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Chitinozoan Biostratigraphy in the Ordovician of Baltoscandia

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ABSTRACT

In the present study the origin, zonation and distribution of chitinozoans and their usefulness in dating different geological events are discussed.

Rapid development of chitinozoans was probably due to their (at least part of them) relatively high evolutionary potential and pelagic mode of spreading. The analysis of the origin and possible mode of life, as well as the species level distribution and variations in the diversity of chitinozoans shows that they represent an ontogenetic stage (eggs) in the life cycle of a cryptic fossil group. The eggs register the variations of environment affecting the group.

The results of the study reveal relatively low diversity of chitinozoans in their early stage of development in Baltoscandian Ordovician succession. Intensive diversification of the group was in the Darriwilian, continued in the Caradoc and Ashgill up to the well-documented end-Ordovician crisis. Chitinozoans appear to be a group suitable for detailed investigations because of their fairly restricted vertical range, worldwide distribution and rather limited number of carefully described species.

On the basis of the material from different areas of the Baltoscandian Ordovician basin the chitinozoan zonation was proposed and compared with the graptolite and conodont zonations. It was established that stratigraphic evolution of chitinozoans surpassed that of other faunal elements in upper Ordovician. Relatively well documented regional zonations serve as a basis for new long-distance correlations between several palaeocontinents, which are complicated due to the well-known differentiation of Ordovician faunas. At the same time, a more detailed biozonation is needed for the correlations of rocks from varied confacies belts within the Ordovician of the Baltic Basin.

LIST OF ORIGINAL PUBLICATIONS

This thesis is based on the following publications:

I Paris, F. & **Nõlvak, J**. 1999. Biological interpretation and paleobiodiversity of a cryptic fossil group: the "chitinozoan animal".*Geobios*, **32,** 2, 315-324.

II **Nõlvak, J.** & Grahn, Y. 1993. Ordovician chitinozoan zones from Baltoscandia. *Review of Palaeobotany and Palynology,* **79**, 245-269.

III **Nõlvak, J.** 1997. Ordovician. Introduction. *In:* Raukas A., Teedumäe A. (eds.). *Geology and Mineral Resources of Estonia.* Estonian Academy Publishers, Tallinn, 52, 54-55.

IV **Nõlvak, J.** 1999. Ordovician chitinozoan biozonation of Baltoscandia. *Acta Universitatis Carolinae. Geologica,* **43**, 1/2, 287-290.

V **Nõlvak, J.** 1980. Chitinozoans in biostratigraphy of the northern East Baltic Ashgillian. A preliminary report. *Acta Palaeontologica Polonica*, **25**, 253-260.

VI **Nõlvak, J.,** Meidla, T. & Hints, L. 1989. The Taučionys Formation (*Holorhynchus* Beds) in the Iljinskoje boring (Pskov District). *Proceedings of the Academy of Sciences of the Estonian SSR. Geology,* **38,** 88-93.

VII **Nõlvak, J.,** Grahn, Y. & Sturkell E.F.F. 1999. Chitinozoan biostratigraphy of the middle Ordovician Dalby Limestone in the Fjäcka section, Siljan District, Sweden. *Proceedings of the Estonian Academy of Sciences. Geology*, **48,** 75-85.

VIII Kaljo, D., **Nõlvak, J**. & Uutela, A. 1996. More about Ordovician microfossil diversity patterns in the Rapla section, northern Estonia. *Proceedings of the Estonian Academy of Sciences*, **45**, 131-148.

IX Grahn, Y., **Nõlvak, J.,** & Paris, F. 1996. Precise chitinozoan dating of Ordovician impact events in Baltoscandia. *Journal of Micropaleontology*, **15,** 21-35.

X Paris, F., Verniers, J., Achab, A., Albani, R., Ancilletta, A., Asselin, E., Chen, X., Fatka, O., Grahn, Y., Molyneux, S., **Nõlvak, J.,** Samuelsson, J., Sennikov, N., Soufiane, A., Wang, X. & Wincester-Seeto, T. 1999. Correlation of Ordovician regional chitinozoan biozonations. *Acta Universitatis Carolinae. Geologica*, **43,** 1/2, 291-294.

XI Kaljo, D., Hints, L., Hints, O., Martma, T. & **Nõlvak, J**. 1999. Carbon isotope excursions and coeval environmental and biotic changes in the late Caradoc and Ashgill of Estonia. *ibid.* 507-510.

XII Brenchley, P.J., Marshall, J.D., Hints, L., & **Nõlvak, J**. 1997. New isotopic data solving an old biostratigraphic problem: the age of the upper Ordovician brachiopod *Holorhynchus giganteus*. *Journal of the Geological Society, London*, **154,** 335-342.

XIII Kaljo, D., Hints, L., Martma, T. & **Nõlvak, J**. 2001. Carbon isotope stratigraphy in the latest Ordovician of Estonia. *Chemical Geology,* **175**, 49-59.

INTRODUCTION

Palynology has been applied to a wide assemblage of geological problems, including palaeobiogeography, palaeoclimatology and the identification of depositional environments. Biostratigraphy, which is "the organization of rock strata systematically into named units based on content and distribution of fossils" (Hedberg 1976, p. 45) and correlation of the Ordovician rocks remain the primary applications. The importance of the chitinozoan biostratigraphy is highlighted by the fact that they constitute a group of organic-walled microfossils, well represented and commonly recovered from marine sediments. Based on worldwide results, it has become evident that many chitinozoan species, when carefully identified, have relatively short ranges, and are useful for stratigraphic correlations.

Although biostratigraphic and chronostratigraphic units are considered as independent elements in stratigraphic classification, there is a growing dependence on the correlation of synchronous horizons to help resolve some problems such as duration of unconformities or gaps, synchroneity of eustatic cycles, relative rate of sediment accumulation, etc. For this reason, the integration of biostratigraphic units into linear time-scales is receiving increased attention. A linear time-scale is one graduated in equal units of absolute time (Christopher & Goodman 1996). The main tools for constructing time scales are relative scales calibrated with unique biostratigraphic datums and zones in fossiliferous deposits. And almost all Ordovician deposits are relatively rich in microfossils. Actuality of that subject is pointed out by the International Geological Correlation Program (IGCP), which approved Project 410 "The Great Ordovician Biodiversification Event", and in which the author is actively involved.

Some results and problems of the biostratigraphy of chitinozoans, their zonal distribution in the Ordovician rocks of the East Baltic and Scandinavia, and usefulness of the chitinozoan zonation in relative timing of several geological events (volcanic ashfalls, impact events, gaps in sedimentation, but also building material of some historical objects, etc.) are the objectives of that summary.

1. CRYPTIC FOSSIL GROUP

One of the recently actively studied microfossil groups in the Lower Palaeozoic sedimentary rocks in the whole world is chitinozoans. They constitute an enigmatic group of fossils whose significance and biologic affinities are still unconfirmed, despite that the first descriptions were given by Eisenack (1931) more than 70 years ago from the Baltic material. His material was collected from Quaternary glacial deposits – erratic boulders scoured from the Ordovician and Silurian bedrock of the northeastern Baltic. In pure limestone boulders fossils are usually very well preserved, which was very important at the early stage of chitinozoan studies. A modified shape-based classification by Eisenack is still used (Paris *et al*. 1999).

Chitinozoans are organic-walled microfossils, occurring in the early Ordovician (Upper Tremadoc) to latest Devonian (Upper Famennian) unexceptionally marine deposits from nearly all Palaeozoic seas. So, they proliferated worldwide during 130 Ma. In Baltoscandia they are preserved only in the lower Palaeozoic part – in Ordovician and Silurian sedimentary rocks. They are found isolated or in rare cases in chain-like structures in most sedimentary (see exceptions below) or low-grade metamorphic rocks. Abundance usually ranges from a small number of specimens to several tens of specimens per gram of rock (exceptionally up to several hundred to thousand), and the highest production of chitinozoans seems to correspond to high-latitude cold-water environments (Paris 1996).

The primary morphological features utilized in chitinozoan taxonomic assignment are overall shape, presence or absence of spines, and other ornament. A chitinozoan individual consists of two main parts — chamber and neck, has an opening closed by operculum or prosome, their black hollow vesicles normally display a radial symmetry and length ranges from 50 up to 2000 μm, usual size is one to several hundred microns. The largest specimen in the world, a 2.562 mm long *Conochitina minnesotensis* has been established by the author in the Baldone core (Latvia), at a depth of 856.3 m (Keila Stage, Upper Ordovician).

The occurrence in chain-like structures does not prove colonial behaviour of chitinozoans (see Paris & Nõlvak 1999 —PAPER I) as each basic vesicle has no communication with neighbouring ones. The actual chemical composition of the wall is still unclear, however, they were named after the chitinoid appearance by A. Eisenack. Biologic affinity and significance of chitinozoans are interpreted in a variety of ways (see review in Miller 1996), but based on the results of Paris (1981) and Paris & Nõlvak (1999 –PAPER I), here the hypothesis of eggs of soft-bodied metazoans is favoured.

Below, some other aspects (following beside the above papers also Paris 1990) allowing us to suggest that chitinozoans were eggs of marine organisms are discussed. In all probability, they were metazoans with a soft body, which were not preserved in ordinary conditions of fossilization. There is no strict concordance between the range of any known fossil group and chitinozoans. Also, many comparisons have been made between only a limited number of chitinozoan species and one or two present-day living organisms, not embracing all variety of forms in the whole group. Therefore, partial solutions have prevailed and chitinozoans have been regarded more like a collection of isolated things than like a biological entity. It is obvious that a drastic change in the behaviour of living organisms (e.g. accession to a terrestrial, lacustrine or marine mode of life) may be followed not only by fundamental modifications of the organism itself but also in a mode of its reproduction. It seems reasonable to try first to identify the biological significance of the chitinozoans and then, to test the available "candidates" among known fossil groups, rather than to search individual similarities with some modern organisms.

One basic feature of all well-preserved chitinozoans is their hermetical closure. It means that an individual is not able to ensure important vital functions as feeding, excretion or reproduction. There are two exceptions from this: survival stages like cysts, which occur among plants (algae), protozoans or primitive metazoans, and reproductive stages such as spores, eggs, egg-capsules, etc. In both cases the inner content is temporarily isolated (cut off) from the environment. Another specific feature distributed among cysts and reproductive stages is the absence of juvenile forms within their populations. This is well documented also among chitinozoans. It can be illustrated by morphometric diagrams (see Paris 1981; Grahn *et al.* 1996 –PAPER IX), where the distribution of dots is always like "cloud" and not arranged along a regression line as for normal populations including growing stages. It seems evident that chitinozoan vesicles arrived into seawater with their definite size (which was not obviously equal: within one population the vesicles of a species can be almost equal or sometimes their size can differ up to three times), like cysts and eggs. The same is true also of their geographic distribution: water dynamics (currents, waves) and temperature control were (and are) the main factors favouring or restricting the dispersion of cysts and pelagic eggs, and most probably also chitinozoans.

It is obvious that soft-bodied metazoans can be fossilized only in some exceptional or peculiar conditions (e.g. Gabbott *et al.* 1998), such as was the case of conodonts (Briggs *et al*. 1983). In future biochemistry of the chitinozoan test can also give some more information about their extreme resistance to many acids, when secondary thermal heat flows have a stronger influence to their preservation.

Finally, in analysing the hypotheses of the systematic position of the chitinozoans, attention has been paid also to their abrupt disappearance from the fossil record in the topmost Famennian (Paris 1981). One can argue that the ultimate disappearance of the chitinozoans was preceded by a progressive decline in their diversity of species (see Paris & Nõlvak 1999, figs. 3, 5 – PAPER I) rather than by significant drop in the abundance of the vesicles. This could be a true extinction, or could also correspond to a fundamental change in the behaviour of the chitinozoan producers (chitinozoophorans by Grahn 1981). It is worthwhile pointing out that the early Carboniferous is usually regarded as a life spreading time and it is possible that the chitinozoan producers had left the marine environment and adopted a terrestrial mode of life. Such a change could be accompanied with modifications of their reproduction strategy or type. In relation to this it is worth noting that eggs of some insects have the composition, shape and structure similar to some chitinozoans. This refers to the consequences of drastic behavioural changes in the fossil record and does not mean that insects are directly derived from chitinozoan producing organisms. On the other hand, the diversity of such eggs through time can be used to test the fluctuations of the biodiversity of cryptic soft-bodied organisms producing resistant eggs on a world-wide scale, or given palaeoplate (Paris & Nõlvak 1999, p. 320 –PAPER I).

USE AS A STRATIGRAPHIC TOOL

Chitinozoans have proved to be useful for palaeogeographic reconstructions and a very efficient stratigraphic tool (see Paris *et al.* 1999). However, there occur some problems as all groups of fossils are to some extent dependent on facies. Minute and relatively light fossils as chitinozoans were, have been subject to much the same processes of sedimentation as different clasts. It is the combined interaction of these processes upon biological and detrital sedimentary material that produces facies (Jenkins 1970). Obviously chitinozoans are distributed not independently of facies, but that compared to other fossil groups, they are relatively free from facies control, and being considered (see Verniers *et al.* 1995) as a planktic group, are distributed relatively independently of minor facies changes (e.g. Nõlvak 1980, fig. 2 –PAPER V). At the same time, some interesting efforts to define clear chitinozoan biofacies (Grahn 1982b, Nestor 1998) need much more precise data for correlation of described beds (e.g. proved by some other fossil group) before so radical conclusions which were presented. Main problems concern the age of the compared layers.

