Silurian faunas of Gotland (Bergman 1995), and *Tetraprion* sp. A, which is common in the late Llandovery of Estonia (Rubel et al. 2007). It is noteworthy that *Symmetroprion* was also recovered in the Broceni Member of the Valga section (Hints 2001).

As a consequence, close links between the faunas of the Valga and Stirnas sections suggest a very similar, though not necessarily contemporaneous, succession of environments. The recovery of several taxa typical of the Silurian of Baltica indicates that polychaetes were affected only moderately by the end-Ordovician extinction.

INTEGRATED RESULTS AND CONCLUSIONS

The succession of sedimentological and palaeontological events in the topmost Ordovician strata of the Stirnas drill core is summarized in Fig. 10. Traditional lithostratigraphical units of the Porkuni Stage in the Livonian Basin were recognized in the core: the Kuldiga and Saldus formations corresponding to shallowing upwards sequences S7 and S8 of Harris et al. (2004, 2005), respectively. Based on palaeontological characteristics, the Hirnantian strata can be subdivided into seven (I–VII in Fig. 10) more or less distinct intervals, which coincide only partly with the lithological complexes described above.

(1) In spite of the gap on the Pirgu–Porkuni boundary in the Stirnas core, the brachiopods, scolecodonts, and conodonts of the basal Porkuni Stage (basal Bernati Member, interval I in Fig. 10) clearly show pre-Hirnantian affinities. To some extent this may be environmentally controlled since the most argillaceous rocks and most deep-water settings are encountered in this interval (Fig. 4). Biostratigraphically, this part of the succession belongs to the lowermost Porkuni *Spinachitina taugourdeaui* Biozone, within the rapidly rising limb of the δ^{13}C curve. These data evidence that the gap at the Pirgu–Porkuni junction is at least to a great extent pre-

Hirnantian (= latest Pirgu), marking a sea level lowstand and a regional regression. The Hirnantian (= Porkuni) begins with a short-lived sea level rise and stabilization.

(2) The bulk of the Bernati Member and base of the Edole Member (interval II in Fig. 10) are characterized by a specific polychaete fauna and the *Hindella–Cliftonia* brachiopod Association, which includes typical representatives of the *Hirnantia* Fauna. At the base of this interval the zonal chitinozoan *Conochitina scabra* and the conodont *Noixodontus* appear.

In the boundary stratotype of the Hirnantian Stage, the Wangjiawan North section, South China, the *Hirnantia* Fauna is contemporaneous with the main peak of the HICE (Chen et al. 2006). In the Stirnas core, by contrast, the *Hindella–Cliftonia* Association obviously predates the maximum $\delta^{13}C$ values. It follows that the *Hirnantia* Fauna in the eastern Baltic is somewhat older than in the stratotype area if the correlation between the $\delta^{13}C$ excursions in Stirnas and Wangjiawan as advocated above is correct. However, the diachronous nature of the *Hirnantia* Fauna is suggested also within South China (Rong et al. 2002; Chen et al. 2006), not to mention also in different continents.

The Bernati–Edole boundary is characterized by an increased content of siliciclastic material and a 1‰ step in the $\delta^{13}C$ curve (Fig. 4). Just above the boundary there are high concentrations of bromium and sulphur, most probably related to increased amounts of organic matter. Palaeontologically this boundary is rather weakly expressed, except in the case of scolecodonts, which display a conspicuous abundance peak in the basal Edole Member (Fig. 9). Similarly high abundance and succession of predominating taxa were recorded in the Valga core at a depth of ca 322 m (Hints 2001), suggesting that placement of the Bernati–Edole boundary in that section may need revision. The coincidence of polychaete bloom and isotope shift in the Stirnas core may be significant, but further data are required to interpret the possible relationships. Typical of the lower part of this interval is also the occurrence of the single chitinozoan species *Rhabdochitina gracilis*. Similar and probably contemporaneous beds where only *R. gracilis* occurs have been identified in NE Poland (Modliński et al. 2002, fig. 3).

(3) The middle part of the Kuldiga Formation (intervals III–IV in Fig. 10) is distinguished by the appearance of low-diversity *Dalmanella testudinaria* brachiopod association, together with the chitinozoan *Lagenochitina prussica* and two distinct polychaete associations. It is interesting to note that two benthic groups, brachiopods and polychaetes, show reversed diversity patterns: a low-diversity brachiopod fauna is associated with a high-diversity scolecodont assemblage, suggesting that the bottom environments were indeed more suitable for vagile benthos than sessile filtrators. With respect to trilobites, interval III is virtually barren, although *Mucronaspis* is rather abundant in under- and overlying strata.