Taxonomic study of the Ordovician chitinozoans of Baltoscandia started by A. Eisenack was continued in Sweden by S. Laufeld from the Fjäcka and Amtjärn sections (Laufeld 1967), and from 1980s mainly by Y. Grahn (see Grahn 1982a and references therein).

During the late 1960s, R. Männil showed interest in the usefulness of chitinozoans in the biostratigraphy of the East Baltic sections (Männil 1971). Later his chitinozoan studies were concentrated to the more restricted parts of middle Ordovician: from Lasnamägi to Haljala stages (upper Llanvirn — lower Caradoc; see Männil 1986, and references therein). So, in Estonia for a relatively long time much effort was put in some definite Ordovician correlation problems, while less attention was paid to the distribution of chitinozoans in the whole Ordovician succession.

The first chitinozoan zonation schemes for the upper Ordovician were published by the author (Nõlvak 1986, 1988) from Estonian and Lithuanian sections and later together with Y. Grahn already for the whole Ordovician of Baltoscandia (see Nõlvak & Grahn 1993; emended in Nõlvak 1999 –PAPERS II, IV and references therein) and for Estonia (Nõlvak 1997 –PAPER III). These schemes have also comparisons with the zonations of graptolites and conodonts, which show clear dissimilarities in the usefulness between these fossil groups as a stratigraphic tool in a particular part of the Ordovician.

MATERIAL AND METHODS

Most of the material studied from more than a hundred sections has been collected by the author since 1966. Additional material (some hundreds of samples) collected in the Institute of Geology (until late 1980s) was also used. The material analysed here for the study of the chitinozoan diversity changes (in press) comprises of 5362 samples and was collected from the sections of Estonia – 2000, Latvia – 500, Lithuania – 325, NW Russia – 370, Poland – 100, Belarus – 90, Ukraine – 25, Norway – 270 (Grahn *et al.* 1994) and Sweden - 1682. In Sweden many of the samples were collected together with Y. Grahn in 1992, 1995 and prepared by the author (Grahn *et al.* 1996 –PAPER IX; Nõlvak *et al*. 1999 –PAPER VII), and some published data were also used (Grahn 1980, 1981, 1982a).

The samples with a usual weight of 0.3–0.5 kg were dissolved using weak hydrochloric or acetic acid. For technical reasons, hydrofluoric acid was not used. However, only few types of rocks, mainly restricted black shales in the East Baltic Ordovician successions need to be dissolved in this strong acid, which is important advantage in the microfossil research. A finely drawn glass pipette was used to gather chitinozoan specimens from insoluble residue, and also specimens of some other microfossil groups as bigger forms of acritarchs, graptolites, melanosclerites and minute inarticulate brachiopods etc. were also collected. All specimens are kept in small plastic boxes in glycerine, which enables re-examination of specimens and justifies the time used. Usually for conodonts such a pipette technique is unsatisfactory, because most of them are buried under the insoluble residue. Hence, together with separating chitinozoans from detailed sometimes bed-by-bed series of samples it was possible to collect and describe also other rare groups of microfossils: radiolarians (Nazarov & Nõlvak 1983), endolithic trace-fossils (Podhalańska & Nõlvak 1995), inarticulate brachiopods (Popov *et al.* 1982; Popov & Nõlvak 1987; Popov *et al.* 1994; Wright & Nõlvak 1997; Lécuyer *et al.* 1998), or muellerisphaerids (see Estonian data by the author in Holmer 1987). The author (unpubl.) has recovered the oldest muellerisphaerids already from the Kunda Stage in the Baldone section of Latvia. So, the total range of this group can now be extended down to the lower Darriwilian. Earlier muellerisphaerids were known from the Caradoc to the Upper Devonian (Hüneke & Reich 2000).

Most of the SEM analyses were made in Tallinn (1978–1995), but also in Rennes (1995) and Stockholm (1995). The main ideas and criteria in a new classification (Paris *et al.* 1999) are accepted and used, but the main attention in this paper is paid to the chitinozoan distribution.

UNFAVOURABLE CONDITIONS

Some parts of the Ordovician succession in Baltoscandia are unproductive for chitinozoans due to environmental or/and preservational reasons. First of all such unfavorable facies were formed in active water environments. Environmental reasons.

(a) Glauconite-bearing sandstones of the Hunneberg and Billingen stages in North Estonia. In many borehole cores these beds are also strongly influenced by dolomitization (e.g. in South Estonia). In these sandstones chitinozoans have a sporadic distribution, indicating to the probability of reworking, at least partly.

(b) Reefs (also bioherms or carbonate buildups in early literature): in the Keila–Oandu boundary beds in northwestern Estonia and in the Pirgu Stage in South Estonia and Sweden (Kullsberg and Boda Limestones). Barren are also samples from the oldest carbonate buildup in Baltoscandia (from early Idavere time) – on the rim of the Lockne impact crater (Grahn *et al.*1996, –PAPER IX).

(c) Real lagoonal facies (very restricted in the Baltoscandian Ordovician): rare oolitic grainstones are known from the Kardla core section in southern Estonia and from the Ukmerge core section in Lithuania from the topmost Porkuni Stage (Nõlvak 1988, fig. 2).

Preservational reasons.

(d) Secondary dolomitization reduced organic-walled microfossils to "powder". This process could occur in all stratigraphical levels in North Estonian and partly Lithuanian Confacies belts. Almost always the topmost Ordovician beds are also dolomitized and chitinozoans are not found, except where all beds of the Porkuni Stage are absent, and have been eroded most probably before the Silurian. Such sections are typical of the transitional area between two main belts, e.g. in the Eikla, Kingissepa, Pärnu and Are core sections in Estonia and Jaksai in Lithuania (Nõlvak 1986, figs. 3, 4; 1988, fig. 4; 1989, figs. 1, 6).

(e) Marine redbeds are totally barren in South and West Estonia, Latvia, Poland and Sweden in the lowermost Middle Ordovician (Volkhov and lower Kunda stages, or partly even higher, in the Aseri–Lasnamägi stages) and in the same regions in the lower Ashgill (lower Pirgu Stage). Rocks with red spots, often distributed in some levels of the Nabala, Vormsi and Pirgu stages, have a different degree of productiveness. This indicates to a possibility of secondary destruction of all kind of organic-walled microfossils in redbeds, when widely distributed pyrite crystals in the rock recrystallized to goethite–hematite and microscopic spatial changes caused dense splitting of the walls of fossils, reducing them to powder.

(f) In Norway and partly in Sweden (e.g. in Scania, author's observations) rocks are influenced by thermal heat flow (Permian volcanics), and the critical palaeotemperature for the preservation of chitinozoans is about 300°C. This temperature range is calibrated according to the CAI values of conodonts or colour changes of acritarchs (Hagenfeldt 1997) from the same beds. Often in Norway (Grahn *et al.* 1994) and in Sweden in the Hallen section near the Lockne impact the chitinozoan vesicles, if preserved, have become very brittle, having a metallic lustre. Such changes were also followed among acritarchs and scolecodonts (Sturkell *et al.* 2000).

All the named aspects influenced our results and were taken into account during the sampling (e.g. there is no need to take samples from redbeds or pure dolomites). At the same time, it is difficult to calculate the fatal degree of dolomitization (or to find real percentages) or density of red spots in the rocks when chitinozoans are not preserved.

In general, the preservation of chitinozoans in the Ordovician limestones is good to excellent, but poor to bad in graptolite shales (sections in Scania and central Poland).

2. CHITINOZOAN BIOZONATION

The practical usefulness and the main task of any fossil group in stratigraphy are their ability (potential) to serve as a basis for the subdivision of the layers in a certain section and then to compare (correlation) with some other sections. For a long time the biostratigraphical potentiality of the chitinozoans was not in use because of the fact that the main material was obtained from erratics with more or less uncertain age (Eisenack 1965). In the Ordovician three levels of comparisons (correlations) can be separated: global – between different palaeoplates, regional – between different areas within one basin, e.g. in Baltoscandia: between five main composite environmental belts (see Männil 1966; Jaanusson 1976, for a general review), and local – within one confacies belt, e.g. in detailed correlations of the oil shale-bearing beds in North Estonia (see Männil 1986).

The first chitinozoans in Baltoscandia appeared in the upper Tremadoc, in the topmost clay beds of the stratotypical section of the Varangu Stage (North Estonia, Nõlvak 1999 –PAPER IV). Higher, in the lower Arenig up to 4 m thick (in the Pakri peninsula, northwestern Estonia) glauconite-bearing sandstones and dolomites of the Hunneberg and Billingen stages with rare chitinozoans are distributed. Higher, in the lowermost Volkhov Stage, dolomites or redbeds (in East Baltic) or overheated beds (in Scania, Sweden) occur. So, there is an about 16 Ma period for which the information about chitinozoans is very scarce, and any kind of more detailed subdivision is practically impossible (Nõlvak & Grahn 1993, Nõlvak 1997, 1999 –PAPERS II, III, IV). It means that detailed zonation is in a good use in predominantly bedded limestones (commonly 200–250 m) of the Middle and Upper Ordovician in the whole Baltoscandia and some adjacent areas (northwestern Russia, Poland, Belarus, Ukraine).

LOCAL LEVEL

The pioneer work of Laufeld (1967) on the upper Ordovician of the Fjäcka section of Sweden, and the first series of bed-by-bed samples from the Lipu core section of North Estonia (see Nõlvak 1972) arosed much interest in use the first data of chitinozoans in the local stratigraphy.

In his papers about the distribution of selected chitinozoans Männil (1971, 1972) noticed that "most of mentioned species seem to represent good index fossils and their vertical ranges may be considered as zones" (Männil 1972, p. 142), and further "...in other cases distinct repetitive zones ... referred to as zonules, may be useful in correlations"(*ibid.*). The latter type is in principle the same as ranges of the Lazarus taxa and the distribution of species from the chitinozoan genus *Cyathochitina* is worthy of detailed studies. This system is built up on the changes of the different species of this genus (*calix, campanulaeformis* and *kuckersiana,* Fig. 1). Most thoroughly were studied Upper Ordovician sections in the North Estonian Confacies Belt. In his work of biomicrostratigraphy Männil (1986) suggested a new type of designation of the beds (zonules – ecologically conditioned repetitive zones) with certain main species of *Cyathochitina*, with female names and mail names when the beds are without that genus (Fig. 1). One advantage in such a system is a possibility of avoiding very complicated markers of individual zonules with numbers, letters, etc., because during detailed correlations often new layers appear between those known earlier, which also need their own specific indexes. This practical system was also used by Nõlvak (1972, figs. 2, 4). It was demonstrated in this study that in comparison with the stratotypical area of the Kukruse Stage the southward sections are more complete on account of the uppermost beds of the Kukruse Stage (see also Männil & Bauert 1986; Hints *et al.* 1995, fig. 4). Well grounded correlations of kukersite-bearing layers of the topmost Kukruse Stage were achieved also in the early 1980s (author's observations, unpubl.) during the comparison of the Kuusiku, Rooküla, Palvere and Sibila sections (near Kehra). This gave similar correlation as in the meridional cross-section from the central part of the kukersite basin (see Männil 1986, figs. 1.1; 2.1.1).

In 1967, during detailed bed-by-bed studies of the acid-resistant microfossils from beds of the topmost Kukruse Stage, the author discovered a layer of bentonite in the Kamariku core section (at a depth of 153.21–153.25 m) laying 1.4 m below the palaeontologically proved lower boundary of the Idavere Substage of the Haljala Stage. This bentonite layer is a good stratigraphical marker as it has a wider distribution than any kukersite seam or specific limestone layer, and according to the data available, is the oldest in the East Baltic area. In Sweden, in the Fjäcka section (Nõlvak *et al.* 1999 –PAPER VII) *Conochitina tigrina* occurs among chitinozoans below and above this bentonite

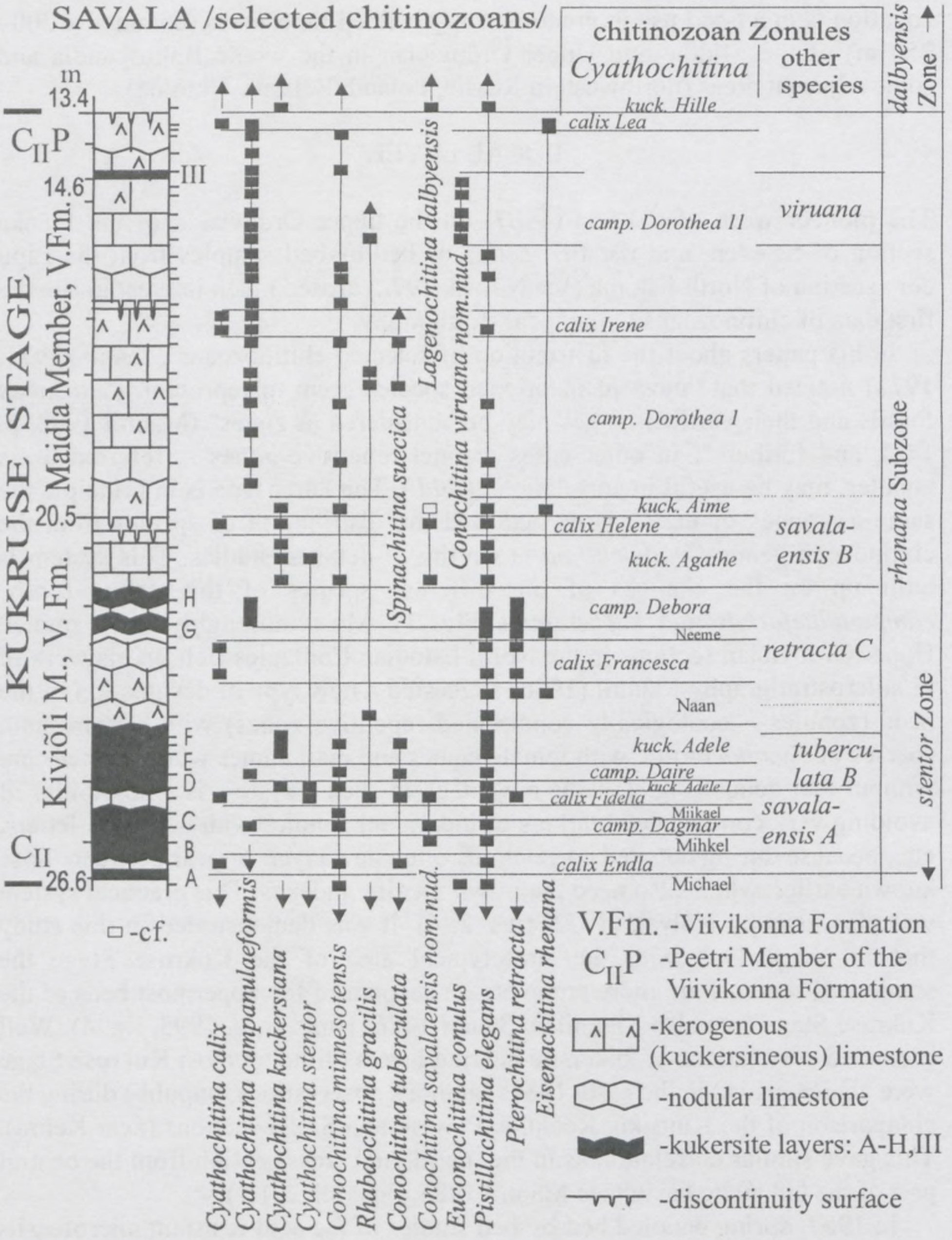

Fig. 1. Different types of chitinozoan zonations. Zonules are useful for correlation in local level. Faunal log of the selected species of chitinozoans (*Desmochitina, Rhabdochitina* and some *Conochitina* are not shown) of the Kukruse Stage in the Savala core section (hypostratotype). The chitinozoan zonule names partly according to the system of R. Männil (unpubl.) and Nõlvak 1972, fig. 2. These beds belong to the *rhenana* Subzone of the *stentor* Zone useful in regional (Nõlvak & Grahn 1993; Modlinski et al. 2002) and global(Paris *et al.*1999) levels. *Conochitina savalaensis, C. viruana nom.nud.* and *Eisenackitina rhenana* are restricted only to the Kukruse Stage in the entire Baltoscandia.

layer (Laufeld 1967). This species has been found only in the topmost part of the Kukruse Stage from many Estonian sections, having a range about 1–2 m, or even less (e.g. in the Valga section, Nõlvak 2001b, app. 8). In northern Estonia this species occurs just below or above that bentonite (e.g. in the Kamariku, Rapla, Palvere, etc. cores). Therefore the same age for these bentonites in Fjäcka and Estonia is suggested (Nõlvak 2002).

During the detailed study of the areal and temporal peculiarities of kukersite accumulation within the upper Uhaku and Kukruse stages (up to 30 m of limestone and kukersite layers) more than 60 described and named zonules were established (Männil 1986). One example presented here is from the Savala hypostratotype section of the Kukruse Stage (Fig.1), showing the possibilities of detailed subdivision.

Another Ordovician sequence in the North Estonian Confacies Belt, where such a biomicrostratigraphical investigation could give in future a good result is the so-called Building Limestone from the Lasnamägi and lower Uhaku stages. These beds, cropping out exceptionally well in the Tallinn area, deserve a complex study of many microfossil groups to find a specific association of fossils for each individual bed. Besides practical value (in archaeology, history of building, etc.) these beds arouse interesting lithological problems concerning problematic widely distributed sedimentological (marked by phosphatic discontinuity surfaces), for which a detailed stratigraphical ground is needed before making far-reaching conclusions. Fossils have also relatively good state of preservation in these beds.

Hence, the described system of using zonules gave a good result within one confacies belt, first of all, in the studies of kukersite-bearing beds, because the specific lithological features of limestones can be used as additional data or "control" in detailed bed-by-bed correlations.

REGIONAL LEVEL

Reliable biostratigraphic correlation is dependent on the ability to define, differentiate and compare biostratigraphic units, and the main task is to find synchronous horizons. A characteristic feature in the regional Ordovician geology of Baltoscandia is the arrangements of several belts having a quite complicated configuration (see Jaanusson 1976, 1995; Nõlvak 1997, fig. 24 – PAPER III). Each of them has specific litho- and biofacies, referred to as confacies belts, show similarities in faunal development and recurrent lithofacies, having a relatively stable position during most of the Ordovician time. However, there were also exceptions. According to facies changes, the transitional area between the main belts was located geographically in different areas (Põlma 1982; Nõlvak 1984, figs. 2–8). The Ordovician time was characterized by an extreme biogeographical differentiation of both benthic and planktic faunas, but with varied degree, at least in the well-preserved and studied part of the East Baltic. This differentiation makes the worldwide correlation of rocks difficult and has resulted in numerous regional stratigraphical schemes.

However, a series of detailed stratigraphical charts compiled for the East Baltic succession (see Männil 1989, and references therein) gave a relatively stable detailed regional classification for all Ordovician rocks, and less detailed data about the distribution of chitinozoans have always been used for compilation of all earlier stratigraphical charts, in which the author was also actively involved.

An emended version of the correlation chart (Nõlvak 1997, Table 7 –PAPER III), compiled for the Estonian succession ranging from 70–180 m in thickness, is based mainly on the above-cited charts and on the chitinozoan zonal schemes elaborated for the whole Baltoscandia (Nõlvak & Grahn 1993, –PAPER II; supplemented Nõlvak 1999 –PAPER IV).

Compared with the other groups, chitinozoans have given most precise stratification and correlation of sequences on many levels in the Viru (Nõlvak 1999), especially in the Harju series (Nõlvak 1979, 1980 –PAPER V, 1984, 1986, 1987a, 1988). In mostly borehole core material, many other groups of fossils are rare (macrofossils, conodonts), or are specifically poor in certain rocks (graptolites). The others are more dependent on particular facies, such as ostracodes (see Meidla 1996, figs. 41–47) or scolecodonts, having also relatively long ranges (see Hints 2000), or are yet insufficiently studied, such as acritarchs (see Uutela & Tynni 1991).

Sometimes very high differentiation of beds can be achieved using chitinozoans. Recently, detailed and repeated sampling of the Fjäcka section in Sweden (in Central Baltoscandian Confacies Belt) has revealed the presence of some very short-range chitinozoan zones, e.g. the *Armoricochitina granulifera* Zone with the thickness not exceeding 15 cm (Nõlvak *et al.* 1999 –PAPER VII). In Lithuania this zone ranges for about 30 cm (Bliudziai core, 1373.6–9 m), and in North Estonia for 20 cm (Imavere core, 215.7–9 m) occurring in three different confacies belts. As an exception, the smallest proved thickness of 1.5 cm of the *Lagenochitina dalbyesis* Zone (found also in the lowermost Idavere Substage in the whole Baltoscandia) is registered in the North Estonian Core No. 6355, between two discontinuity surfaces.

The proposed 26 chitinozoan zones (Nõlvak 1999 –PAPER IV) are generally range zones, preferably total range zones. The index species must be common throughout Baltoscandia (at least in two confacies belts, Nõlvak 1986, 1988) and have a restricted stratigraphic range. Sometimes, however, the index species is very rare, e.g. *Angochitina multiplex*, found in North Estonia, South Estonia and Sweden just above the Kinnekulle K-bentonite and in an erratic boulder in Germany (Schallreuter 1963), but not in the lower boundary-stratotype section near Tallinn (Hints & Nõlvak 1999). It is not clear, whether there is a stratigraphical gap, or the studied samples were too small.

Some problems appear also in practical application of the total range zones. In first approximation the defining of range zones was comfortable; both zonal boundaries were fixed and clear according to the data available, especially in the middle Ordovician (Nõlvak & Grahn 1993, Nõlvak 1999 –PAPERS II, IV). But later, in some sections, where new layers appear (without stratigraphical gaps, because there changes in the faunal composition are sharp), the earlier data about known ranges of important species may be already out of date. Then earlier criteria of the zonal boundaries should be redefined or changed. Such a case appeared in the Valga 10 section (see Nõlvak 2001b, p.9). Sometimes the first and last occurrences (specimens), which are as a rule rare, have some transitional features and are not typical.

Furthermore, the selection of the levels of the bases of the *Cyathochitina angusta* and *Armoricochitina reticulifera* subzones in the Viljandi core section (Fig. 2) depends on how we define transitional (*cf.*) specimens. Obviously, this problem should be analysed in many sections taking into account all chitinozoan assemblages to avoid too far-reaching stratigraphical conclusions based on occasional occurrences (sporadic data). In principle, we use mainly acme of certain chitinozoan species due to their specific nature – they are relatively very numerous, often many tens of specimens per gram of rock. Therefore the occurrences of single specimens should be identified very carefully, and often there is no use in practice, when describing a new chitinozoan species referring to one specimen, because this could be as a teratological (aberrant) form.

The real ranges of species are important to know when comparing the results of the conventional biozonation of chitinozoans obtained empirically with those got according to the algorithm STAND (Rubel & Pak 1984). The first and last occurrences appears to be decisive, despite that the results show a relatively precise and well represented chitinozoan standard: using different databases, 75–90% of chitinozoan species have been included into the standard (Nõlvak 1989). This result was the highest among the investigated groups (see Oleinikov & Rubel 1989). This study (Nõlvak 1989) confirmed that chitinozoan classification and biozonation used are quite realistic and useful.

One important regional aspect is the study of the diversity dynamics of different groups of organisms describing temporal changes in the taxonomic diversity. Diversity dynamics discloses much about evolutionary and environmental biotic changes: radiation, origination, extinction, etc. provided that adequate methods of description are used. The first results of the comparison of the chitinozoans and acritarchs in the Rapla core section (Kaljo *et al.* 1995, and 1996 –PAPER VIII) show significant correlation in the pattern of the most general processes represented by coinciding late Arenig–Llanvirn (Darriwilian) radiation–origination events, diversity rise with a maximum in early–middle Caradoc, followed by a late Caradoc diversity drop and mass extinction at the end of the Ordovician. It also advocates that some general tendences of biotic diversity might be observed in one section, but in future also in different areas of Baltoscandia (East Baltic, Scandinavia, Ukraine, Poland, etc.) such studies are needed. However, some problems could arise by the subdivision and precise correlation between the sections from those areas (see Modliński *et al.* 2002), which have to be ascertained before understanding the

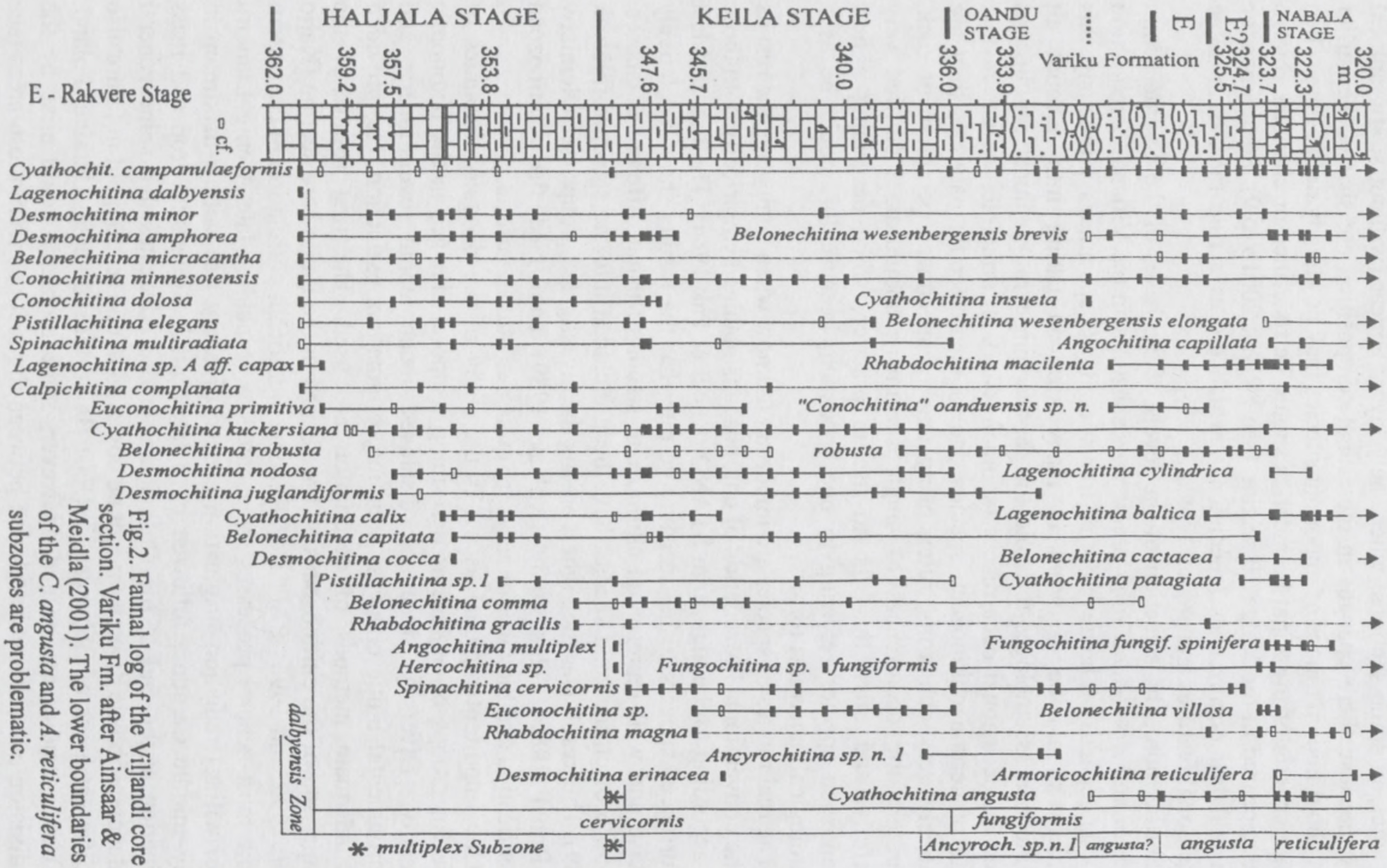

Fig.2. Faunal log of the Viljandi core section. Variku Fm. after Ainsaar & Meidla (2001). The lower boundaries of the *C. angusta* and *A. reticulifera* subzones are problematic.