As discussed above, *Lagenochitina* has not been recovered previously from the Porkuni Stage in the Baltic area, suggesting that these strata are poorly known or missing in other sections studied. This interpretation is supported partly by the polychaete fauna. Scolecodont association 3 has no directly comparable counterpart in the Valga section (cf. Hints 2001). Association 4, however, is quite similar to that occurring just above the *Pistoprion*-dominated assemblage (= association 2 in Stirnas; see Fig. 10) in the Valga core. This fits well with isotope data: in both sections similar polychaete faunas occur in the interval of the highest isotope values. Moreover, the general shape of the $\delta^{13}C$ excursion, if compared with those in the southern Estonian sections analysed by Brenchley et al. (2003) and Kaljo et al. (2008), suggests a more complete record in the Stirnas core.

Interval III is also noteworthy with respect to the occurrence of the conodont *Amorphognathus ordovicicus* (Fig. 10). For the first time this species has been recovered definitely from the Edole Member – all previous Hirnantian finds in the Baltic area come from the basal Bernati Member (Kaljo et al. 2008). As in Wangjiawan North (Chen et al. 2006), *A. ordovicicus* occurs below the main $\delta^{13}C$ peak in the Stirnas core.

Interval IV, coinciding with the highest $\delta^{13}C$ peak, marks the decline of shelly fauna. This decline was caused by continuing regression and possibly ocean-water cooling as indicated by 2–3‰ shift in the $\delta^{18}O$ curve.

(4) The topmost part of the Edole Member (interval V in Fig. 10) almost lacks shelly faunas. Chitinozoans and scolecodonts are very rare, but conodonts are fairly common.

(5) The boundary between the Saldus and Kuldiga formations is marked by a notable lithological change (Fig. 4) and a strong discontinuity surface, indicative of a sequence boundary. This level marks a regression and the gap HA *sensu* Bergström et al. (2006), which is established in many shallow-water successions across Europe and North America (Bergström et al. 2006; Schmitz & Bergström 2007). The Piltene Member (interval VI in Fig. 10) contains deposits of high water energy shoal settings (oolitic limestone, cross-bedding structures). Noteworthy is the barium anomaly, which may originate from freshwater influx, further supporting the proximity of a shoreline. The rich conodont fauna recovered in this interval shows clear signs of wear and is considered as being reworked from older strata. No macroscopic shelly fossils or organic-walled microfossils were found in the Piltene Member. However, comparison

with the Ruhnu core suggests that at least the lack of organic-walled microfossils in the topmost Edole and Piltene members is due only to preservational issues and not a sign of a hostile habitat.

(6) The topmost Hirnantian in the Stirnas core is represented by the Broceni Member (interval VII in Fig. 10), which is rich in redeposited conodonts. In addition, abundant chitinozoans and scolecodonts are found in this interval. The chitinozoans are strongly flattened and the same species occur also in underlying strata, so that reworking might be suggested. However, the scolecodont assemblage contains several new forms, indicating that at least some of the organic-walled microfossils in this interval are indigenous. The contact between Ordovician and Silurian strata (= HB in Bergström et al. 2006 and Schmitz & Bergström 2007) coincides with a sharp lithological change and a probable lack of the earliest Silurian deposits.

In summary of the above described patterns, the fossil data provide several criteria for subdivision and correlation of Hirnantian deposits in the eastern Baltic area. In many cases there were also close biogeographical links with other regions: faunas similar to those from the Stirnas core are known also from Scandinavia, British Isles, Belgium, and North America, but also from South China and North Africa. In the course of this study the biostratigraphical utility of the chitinozoans *S. taugourdeaui* and *C. scabra*, and the conodonts *A. ordovicus* and *N. girardeauensis* has been re-confirmed.

The stable isotope record in the Stirnas core, although generally similar to those in other sections, has some distinctive features and poses a number of questions, some of which are already addressed above. The temporal relationships between the main peak of the HICE and the distribution of shelly faunas still requires an additional comment. The maximum diversity of shelly faunas in the Stirnas core falls within the slow stepwise rise of the carbon isotope values. The highest peak of the isotope curve, however, corresponds to the level of the low-diversity brachiopod association, and the continuing isotope values in the topmost Kuldiga and lower Saldus formations contain no shelly faunas. This does not correspond exactly to the global distribution of the *Hirnantia* Fauna, whose diversity rises up to the maximum peak of the isotope curve (Harper & Rong 2008). Also, in the Taagepera core of southern Estonia (Brenchley et al. 2003), most shelly fossils, including the brachiopods *Dalmanella* sp. and *Plectothyrella* sp., occur in the interval with maximum values of the carbon isotope excursion (ca 422.5–420.0 m). On the basis of these factors it may be argued that the Hirnantian benthic faunas appeared first and were more diverse in relatively deeper shelf settings of the Livonian Basin (see also Chen et al. 2006 and the discussion above).