causes and possible mechanisms of bioevents. Still too little data are available for the resolution of criteria of some important stratigraphical levels, e.g. lower boundaries of the Oandu or Rakvere stages in South Estonian sections (see discussion in the Valga 10 section, Põldvere 2001, p. 5 and 39–40).

The first attempt to summarize the data on 157 species of 26 genera from North, West and South Estonia, Latvia, Lithuania, Sweden (Grahn 1980, 1981, 1982a, Nõlvak *et al.* 1999 –PAPER VII, and author's observations), Norway (Grahn *et al.* 1994), Poland, Ukraine and Russia (St Petersburg Region) gave an overview about the main tendences in the chitinozoan diversity changes. They were analysed separately for each region, and the comparison shows very similar rates and order of fluctuations (originations, extinctions). Regional differences in the content of taxa are small, not exceeding 10%, which gives a good ground for the "workable" biozonation schemes (Nõlvak & Grahn 1993, Nõlvak 1999 –PAPERS II, IV). The identifications on the generic level where left out from the number of species, undescribed species were included, and only the real first and last records of species were considered. The taxonomic diversity of Ordovician chitinozoans in the whole Baltoscandia is presented in Fig. 3 using a more generalized time-scale compiled for Project 410 (see below).

After the first chitinozoan occurrences in the Tremadoc (*Lagenochitina esthonica* and *L. destombesi* in the topmost Varangu Stage from a single section and sample both in North Estonia and Latvia), the most intensive origination and diversification took place in the late Volkhov time (early Darriwilian). It is remarkable that the number of investigated samples has no direct bearing on the sudden rise of all curves (Fig. 3). In the Lower Ordovician, except rare finds in North Estonia, condensed beds of the Hunneberg and Billingen stages in the other regions of Baltoscandia turned to be barren due to the preservational reasons (see above).

In late Volkhov and Kunda time the number of short-ranging species was relatively large, and the origination and extinction rates reached their maximum values. Higher the diversity curve shows rising values up to the maximum (35 species) in the Uhaku Stage (latest Darriwilian) with numerous short-ranging species, but the origination rate curve shows a small drop. In younger beds (from Haljala to Nabala stages) the origination and species diversity curves are relatively stable, but the generalized time-scale in the Project 410 is too generalized and the noticeably clear change in the diversity, recovered at the Keila–Oandu boundary in Rapla section (Kaljo *et al.* 1996 –PAPER VIII), is not reflected here (Fig. 3). The drop of the diversity and origination curves starts near the end of the Caradoc (in the Vormsi Stage) and continues in lower Pirgu Stage partly due to the wide distribution of redbeds in this part of the succession. In the upper Pirgu Stage origination intensifies during a short period, followed by major extinction (Late Ordovician crisis) in the Porkuni deposits (see below), where *Conochitina scabra* appears to continue, together with only three other species, into the Silurian (Fig. 3).

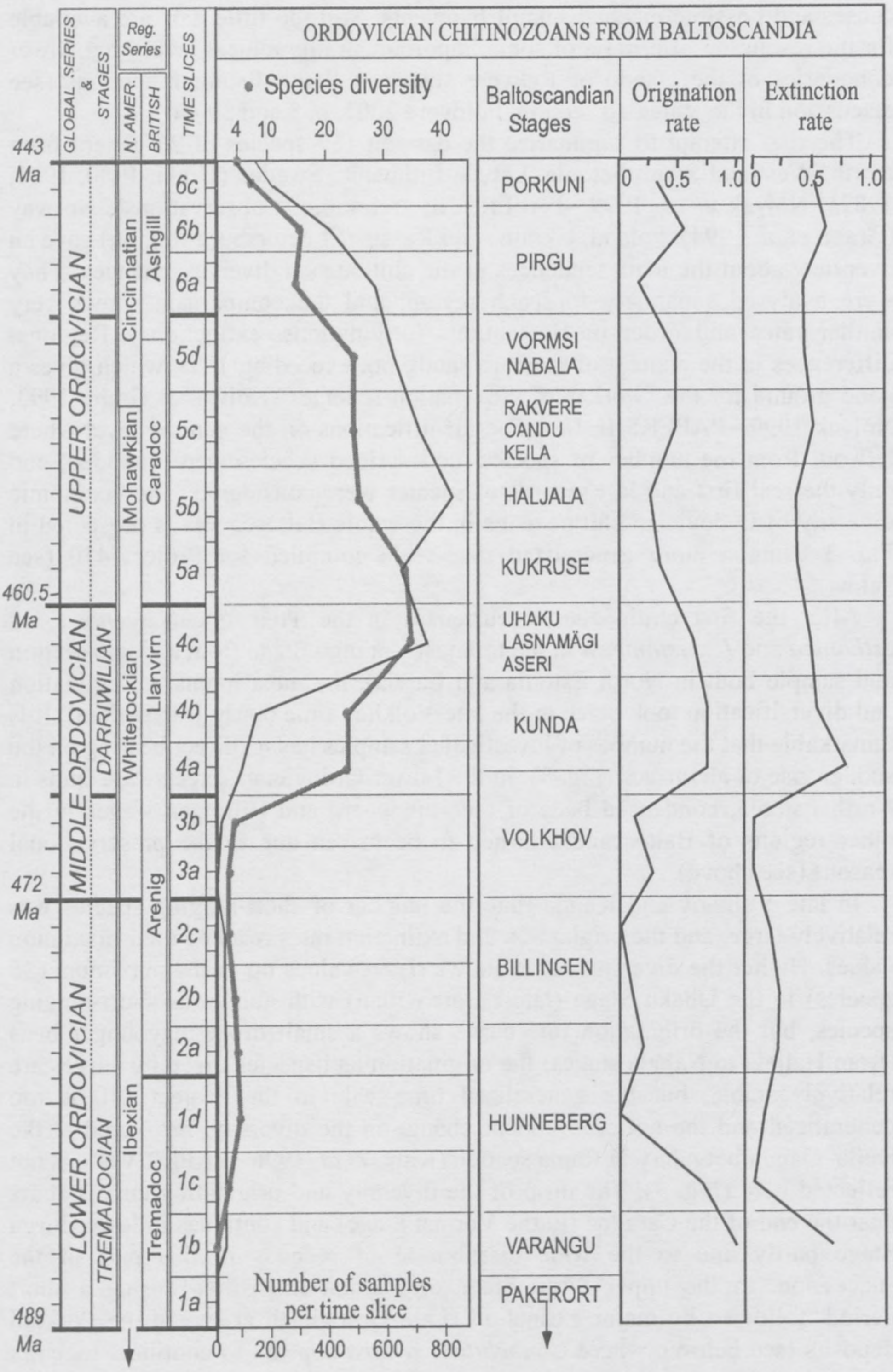

Fig. 3. Biodiversity of Ordovician chitinozoan species in Baltoscandia.
Time slices after B. Webby, R. Cooper, S. Bergström & F. Paris (IGCP Project No.410).

Although several genera show the cosmopolitan behaviour and other genera have been palaeogeographically differentiated into latitudinal bands, especially in the Ordovician, where separate biozonations have been erected for Baltica, North Gondwana and Laurentia (see Paris *et al.* 1999 and references therein – PAPER X) it is possibile to document the correlation between the available chitinozoan biozones on the species level. However, well-known difficulties arise: palaeocontinents were dispersed and their specific configuration had important consequences for the provincialism of the Ordovician faunas. The distribution of pelagic faunas is mostly controlled by temperature and indirectly by palaeolatitude and hence by climate. This pattern of distribution is subsequently modified by palaeocurrents. Mainly for these reasons a global chitinozoan zonation for the Ordovician has not been achieved.

The main goal of the IGCP Project 410 is to provide a more comprehensive global view of the Ordovician biotas, focusing on their patterns and significance of appearances, on the diversification of each major group and the nature and extent of faunal extinctions during the Period. For that purpose a new Ordovician time-scale (working version, see Fig. 3) using three main groups of fossils: graptolites, conodonts and chitinozoans. Their distributional data (zonal schemes) are plotted on the chronostratigraphic column. This is subdivided into 16 time slices and collated with the main regional chitinozoan zonations from Baltoscandia, North Gondwana and Laurentia. Besides the constructing of the global biodiversity curve of the Ordovician, one goal was to register the local anomalies in the diversity of the group (Paris *et al.* 2001; Nõlvak 2001a). In this first attempt the Baltoscandian zonal scheme of chitinozoans is considered to represent the whole Baltica palaeoplate (Achab *et al.* 2002), that embraces about half of the samples used in this contribution from the whole world.

The first comparisons between zonations show that the distinctiveness of the regional microfaunas can also be expressed in terms of common species. Of 390 known and unanimously accepted (*pers. comm.*, F. Paris *et al.*) Ordovician species only 21 are shared by the three palaeoplates and 38 species have been recorded from two palaeoplates. During the Darriwilian (from the upper Volkhov to upper Uhaku stages) to the lower Caradoc (Kukruse–Haljala stages) common species are recognized in Baltica and North Gondwana, and during the Late Ordovician more open communication has existed between Baltica and Laurentia (Paris *et al.* 1999, –PAPER X). The investigated sections in the latter two palaeoplates show great similarities during late Pirgu and early Porkuni times corresponding to the upper *Conochitina rugata, Belonechitina gamachiana* and *Spinachitina taugourdeaui* zones. According to the new data (e.g. in the Kardla core section, Fig. 4), it seems reasonable to erect a new zone, *Tanuchitina anticostiensis*, which is known from Laurentia (Achab 1989) and gives more evidence about the similarities in the vertical distribution and zonation of chitinozoans.

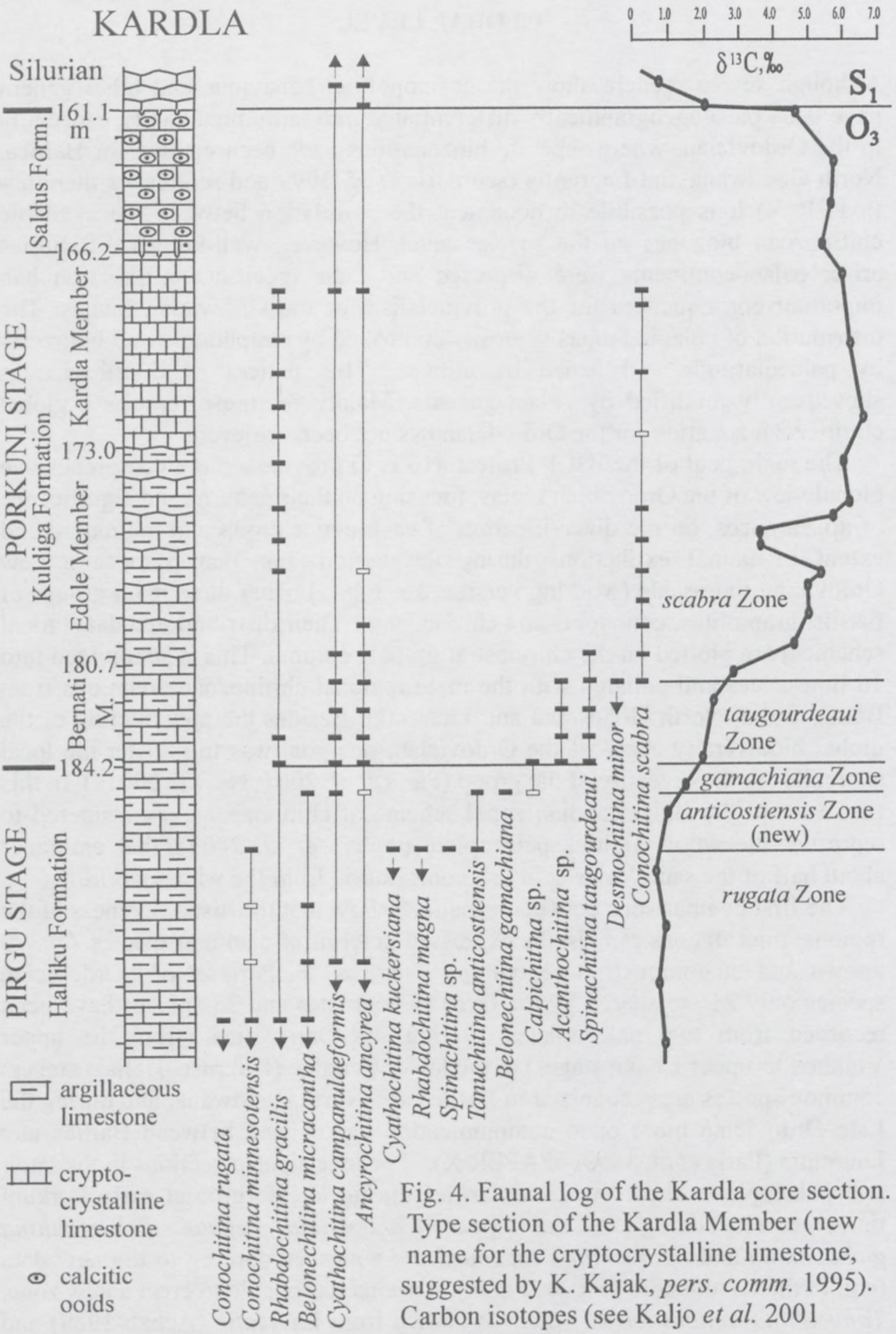

Fig. 4. Faunal log of the Kardla core section. Type section of the Kardla Member (new name for the cryptocrystalline limestone, suggested by K. Kajak, *pers. comm.* 1995). Carbon isotopes (see Kaljo *et al.* 2001 -PAPER XIII).

An example of a new zone for Baltoscandia: *T. anticostiensis* erected in Quebec and correlated with the *D. anceps* Zone (Achab 1989), in the Estonian sections in the topmost part of the *C. rugata* Zone and just below the *B. gamachiana* Zone.

At the same time, it seems more realistic that the topmost beds of the Porkuni Stage in Baltoscandia (*Conochitina scabra* Zone) are younger and more widely distributed than those in the Anticosti Island section (Kaljo *et al.* 2001 –PAPER XIII). The youngest zone – *Ancyrochitina ellisbayensis* of Anticosti (Soufiane *et al.* 1999; Soufiane & Achab 2000, p. 94) could belong already to the Silurian (see Long & Copper 1987, p. 1818).

However, not all parts of the Ordovician succession in different palaeocontinents are equally exposed and studied and often there exists strict connection between diversity and available data curves: the higher the number of samples, the higher the diversity. It means that data about chitinozoan distribution from some portions of Ordovician are poor. In Baltoscandia such direct relation is not observable (see Fig. 3), which shows the relatively high level of study.

Despite differences all three named palaeoplates have also some points in common. The pulses in the diversity and origination rate are concentrated in the Darriwilian, and chitinozoan diversity starts to decline in the upper Caradoc, reaching the minimum level in the upper Ashgill. This observation implies that the chitinozoan microfauna was under environmental pressure beginning with the late Caradoc, and the Gondwanan glaciation was the final stroke for its populations. The level of that stroke should be followed at the upper boundary of the *Spinachitina taugourdeaui* Zone in Baltoscandia and Laurentia, but most probably is not reflected in the North Gondwanan NE Algerian Sahara well NI-2 section (Paris *et al.* 2000).

This intriguing conclusion can be explained by the suggestions that (1) the zonal species *Spinachitina oulebsiri* Paris *et al.* is conspecific with *S. coronata* (Eisenack) from the Baltoscandian lower Pirgu Stage. (2) The association (incl. *Lagenochitina baltica*, *L. prussica, Rhabdochitina magna, Desmochitina minor, Tanuchitina elongata (= T. bergstroemi*, see Paris 1990; Achab & Asselin 1995, p. 76), *Belonechitina micracantha*) is very similar to that found in the lower Pirgu Stage in the North Estonian Confacies Belt, in beds below the *Conochitina rugata* Zone. It means that chitinozoans in this Algerian section could be older than suggested (Paris *et al.* 2000) and belong to the *Tanuchitina bergstroemi* Zone. The Late Ordovician crisis among chitinozoans is probably not reflected in this section, it is in a gap. If such comparison gets proved by chemostratigraphy (e.g. with low values of carbon isotopes, see Marshall *et al.* 1997), then the stratigraphical hiatus is longer in this Saharan section. Described glacio-marine diamicrites, interpreted as resulting from the melting of the ice cap (Paris *et al.* 2000, p. 90) could be 1–2 Ma older than glaciation "fingerprints" in the farther-lying palaeocontinents.

3. DATING OF GEOLOGICAL EVENTS

Besides the subdivision and correlation of sections, which serves as a basis in geological mapping, chitinozoan zonation can also be used in dating of some geological events, such as volcanic eruptions, impacts, abrupt sea-level changes and extinctions, or sedimentological breaks. The Baltoscandian Ordovician succession contains fingerprints of such big events that can be followed in the fossil record. However, a question is how precisely we can date them according to the commonly used (available) timescale. In the following some examples of the use of microfossil data in practice are given.

PRACTICAL USE

MASS EXTINCTION EVENT. From 157 known species of chitinozoans only 4 can be distinguished in the topmost Ordovician beds, which cross the Ordovician–Silurian boundary, but higher some *Cyathochitina* species reappear in the transitional beds of the Juuru and Raikküla stages (Nestor 1994). The abrupt change among chitinozoans in Baltoscandia in the Porkuni Stage is described in several papers (Nõlvak *et al.* 1989 –PAPER VI; Brenchley *et al.* 1996; Brenchley *et al*. 1997 –PAPER XII; Kaljo *et al.* 1998; Kaljo *et al.* 1999 –PAPER XI; Kaljo *et al.* 2001 –PAPER XIII). This level can be followed among others in the Kardla core section (Fig. 4), where the beds under discussion are lithologically without clear features of breaks in sedimentation. So, it is one of the few sections, where the *Spinachitina taugourdeaui* Zone has its natural boundaries, without barren dolomites above (in North Estonia) or redbeds below this zone (in South Estonia). When analysing the reasons for that stroke, one can speculate that the water was too warm for chitinozoans during the regression maximum.

In a many aspects similar situation can be followed in the Rapla section in the latest Keila time: 63% of the species disappeared – a substantial extinction event (Hints *et al.* 1989, Kaljo *et al.* 1996; fig. 5 –PAPER VIII; Meidla *et al.* 1999). This event is contemporaneous with the appearence of the reefs (Vasalemma Formation in northwestern Estonia and Kullsberg Limestone in Dalarna, Sweden). However, in the Viljandi section (Fig. 2) this change (extinction event) can be followed at a level of 336.0 m, and the beds in the top of the Keila Stage (interval 336–338 m) have conspecific chitinozoan assemblage with that from the Lehtmetsa Member having approximately the same age with reefs (see Põlma *et al.* 1988, fig. 32). Breaks in sedimentation characteristic of the North Estonian sections (below the Hirmuse Formation) are not established in Viljandi section. Anyhow, the gap at the boundary of the Keila and Oandu stages seems to be not the only cause of the described abrupt change among chitinozoans in the Rapla section (Kaljo *et al.* 1996 –PAPER VIII); the sea may have been too warm at that time.

In several papers main attention is paid to the carbon isotope stratigraphy in the latest Ordovician, where the correlation between chitinozoan biodiversity and the carbon isotope curve can be observed (e.g. in Kardla, Fig. 4), showing that both were influenced by the same environmental agents (Kaljo *et al.* 1999, 2001 –PAPERS XI, XIII). In this portion (upper Pirgu–Porkuni stages), among investigated fossils chitinozoans allowed the most precise subdivision and correlation, and were used as stratigraphical background in comparisons of isotope curves.

ERRATIC BOULDERS. Eisenack's pioneering studies in the classification of chitinozoans was based mainly on the information collected from erratic boulders (see Eisenack 1968 for overview) and nowadays approximately half of them could be indentified also stratigraphically. Anyway, although there is no close stratigraphical connexion, information from a large submarine area of the Ordovician sequence between West Estonian and Swedish Islands, but also from the Bothnian Sea (see Tynni 1975, Uutela 1989 for overview and references therein; Nõlvak *et al.* 1997) is lacking and erratic boulders are still the only source of knowledge about facies distribution in these regions. However, there are well-known limits set by the composition of rocks – only hard varieties of limestone have been saved during the transport, but the preservation of fossils is often very good for palaeontological studies.

During the study of Ordovician erratics in Åland Islands interesting rich fauna were collected from the boulders in Överby in the surroundings of the Lumparn impact crater. The timespan of partly strongly argillaceous limestones includes equivalents from the Aseri to Nabala stages, indicating the possibility of a not very distant source area. It may be even in Åland Islands, because a large limestone area under the Bothnian Sea seems to be too far for so argillaceous rocks to have persisted during glacial transportation.

VOLCANIC SEDIMENTS. The influence of volcanic ashfalls on the marine environment has been studied from Männamaa and Rapla sections (Hints & Nõlvak 1999, figs. 4, 5). The analysis of the detailed bed-by-bed series of samples from the beds near the Middle Ordovician Kinnekulle K-bentonite revealed that this influence to the chitinozoan fauna was much smaller than hitherto presumed (Nõlvak 1996, 2002) due to the planktic mode of life suggested for chitinozoans. No remarkable changes took place in the content of the chitinozoan fauna in these beds, except, as a specific feature, just above the Kinnekulle K-bentonite where the zonal form *Angochitina multiplex* appeared. These aspects together allow us to trace this horizon with very high precision.

Another example of datings among Ordovician volcanic events concerns the relations between the Avalonia and Baltica palaeocontinents. The results of the detailed biostratigraphy by chitinozoans from the Brabant Massif, Belgium and correlations with the Baltoscandian chitinozoan zonation (Nõlvak & Grahn 1993 –PAPER II) enable precise dating of the subduction related to the Late Ordovician volcanism in the Brabant Massif. This corresponds to the lower and middle parts of the Baltoscandian Pirgu Stage (Van Grootel *et al.* 1996 p. 69), indicating the time of about 447–449 Ma. It is interesting to note that bentonite layers of about the same age occur in Estonian sections: in one level in the upper Vormsi Stage (cores no. 309, 146, Padaküla; Nõlvak 1987b) and at least in three levels in the lower Pirgu Stage in many sections (see Nõlvak 1984, 1986). In Belgium this part of the sections corresponds to the *Tanuchitina bergstroemi* and *Conochitina rugata* zones, showing approximately the same age.

IMPACT CRATERS. Detailed information about chitinozoan distribution allowed precise dating of all four Ordovician impact craters known in Baltocsandia. The craters are not the same age as earlier suggested (Grahn & Nõlvak 1993): Granby – early Kunda, Tvären – late Kukruse, Kärdla and Lockne – early Haljala. Later many detailed studies (Grahn *et al.* 1996 – PAPER IX; Nõlvak 1996; Ormö *et al.* 1998; Sturkell *et al*. 2000) of the fillings of those craters were carried out. More precise datings are effective when series of chitinozoan samples are available from the sections outside the structures, as near Kärdla (in the Männamaa section, Grahn *et al.* 1996, fig. 9 –PAPER IX) and Lockne (in the Hallen section, Sturkell *et al.* 2000, fig. 2).

Lumparn crater (Åland Islands, southwestern Finland) is dated as older than Lower Ordovician (Abels *et al.* 1998). It has become evident that the beds within the crater are not only of Nabala age (Merrill 1979; Tynni 1982), and chitinozoans indicate the Haljala–Keila time, represented by beds of the *Spinachitina cervicornis* Zone. All efforts to find some organic-walled microfossils in the topmost beds inside the Lumparn crater (up to 38 m thick in the borehole core Tranvik 7) have been unsuccessful because of the abundance of red spots in the cryptocrystalline (aphanitic) limestones. In part this rock is totally rose-coloured or tan, probably due to the influence of the Devonian sediments – totally eroded red sandstones, with age not earlier than the lower Middle Devonian. If this is correct, it can be suggested that the organic-walled fossils are destroyed under the influence of the Devonian cover on the underlying aphanitic limestones (secondary processes) in the same manner as in Ordovician redbeds (described above). It is evident that the rock type is important for the preservation of chitinozoans – they are less influenced in more argillaceous limestones and marls with red spots, as in the beds around the red "Jonstorp-facies" not depending on the age of the beds, as in the Halliku or Tudulinna formations in the South Estonian sections (resp. Pirgu or Vormsi stages).

GAPS. The most extensive gap in the chitinozoan distribution of about 13–14 Ma occurs in the Lower Ordovician (Hunneberg and Billingen stages) and in the lowermost Middle Ordovician (lower Volkhov Stage). It has mainly sedimentological but also preservational reasons, which diminishes the usefulness of chitinozoans in the stratigraphy of that part of the Baltoscandian succession. The gap can be observed clearly in the chitinozoan zonation schemes: the succession is less stratified and the vertical range of chitinozoan zones is relatively long compared to graptolites and conodonts (see Nõlvak &

Grahn 1993; Nõlvak 1999 –PAPERS II, IV). However, more detailed studies may give some new zones (unpublished), e.g. in the *Cyathochitina regnelli* Zone, near the base of the Darriwilian, which is an important stratigraphical level for global correlations.