The present study demonstrates that in addition to palaeontological criteria and stable isotope excursions, a chemostratigraphic approach may provide new regional tie-points within topmost Ordovician strata. However, few such data are currently available, so it remains to be tested whether the anomalous concentrations of, e.g., zirconium, chromium, and barium identified in the Stirnas core can be traced elsewhere.

The data from the Stirnas core also shed some light onto problems relating to regional usage of lithostratigraphical units, which, in turn, are often treated as chronostratigraphical proxies. In the type area the Saldus Formation comprises cross-bedded silty limestones with ooids (Brangulis et al. 1989) as well as typical oolitic limestones. In southern Estonian sections only the latter type of rocks is included commonly in the Saldus Formation. For example, in the Ruhnu core the Saldus Formation is identified in the interval between 603.0 and 601.0 m (Põldvere 2003). The same concept is followed by subsequent authors including, e.g., Brenchley et al. (2003) and Bergström et al. (2006). Comparison with the succession in the Stirnas core and other western Latvian sections suggests, however, that the lower boundary of the Saldus Formation in the Ruhnu core should be drawn considerably lower, at the discontinuity surface at 609.5 m above which peloidal limestones with silty and sandy interlayers first appear. Such reinterpretation of the Ruhnu section also alters the interpretation of the stable isotope curve as discussed above. A similar problem occurs with different interpretations of the Bernati Member, whose thickness in southern Estonian sections may be greater if more detailed comparisons are made with the type area in Latvia.

Thus, in summary, Hirnantian strata of the Stirnas core form one of the most complete successions in the eastern Baltic studied as yet in detail, at least with respect to the Kuldiga Formation, and the section would serve as a valuable reference for future studies. In order to provide a more comprehensive understanding of terminal Ordovician geo- and bioevents, further data are required particularly on bioclast carbon and oxygen isotopes as well as from faunal changes along the basin transect in other Baltic sections.

Summarized results of XRF analyses of main components and trace elements from the Stirnas-18 core

Component, %	Median	Minimum	Maximum	Samples with high concentrations, m	Component, ppm	Median	Minimum	Maximum	Samples with high concentrations, m
LOI	30.3	22.2	42.1		As	<10	<10	45	889.4; 895.65
SiO_2	18.64	1.6	36.74		Ba	118	<20	533	892.3; 892.6
TiO_2	0.25	0.02	0.49		Br	5	<5	19	907.9; 908.23
Al_2O_3	4.56	0.38	11.09		Ce	<20	<20	56	
Fe_2O_3	2.01	0.32	4.06		Co	<20	<20	315	911.1
MnO	0.25	0.18	0.66	908.7–911.65; 912.9	Cr	68	<10	161	901.2–903.6; 904.3
MgO	2.51	0.39	8.96		Cu	8	<10	237	912.6
CaO	34.26	15.33	53.18		Ga	<10	<10	10	
Na_2O	0.28	0.05	0.58		La	<20	<20	26	
K_2O	1.66	0.14	3.38		Ni	37	5	280	911.1
P_2O_5	0.06	0.02	0.09		Pb	<10	<10	54	889.4; 895.65
Cl	0.14	0.05	0.29		Rb	46	<5	98	
S	0.09	0.01	0.69	889.4; 907.9; 908.2	Sr	149	110	205	
					V	28	<10	64	
					Y	13	4	17	
					Zn	25	4	51	
					Zr	94	7	262	889.4; 890.0

Whole-rock ($\delta^{13}C_{bulk}$) and bioclast ($\delta^{13}C_{brach}$, $\delta^{18}O_{brach}$) stable isotope data from the Stirnas-18 drill core