Most of the Estonian chronostratigraphic units (regional stages) are defined as intervals characterized by a particular composition of rocks. Considerable breaks in sedimentation have been revealed and often used on the boundaries of most chronostratigraphic units established in the North Estonian Confacies Belt (see Männil 1966; Rõõmusoks 1983 and references therein). However, problems in correlation arise when in some other regions of the Baltoscandian basin the beds, which fulfilled these gaps, contain earlier unknown ranges of fossils considered as typical of a certain subdivision (see Hints *et al.* 1995; Meidla *et al.* 1999, fig. 6). The designation of the boundary-stratotypes might be of some help (Hints & Nõlvak 1999). Nevertheless, some difficulties appear when selecting the boundary stratotype in borehole cores without gaps, which is correct, but practically the limited amount of rock material in subsurface sections (borehole cores) is then the main restriction. Moreover, precise definitions of boundaries by only one microfossil group leave such definitions often questionable.

A disputable stratigraphical gap occurs between the Nabala and Vormsi stages according to the distribution of chitinozoans (Nõlvak 1980, fig.2, –PAPER V), but questioned by Hints & Meidla (1997, p.82; Ainsaar & Meidla 2001) by ostracodes. Further detailed studies are needed, but in support of the conclusion about a regional hiatus the geophysical data from the northern Baltic (Tuuling 1998) can be submitted, which show clear changes and a prominent gap in the sequence.

On the other hand, using series of chitinozoan samples it is possible to find also relatively very short hiatuses in sedimentation, e.g. in the boundary beds of the Kukruse and Haljala stages (Nõlvak 1972; Nõlvak *et al.* 1999 –PAPER VII; Sturkell *et al*. 2000).

HISTORICAL OBJECTS. (1) The first object was a small piece of rock from the portal of Niguliste Cathedral in Tallinn (first mentioned in 1308). The question was simple: from where was this rock material taken? By historical data (V. Raam, *pers. comm.* early 1970s) the masters were from Sweden, who may have taken the material with them. According to chitinozoans this portal was made of local "Building Limestone" of the Lasnamägi–lowermost Uhaku stages (by present classification) and not of the Middle Ordovician limestone from Öland, or Silurian limestone from Gotland, which were widely used in Sweden.

(2) Very interesting objects were limestones from the wrecks of two drowned medieval ships found in the Gulf of Finland (near Aegna Island) and from the Oslofjord (Norway). Up to 10 cm thick plates of hard limestone measuring at least 0.5 x 1 m were discovered. According to the content of chitinozoans and lithology, these plates were both from the layers

corresponding to the *C. clavaherculi* Subzone of the *C. striata* Zone belonging to the upper Lasnamägi Stage (also "Building Limestone") and were probably taken from a quarry near Tallinn as a ballast to keep the ship steady.

(3) Many dredged samples of "Baltic Limestones" from different localities of the Baltic were studied. These data gave additional information for mapping, but often there was some uncertainity about their precise age and only joint study of different microfossil groups can give a reliable result (see Nõlvak *et al*. 1997).

(4) Interesting objects were also some medieval memorial plates or ledgers and building stones, where microfossils have clear advantage in efforts to throw light on their age and origin. In Old Tallinn these were mainly of the Ordovician "Building Limestone", but more decorative ones of the Silurian dolostones (see Perens & Kala 2002).

The above confirms that biomicrostratigraphy of chitinozoans is needed.

CONCLUSIONS

The very brief history of the investigations of Ordovician chitinozoans in Estonia can be tentatively divided into three periods focusing on different aspects. First of all the discovery of this group by Eisenack (1931) and his palaeontological studies (for overview see Eisenack 1968). Then, from the late 1960s the first observations about the stratigraphical applicability of some chitinozoan species during the compilation of several stratigraphical charts for the whole Baltoscandia and microbiostratigraphy of kukersite-bearing beds in one confacies belt by Männil (1972, 1986, 1989), when the author had his first experience in chitinozoan study (Nõlvak 1972). Later, the author carried out the zonation of chitinozoans independently for the upper Ordovician (Nõlvak 1984, 1986, 1988) and together with Grahn (Nõlvak & Grahn 1993 –PAPER II) for the whole Baltoscandia. For the first time data from the whole Ordovician succession became available (Rapla core section). The author has continued this approach and this has given him a possibility of taking part in global projects (e.g. Paris *et al*. 1999 –PAPER X; IGCP Project 410; etc.). The studies through the whole Ordovician in different areas of Baltoscandia are of great importance for the systematics and biostratigraphy of chitinozoans. To sum up, the following aspects should be noted.

(1) It is obvious that practically we should not go too far with the density (see discussion in Jaanusson 1976, p. 303) and amount of rock samples. Long experience shows that only 50 grams (often used in early studies) for a limestone sample is not representative, at least five to ten times more is needed.

(2) One of the main aims was to elucidate in which part of the Ordovician basin and in what levels chitinozoan biostratigraphy can give a profitable result. In all regions where rocks are preserved it appears to be Middle and Upper Ordovician (Darriwilian, Caradoc and Ashgill – from upper Volkhov to Porkuni stages), except the redbed parts in Baltoscandia. Compared to other fossil

groups, the most precise differentiation among chitinozoans can be followed in the upper Caradoc and Ashgill (from Nabala to lower Porkuni stages), in beds where they are preserved, being the best tool for high-resolution correlations.

(3) Different steps and reliability of chitinozoan zonation are used: above-described zonules (Fig. 1) should be used only in one confacies belt together with some other (lithological) features. Some range zones presently in use might be revised (see Nõlvak 2001b).

(4) It may not be fruitful and practically needed to follow in future all traditional regional subdivisions (stages, see Männil 1966 and references therein) and to make every effort to find boundaries in some certain section, as in case of the earlier discussed Oandu–Rakvere, or Nabala–Vormsi boundary problems. These historical subdivisions were erected without definitions of their boundaries. Only one boundary stratotype has been erected: Haljala–Keila (Hints & Nõlvak 1999); potentially there are some more suitable outcrop sections, but not for every subdivision (see Hints *et al.* 1995, fig. 3).

(5) For the global comparisons it seems more interesting and reasonable to concentrate joint efforts on those levels, which were selected (IGCP Project 410) as boundaries of the Ordovician time slices (4c – 6a, 6c; see Fig. 3) and in six levels coincide with the boundaries of the East Baltic Regional Stages. These are the lower boundaries of the Aseri, Kukruse, Keila, Nabala, Pirgu and Porkuni stages. This selection is mostly based upon our detailed and reliable chitinozoan zonation in Baltoscandia, which has proved its usefulness in global correlations and is one of the best results in our work.

One of the main future tasks is to determine how much time the preserved sediments represent in the Ordovician in the Baltoscandian succession and how long the hiatuses were. Detailed zonation of chitinozoans can be successfully used owing to their best preservation in the world in some portions of the Ordovician.

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KITIINIKUTE BIOSTRATIGRAAFIA BALTOSCANDIA ORDOVIITSIUMIS

Käesolevas väitekirjas, mis koosneb sidustekstist ja varemtrükitud artiklitest, on käsitletud kitiinikute päritolu, nende jaotumust ja levikut ning nende kasutamise võimalusi erinevate geoloogiliste sündmuste dateerimisel.

Pikaajalise uurimistöö käigus on selgunud, et kitiinikud kujutasid endast varapaleosoikumis (Baltoskandias Ordoviitsiumis ja Siluris) elanud ja planktoni hulka kuulunud, kuid senini ebaselge päritoluga ning tänapäevaks väljasurnud oleste orgaanilise kestaga mikrokivististe rühma. Tõenäoliselt vähemalt osa neist on kivimites mittesäilinud organismide munad, millede liigilises kooseisus toimunud suhteliselt kiired muutused võimaldavad geoloogilisi läbilõikeid liigestada ja korreleerida. Kokkuleppelise taksonoomia alusel kirjeldatavate liikide levik on erinev. Osa neist olid väga lühiajalise elueaga, reageerides kiiresti keskkonna muutustele. Need on aluseks vertikaalse liigestusskeemi koostamisel, milles peetakse oluliseks koosseisu muutuste järjestust. Ilmneb, et Baltikumi, Skandinaavia ja sellele lähedaste alade Ordoviitsiumi settekivimites on kitiinikute jaotumus piisavalt sarnane, et koostada erineva territoriaalse kasutusala ning detailsusega tsonaalseid skeeme, mille alusel saab kihte rööbistada ja suhtelist vanust määrata. Töö tulemused võimaldavad võrdlust ka erinevate paleokontinentide vahel, mis näitab, et eriti Ülem-Ordoviitsiumi settekivimites on kitiinikute jaotumus tunduvalt detailsem võrreldes teiste seniuuritud faunarühmadega. Kõrgtäpsusega biostratigraafilise uurimistöö tulemus Baltoscandia läbilõigetest on aluseks ka kuue taseme valikul uue globaalse korrelatsiooniskeemi koostamisel rahvusvahelise projekti (IGCP 410 raames).

Kitiinikud ilmusid Baltoskandia Ordoviitsiumi läbilõikesse Tremadoc'is, Varangu lademe ülemistes kihtides. Sellele järgneva umbes 16 miljoni aasta vältel olid kitiinikute kooslused liigivaesed ja sporaadilised ning eraldatud üksteisest pikaajaliste settelünkadega. Darriwilianist (Volhovi ea teisest poolest) alates toimub intensiivne koosluste mitmekesistumine, mis võimaldab järsult tõsta korrelatsioonitasemete arvu ja biotsonaalsete skeemide liigestuse detailsust. Mõningane mitmekesisuse vähenemine, lokaalsete tõusude taustal, on jälgitav Ashgilli algusest alates (Nabala–Vormsi eal), millele järgneb Ordoviitsiumi kitiinikute drastiline väljasuremine seoses globaalse Hirnantia jääajaga Porkuni ea teise poolel. Selline üldine areng on jälgitav nii regionaalsel kui ka globaalsel tasemel, mis ei välista aga lokaalseid erinevusi.

Kitiinikute leviku alusel koostatud ajaskaala taustal on määratud ka mitmete Ordoviitsiumiaegsete geoloogiliste sündmuste nagu settelüngad, vulkaanilised setted, meteoriidikraatrid, või objektide (rändkivid, ajaloolised ehitusmaterjalid, arheoloogilised leiud, jt.) suhtelist vanust, tähtsustades uurimistöö praktilist väärtust.

ORIGINAL PUBLICATIONS

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Paper I

Paris, F. & Nõlvak, J. 1999. Biological interpretation and paleobiodiversity of a cryptic fossil group: the "chitinozoan animal". *Geobios*, **32,** 2, 315-324.

BIOLOGICAL INTERPRETATION AND PALEOBIODIVERSITY OF A CRYPTIC FOSSIL GROUP: THE "CHITINOZOAN ANIMAL"

FLORENTIN **PARIS** & JAAK **NÕLVAK**

PARIS F. & NÕLVAK J. 1999. Biological interpretation and paleobiodiversity of a cryptic fossil group: the "chitinozoan animal". [Interprétation biologique et paléobiodiversité d'un groupe fossile cryptique: "l'animal-chitinozoaire"]. GEOBIOS, **32**, **2**: 315-324. Villeurbanne, le 30.04.1999.

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ABSTRACT - Chitinozoans are used to evaluate, at the species level the variations of diversity within a cryptic fossil group from their emergence to their ultimate demise. The chitinozoans appear to be a group suitable for such investigations because of their fairly restricted range, world-wide distribution and rather limited number of described species. However, before using this group, it was necessary to demonstrate that chitinozoans represent an ontogenic stage (eggs) in the life-cycle of a cryptic group (i.e. virtually non fossilisable soft bodied organisms). The eggs register the variations of diversity affecting the group and are therefore valid indicators of species diversity. Our results reveal the dramatic diversification of the group in its early stage, reflecting the incorporation of numerous innovations and an accelerated radiation phase lasting twenty millions years, from the late Tremadoc to the Llanvirn. The end-Ordovician faunal crisis and its Eo-Silurian recovery are also well documented. The decline of the group occurred as early as the Lower Devonian and its extinction at the very end of the Devonian, after a brief increase in diversity during the Givetian. However, depending on the time units used, the shape of the biodiversity curves varies significantly and some artefacts may be present. At the series level, only the major tendencies are recorded, whereas at the stage level, additional second-order signals, not firmly related to abiotic events, are revealed.

KEYWORDS: CHITINOZOANS, EGGS, CRYPTIC FOSSILS, BIODIVERSITY, ONTOGENY, PALEOZOIC.

RÉSUMÉ - Nous utilisons les chitinozoaires pour évaluer les variations de diversité au niveau spécifique, au sein d'un groupe fossile cryptique, depuis son émergence jusqu'à son extinction définitive. Par sa durée assez brève, sa distribution quasi mondiale et le nombre relativement limité des espèces qui le constituent, ce groupe se prête en effet assez bien à ce type d'analyse. Il existait cependant un préalable à l'exploitation de ce matériel. Il fallait démontrer que ces chitinozoaires représentent un stade de l'ontogénie, en l'occurrence les oeufs, d'organismes à corps mou difficilement fossilisables et, qu'à ce titre, ils ont également enregistré les fluctuations de biodiversité du groupe cryptique qu'ils matérialisent. Nos résultats, obtenus à partir de l'exploitation d'une base de données, soulignent la spectaculaire biodiversification initiale du groupe, correspondant à la mise en place de nombreuses innovations et à une phase de radiation accélérée pendant une vingtaine de millions d'années, du Tremadoc supérieur au Llanvirn. La crise fini-ordovicienne et sa phase de cicatrisation éo-silurienne sont bien documentées. Le déclin du groupe s'amorce dès le Dévonien inférieur et son extinction définitive intervient au Dévonien terminal, malgré une légère recrudescence de la diversité spécifique au Givetien. La durée des unités de temps choisies pour le calibrage temporel influence nettement la forme des courbes de biodiversité obtenues et peut même introduire des artefacts. Le découpage au niveau des séries ne révèle que les tendances majeures alors qu'un découpage plus fin, au niveau de l'étage, met en évidence des signaux de second ordre que nous n'arrivons pas à relier de façon indiscutable aux événements abiotiques déjà connus.

MOTS-CLÉS: CHITINOZOAIRES, OEUFS, FOSSILES CRYPTIQUES, BIODIVERSITÉ, ONTOGENÉSE, PALÉOZOÏQUE

INTRODUCTION

Variations of faunal and floral biodiversity, usually expressed at family or sometimes at the generic level, have been widely used to quantify the impact of terrestrial or extra-terrestrial events on the composition of the biosphere trough time (e.g. Sepkosky 1995a,b). This approach has been extensively employed to document the effects of the Cretaceous-Tertiary Event on the extinction, versus radiation, of both marine and terrestrial organisms in the Late Cenomanian and the Early Paleocene (see references in Sheehan et al. 1996)

In the present paper, our approach is different, because our goal is to depict the pattern and the tempo of the biodiversification, at the species level, of a single fossil group, from its emergence to its final extinction. We have selected the chitinozoan group to illustrate such biodiversification

FIGURE 1 - Geographic distribution of the main chitinozoan-yielding localities (from Paris 1996, updated). *Distribution géographique des principaux gisements de chitinozoaires (d'après Paris 1996, mis à jour).*

pattern for several reasons: 1) the range of the group, from the late Tremadoc to the topmost Devonian, is well documented, 2) the group has a world-wide distribution and was only weakly affected by climatic and paleogeographic controls, especially during the Silurian and the Devonian when the principal paleocontinents were converging, 3) the group contains a moderate number of species (1078 species recorded up to the end of 1997), and 4) a taxonomic database (Paris & Bernard 1994) is available, including all the chitinozoan species described to date. After some slight modifications [e.g. indication of the First Appearance Datum (FAD) and of the Last

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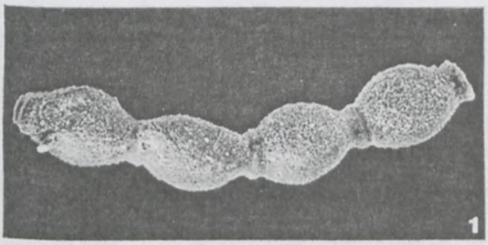
FIGURE 2 - **1.** Simple chain of *D.* cf. *nodosa* EISENACK; Ch 940/4360, Värka borehole, depth 369.0 m, Jõhri Substage of Haljala Stage, southwestern Estonia [x160]. **2.** Operculum in situ showing the sealing of the vesicle of *D. nodosa* (note the central scar in the centre of the operculum, corresponding to the remains of its linkage to the bottom of the neighbouring vesicle); Ch 925/7931, Männamaa borehole, depth 164.4 m, Idavere Substage of Haljala Stage, western Estonia [x 1200]. **3.** Coiled catenary structure (helical spring-like structure) of *D. nodosa*; Ch 928/6799, Ketrzyn borehole, depth 1573.6 m, Idavere Substage of Haljala Stage, northern Poland [x 120]. **4.** Detail of the linkage of the structure represented on figure 2.3 [x 425]. **5.** Cluster of *D. nodosa* interpreted as the laying of the corresponding "chitinozoan animal" species (note the opercula still in place indicating an infertile laying); Ch 923/7931, Männamaa borehole, depth 164.4 m, Idavere Substage of Haljala Stage, western Estonia [x 110]. **6.** Detail of the preceding figure showing the vesicle arrangement of the opercula still in place [x 120]. **7.** Large cluster of *D. nodosa* interpreted as a laying of the corresponding "chitinozoan animal" species; most of the opercula are removed (the juveniles may have escaped from the vesicles); Ch 938/7970, Tartu borehole, depth 325.4 m, Idavere Substage of Haljala Stage, southern Estonia [x 70]. **8.** Detail of the preceding figure showing the arrangement of the vesicles, their opening and empty chambers [x 220]. ***1.** Chaîne simple chez* D. *cf.* nodosa *EISENACK; Ch 940/4360, forage de Värka, 369,0 m, sous étage Jõhri, étage Haljala, Sud-Ouest de l'Estonie [x 1200]. **2.** Opercule en place montrant la fermeture de la loge de* D. nodosa *(noter la cicatrice au centre de l'opercule, correspondant aux restes de sa fixation sur fond de la loge la plus proche); Ch 925/7931, forage de Männamaa,164,4 m, sous étage Idavere, étage Haljala, Ouest de l'Estonie [x 1200] **3.** Structure caténaire hélicoïdale (en "ressort à boudin") chez* D. nodosa*; Ch 928/6799, forage de Ketrzyn, 1573,6 m, sous étage Idavere, étage Haljala, Nord de la Pologne [x 120]. **4.** Détail des liaisons interloges de la structure représentée sur la figure 2.3 [x 425]. **5.** Agrégat de* D. nodosa, *interprété comme une ponte de "l'animal-chitinozoaire" correspondant à cette espèce (noter les opercules toujours en place suggérant une ponte non fertile); Ch 923/7931, forage de Männamaa, 164,4 m, sous étage Idavere, étage Haljala, Ouest de l'Estonie [x 110]. **6.** Détail de la figure précédente montrant l'agencement des loges et les opercules en place [x 120]. **7.** Large agrégat de* D. nodosa, *interprété comme une ponte de "l'animal-chitinozoaire" correspondant à cette espèce; la plupart des opercules sont partis, indiquant que les larves ont pu s'échapper des loges; Ch 938/7970, forage de Tartu, 325,4 m, sous étage Idavere, étage Haljala, Sud de l'Estonie [x 70]. **8.** Détail de la figure précédente montrant l'agencement régulier des loges et leur ouverture s'ouvrant sur une panse vide [x 220].*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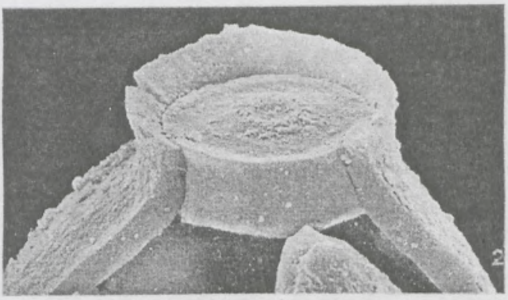
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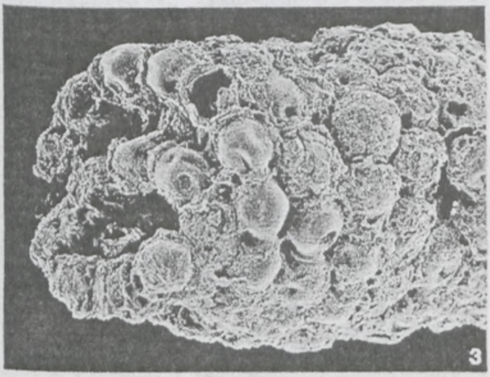
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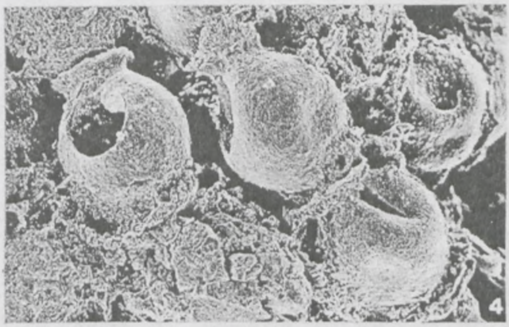
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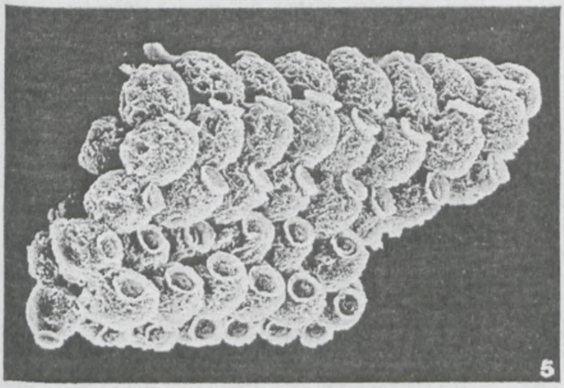
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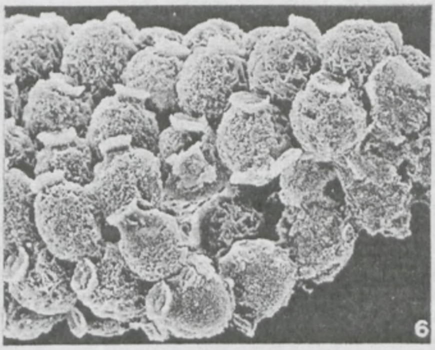
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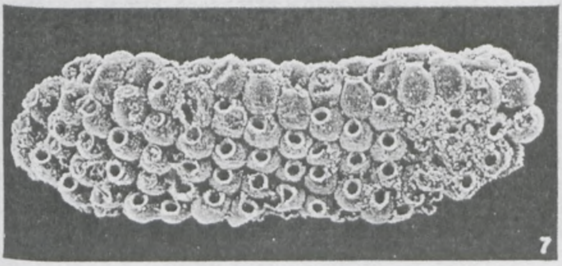
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Appearance Datum (LAD) of each species, when available] the database can be used to evaluate the variations of diversity of the chitinozoans. However, besides these favourable characteristics for carrying out investigations on biodiversification patterns and tempo, the chitinozoan group necessitates some explanations as its has been long regarded as a group of unknown or dubious affinities. Firstly, we must demonstrate that the organic walled-microfossils, known in the literature as chitinozoans, correspond to a stage in the development of marine organisms. Secondly, one must ensure that what is commonly labelled as a species by experts effectively represents a true species in the biological sense, i.e. it must be based on the principle of inter-fecundity. These aspects are documented below.

BIOLOGICAL SIGNIFICATION OF THE CHITINOZOANS

HISTORICAL BACKGROUND

When Eisenack (1931) in his pioneer work introduced the term "chitinozoan" and described the first species, he enumerated possible biological affinities for these organic-walled microfossils. However, despite the animal origin he suggested by the etymology of "Chitinozoa", he retained no conclusive biological interpretation and no systematic assignment. Up to the seventies, with the exception of a brilliant and stimulating paper of Kozlowski (1963), only few new ideas were proposed on the affinities of the chitinozoans (see references in Miller 1996). Subsequent investigations, based on Scanning Electron Microscope (SEM) observations, and a larger chitinozoan record, led succeeding generations of chitinozoan workers to advocate several alternative hypotheses, such as an affinity to the protists, the fungi, the metazoans or that chitinozoans were eggs of marine animals (for the most significant references, see Miller 1996). Whereas several authors listed and/or meticulously discussed the various contentious hypotheses (e.g. Taugourdeau 1981; Sutherland 1994; Miller 1996), others clearly expressed contrasting points of view on the systematic affinities of the chitinozoans (e.g. Jenkins 1970; Laufeld 1974; Grahn & Afzelius 1980; Paris 1981; Grahn 1981; Locquin 1981; Cashman 1990; Jaglin & Paris 1992; Geng et al. 1997; Gabbott et al. 1998). Locquin (1981), for instance, concluded that chitinozoans were fungi, and changed their name to "Chitinomycetes". Cashman (1990), on the other hand, claimed that chitinozoans had affinities with the Rhyzopoda (Foraminifera). Because both the latter hypotheses were based on non-reproducible observations and/or on erroneous descriptions of basic morphological chitinozoan features, they are not considered herein. Much more significant is the hypothesis of reproductive stages originally documented by Kozlowski (1963) on the basis of exceptionally well preserved clusters ("cocoons") of chitinozoans (grouped vesicles with or without a geometric arrangement). Whilst he admitted a possible relationship with the annelids, Kozlowski (1963: 443), was less conclusive on the biological meaning (function) of the chitinozoan vesicles and stressed that they displayed only a remote analogy with eggs or egg capsules of present-day animals. Nevertheless, this "egg theory" has been favoured by numerous authors. Laufeld (1974), for instance, argued in favour of eggs of polychaetes and gastropods and thus for a polyphyletic origin of the chitinozoans. One of us (F. P.) discussed and tested different hypotheses dealing with possible biological interpretation and the affinities of the group (Paris 1981: 78-84) and showed that no known fossil groups can match the range of the chitinozoan group (Paris 1981: 83). The main conclusion, translated here, was: "...chitinozoans are eggs of a soft-bodied marine metazoans a few millimetres in length, eventually with a pelagic or nectic mode of life" (ibid.: 84). The analysis of teratologic individuals supported this conclusion and later led Jaglin & Paris (1992: 163) to state: "...we are still convinced that chitinozoans represent eggs laid by marine metazoans. These eggs developed and reached maturity within the parent organism, and were liberated within the marine environment with a definite size and morphology". Grahn (1981: 30), taking in consideration that "...chitinozoans are the reproductive parts of yet unidentified invertebrates" introduced the name "chitinozoophorans" for these animals. More recently Geng et al. (1997: 42) advocated the egg-laying theory, as did Gabbott et al. (1998) who provided new and impressive evidence of in situ clusters and aggregate masses of chitinozoans. Gabbott et al. (1998) gave a brilliant discussion on the possible "parent" animals. However, we consider premature the conclusion of Gabbott et al. (1998: 451): "Chitinozoans could be the eggs of orthocone cephalopods, although the possibility of a conodont connection is equally tenable", as there are no obvious relationships between the occurrence and the range of the chitinozoans and that of the cephalopods and conodonts in other areas (see discussion of Paris 1981: 83).