Depth, m	$\delta^{13}C_{bulk}$	Depth, m	$\delta^{13}C_{bulk}$	Depth, m	$\delta^{13}C_{bulk}$
889.40	2.23	896.05	4.62	904.90	3.00
889.43	2.58	897.00	4.06	905.40	2.65
889.89	2.68	898.40	2.49; 2.31; 3.56	906.40	2.70
890.00	2.70	898.80	4.88	907.20	2.88
890.20	3.16	899.00	4.27	908.10	3.43
890.60	3.04	899.90	2.85	908.30	2.39
891.85	4.59	900.75	3.83	908.70	2.17
892.70	4.74	901.25	2.44	910.00	2.74
893.70	4.06	901.60	2.14	910.30	2.55
894.25	4.10	902.30	2.37	911.15	2.52
894.65	4.03	902.95	2.54	911.50	2.92
895.00	3.85	903.10	3.17	911.75	1.90
895.25	3.98	903.40	2.21	912.10	1.35
895.35	4.44	903.80	2.59	912.45	1.08
895.65	4.33	904.30	3.76	912.90	1.04

Depth, m	$\delta^{13}C_{brach}$	$\delta^{18}O_{brach}$
899.00	6.59; 6.98; 7.40	−1.50; −1.13; 0.26
899.60	6.43; 6.53; 7.05; 7.32	−1.26; −1.15; −0.52; −0.48
901.80	5.87; 5.89; 5.73; 5.10; 5.15	−3.52; −3.57; −3.91; −4.52; −3.30
903.60	5.43; 5.02; 5.11; 5.15	−3.86; −3.00; −4.32; −3.25
909.00	4.33; 5.12; 5.00	−4.76; −4.87; −3.96

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Hirnantia lademe (Ülem-Ordoviitsium) bio- ja kemostratigraafia Stirnas-18 puursüdamikus Lääne-Lätis

Linda Hints, Olle Hints, Dimitri Kaljo, Tarmo Kiipli, Peep Männik,
Jaak Nõlvak ja Helje Pärnaste

On esitatud andmed Porkuni lademe litoloogilis-geokeemilise, sh isotoopkoosseisu, makro- (brahhiopoodid, trilobiidid) ja mikrofossiilide (konodondid, kitiinikud, skolekodondid) leviku ning taksonoomilise koosseisu kohta Stirnas-18 puursüdamikus. Puurauk asub Lääne-Lätis Porkuni lademe suurimate paksuste vööndis, mis on ladet esindavate Kuldīga ja Salduse kihistu tüüpala. Süsiniku ja hapniku isotoopkoosseisu muutused on võrreldavad varasemate andmetega nii Läti kui Lõuna-Eesti läbilõigetest. Uuritud läbilõikele on iseloomulik isotoopkõvera järkjärguline tõus ja suurimate väärtuste esinemine Edole kihistu ülemises osas. Erinevate fossiilikoosluste levik võimaldab Porkuni lademes eritleda 7 faunaintervalli. Konodont *Noixodontus*'e faunat sisaldavas Kuldīga kihistus on brahhiopoodid esindatud kahe kooslusega koos Porkuni lademele iseloomuliku trilobiidiga *Mucronaspis mucronata*. Samasse intervalli jääb kitiiniku *Lagenochitina prussica* levik, mis oli seni teada ainult Porkuni-eelsetest kihtidest. Edole kihistu keskosast leiti seni noorimad tsonaalse konodondi *Amorphognathus ordovicicus* esindajad. Polüheetide (skolekodontide) koosseisu ja leviku põhjal on piiritletud viis kooslust, millest mitmed on jälgitavad ka Lõuna-Eesti läbilõigetes. Porkuni lademe ülemises osas (Salduse kihistus) makrofossiile ei esine ja konodondifauna viitab ümbersettimisele. Käesolev töö ja võrdlus stratotüüpsete läbilõigetega osutab vajadusele ümber hinnata mitmete Lõuna-Eesti läbilõigete senist litostratigraafilist liigestust.

   Stirnas-18 läbilõike uurimistulemused toetavad meie varasemat seisukohta, et Porkuni lademe alumisse ossa jääv kitiiniku *Spinachitina taugourdeaui* biotsoon Baltikumis vastab Hirnantia lademe alumisele, mitte aga ülemisele osale, nagu seda on arvanud mõned stratigraafid. Viimaste arvates kuulub Hirnantia lademe alumisse ossa *Belonechitina gamachiana* biotsoon. Baltikumis esineb nimetatud kitiinik aga Pirgu lademesse kuuluvas Taučionyse kihistus, kus esineb brahhiopood *Holorhynchus*, mis kõikjal Ordoviitsiumi läbilõigetes on Hirnantia-eelse vanusega.