NEW DATA

From the preceding review, it is evident that many chitinozoan experts believe that chitinozoans represent the eggs or egg capsules of (a) marine metazoan group(s). However, chitinozoans occur in three different ways: 1) as free vesicles usually dispersed in palynological preparations, 2) less frequently as catenary structures ("chains") recorded in palynological residues or on the surface of rock slabs (e.g. Paris & Mergl 1984, pl. 2, fig. 1; Gabbott et al. 1998, figs 4-5), and 3) as "cocoons", illustrated by Kozlowski (1963) and recalling eggs-capsules. The actual link between these three different kinds of chitinozoan occurrences has not yet been fully explained.

New material collected by one of us (J. N.) from the Ordovician of the Baltic area provides the missing elements for a plausible explanation of the relationship between the various types of occurrences. This "Rosette stone", which in our opinion resolves in part the enigma of the chitinozoans, is represented by a coiled catenary structure (Fig. 2.3-4). In this structure, each vesicle (excepted the last one) is linked to the adjacent one by the attachment of the collarette and the centre of the operculum to the bottom of the nearest vesicle (Fig. 2.2). This connection corresponds to the linkage by double adherence sensu Paris (1981: 75). In this coiled structure, each ring is in full contact with the adjacent ones and the resulting helical spring-like structure appears to be coated by some organic remains, possibly corresponding to residues of mucus or rotten organic matter, originally adhering to the chain. In addition to this coiled catenary structure, free vesicles and chains of several vesicles (Fig. 2.1) and clusters (Fig. 2.5-8) belonging to one species, i.e. *Desmochitina nodosa* EISENACK, have been extracted from the rock samples (Nõlvak 1993).

In the eggs of various present-day organisms, such as gastropods and insects, both those with or without an operculum, the embryo on reaching maturity must leave from the vesicle for further development. If chitinozoans are the eggs of a metazoan group, then the arrangement of vesicles into catenary structures, meant that the larvae or the juveniles were not able to escape from the chamber because of the presence of a tightly sealed operculum (Fig. 2.2). Consequently, these catenary structures cannot represent a final laying of eggs but an immature stage in the development of the eggs, i.e. an intra-oviduct stage. We regard the occurrence of partially fragmented chains within the sediment as resulting from an early liberation of immature eggs. This was probably due to the early death of females and to the subsequent decay of their soft tissue. Depending on the hydrodynamic environment, the chains were more or less disaggregated, but included a few to several tens of linked vesicles.

Depending on the taxa, the normal situation, once the eggs reached maturity, was a laying of the eggs either as clusters around some support, in the form of egg capsules, or as single vesicles in the sea water. The latter was obviously the most common mode of laying for the Lagenochitinidae and the Conochitinidae. Nevertheless, some peculiar helicoidal linkage keeping the opening free (e.g. *Lagenochitina navicula* in Paris 1981, pl. 31, figs. 1-3; or *Lagenochitina* sp. nov. aff. *navicula* in Grahn & Paris 1992, pl. 1, fig. 11) might also have represented a functional laying. In the case of the clusters, we have observed (Fig. 2.5-8) that the opening of each vesicle is free, and directed outwards from the general surface of the aggregate (like amphorae on the bottom of a boat). To achieve such a regular pattern, the "chitinozoan-animal" responsible for the laying most likely possessed an ovipositor, and the egg-clusters were made outside the animal. This interpretation differs basically from the interesting scheme proposed by Dzik (1992, fig. 9-8a), who figured a cluster of "opening-free vesicles" within the uterus.

Because of the size of the lay-out, we postulate that the "chitinozoan animals" were metazoans ranging from a few millimetres to a few centimetres in length. Since no known shelly fossil group fits perfectly with the time-range, the environmental control, and the paleogeographic distribution of the chitinozoans, we favour a group of completely soft-bodied organisms for the "chitinozoan animals". The lack of fossilized soft part may be explained by it rapid decay, as demonstrated experimentally by Briggs & Kear (1993) for the polychaetes. However, a patient and meticulous examination of the bedding surfaces of very fine sediments deposited in anoxic and extremely quite environments (e.g. "Konservat-Lagerstätten", or black shales) should reveal at least a carbonaceous film or an impression of the body outline containing a coiled chain of chitinozoans. The best candidates for such discoveries are sediments containing long catenary structures of species usually not encountered as chains in palynological residues (e.g. Paris & Mergl 1984, pl. 2, fig. 1; Gabbott et al. 1998, figs 4-5). This potential has been little explored and, therefore, we are confident of eventual positive findings, similar to that of the conodont animals (see Aldridge & Purnel 1996). Most of the chitinozoan vesicles suggest a pelagic mode of dissemination, but in some cases the "chitinozoan animals" might lay out within empty shells of dead animals on the bottom of the sea. The "chitinozoan animals" themselves were not necessarily all pelagic organisms, some could have been nectic elements, the wide paleogeographic distribution of the group being the result of the drifting of the eggs.

APPLICABILITY FOR BIODIVERSITY EVALUATION

If chitinozoans are regarded as eggs, they belong to the ontogenic cycle of organisms and therefore their morphology is genetically controlled. Consequently, we consider that each morphological unit in the chitinozoan group (i.e. the "species" as currently accepted by the experts of this group) is representative of the "chitinozoan animal" species that produced these tiny vesicles. In other words, the observed specific diversity of the chitinozoans mirrors the paleobiodiversity of their cryptic parents. This assertion is easy to accept as it can be verified on present-day organisms: the mature eggs of each egg laying metazoan species are morphologically discriminant at specific level if the investigation is carried out with the required accuracy. This is well illustrated by the eggs of different species of insects genera belonging for instance to the Hemiptera (e.g. *Cimex, Picromerus*), Phasmidoptera (e.g. *Cyphocrania, Carausius*), or Lepidoptera (e.g. *Pieris*). It follows that in the field of paleontology, the diversity of such eggs through time can be used to test the fluctuations of the biodiversity of cryptic soft bodied organisms producing resistant eggs, on a world-wide scale or a given paleoplate. We consider that this principle can be extended to microfossils such as the chitinozoans, as well as to other cryptic Paleozoic groups such as the "conodont animals", the polychaetes, the biodiversity of which is better appraised through their conodont or scolecodont apparatus, rather than through their rarely preserved fossil bodies. Finally, we are convinced that the systematic affinities of a fossil group is not a prerequisite for the evaluation of its biodiversity through time. For this reason, we are using the chitinozoans to access the biodiversification pattern and tempo of the "chitinozoan animals" themselves, from their appearance in the late Tremadoc to their extinction at the end of the Famennian.

CHITINOZOAN BIODIVERSITY

The biodiversity of the chitinozoan group can be documented either globally (i.e. for the whole range of the group and on a worldwide scale), or for a more restricted time interval (e.g. V. Nestor in Kaljo et al. 1995a), for some climatic belts, or for peculiar environments (shore face, offshore, slope...) or even within a single section (e.g. J. Nõlvak in Kaljo et al. 1995b).

The aim of the present paper is to point out the abrupt changes observed in the composition of chitinozoan assemblages through time and worldwide. However, our ultimate goal is to depict the respective role of abiotic factors, regardless whether of terrestrial or extra-terrestrial origin, and of the biotic dynamics on the biodiversity of the group.

To document the global diversity of the chitinozoans from the late Tremadoc to the end of the Devonian, we have used the CHITINOVOSP database (Paris & Bernard 1994) which was initially designed for recording all chitinozoans taxa described to date (end of 1997). It includes 1078 species listed on the NET. The available data are more or less equally distributed in Laurentia, Baltica and Gondwana and to a lesser extent from sites in China (Fig. 1). In order to work out the most reliable data, we have selected only stratigraphically and taxonomically well defined species (junior synonyms and dubious forms have been omitted). We have slightly modified the original database in order to introduce additional chronostratigraphic information for each selected species, i.e. the FAD (First Appearance Datum), the LAD (Last Appearance Datum) and all the stages corresponding to its total range. For these chronostratigraphic data, we have used the global standard stages and series for the Silurian and the Devonian, and the British regional series and stages (Fortey et al. 1995) for the Ordovician since the new global Ordovician chronostratigraphy is not yet available. The radiometric time scale of Odin & Odin (1990) is used for the numeric calibration of the series and stages. Because the calibration of the boundaries of each stage is poorly constrained, we have preferred to subdivide equally the duration of the series by the number of stages they contain. The resulting numeric time slicing is not precise, but at least, it does not suggest an accuracy that we cannot attain at present. However, in these schematic diagrams (Figs 3-5), the ratio of the duration of the Silurian series with regard to the Ordovician and Devonian series has been preserved.

Additional paleogeographic and bathymetric data have been registered in the database, but they have not been used in this preliminary work, where some 700 species were finally taken into account.

BIODIVERSITY AT THE SCALE OF THE SERIES

Because precise chronostratigraphic data are not available for each selected species, as a first approach we assigned them to one of the Ordovician, Silurian or Devonian series, or exceptionally to more than one. This seems to be acceptable as the total range of almost all chitinozoan species is much shorter than the duration of a series (see Paris 1996, text-figs 2, 6, 8). However, the 866 "answers" sorted from the database by the program used in this study do not correspond exactly to the actual total number of the species used. Indeed, those ranging beyond a series boundary, even if very short ranging species, are counted twice.

The resulting curve is roughly bell-shaped, with a deep indentation corresponding to the dramatic drop in diversity recorded during the Late Ordovician. The time slicing method used produced a smooth curve, and only the events of highest magnitude could be detected by this method. It

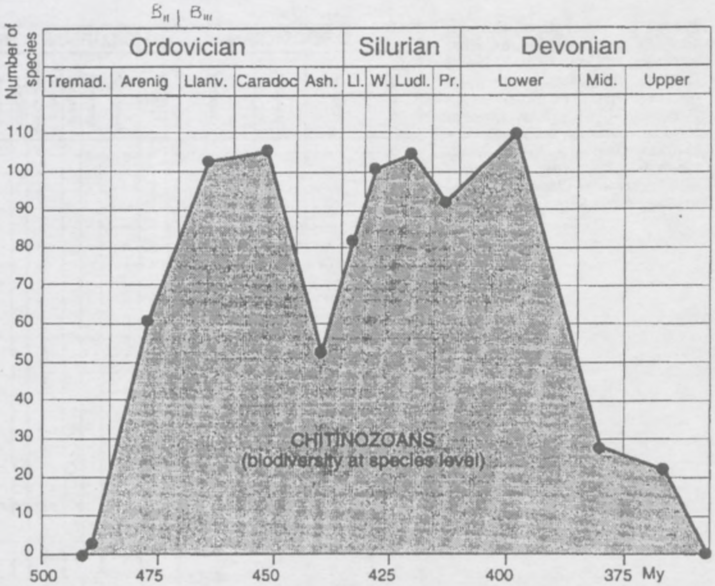

FIGURE 3 - Biodiversity curve of the chitinozoans (i.e. of the "chitinozoans animals") at the series level (radiometric scale from Odin & Odin 1990). *Courbe de diversité des chitinozoaires (c'est-à-dire des "animaux-chitinozoaires") à l'échelle de la série (échelle radiométrique d'après Odin & Odin 1990).*

must be noted that a regular time subdivision (e.g. duration of 20 millions years for each time slice) would produce much higher values for the Silurian because the retained data would be the sum of the Llandovery and Wenlock species and the total of the Ludlow and Pridoli species, respectively.

Despite some incompleteness, mainly due to the time subdivisions used, this curve shows a rapid radiation during the Early and early Middle Ordovician. This radiation is concomitant with the introduction of numerous innovations (Paris et al. 1999), representing more than one-half of all the innovations recorded during the whole range of the group (Fig. 4). Then, a drop of diversity, which began in the Caradoc, ends with a dramatic drop in diversity in the late Ashgill. A rapid recovery is registered during the Silurian before the slow but ultimate Middle and the Late Devonian decline ending with the extinction of the group in the latest Devonian. The Late Ordovician excursion of chitinozoan diversity may be related to the Hirnantian glacial event. However, the power of resolution at series level is not sufficient to document precisely the actual relationship between this drop of diversity and the beginning of the Late Ordovician glaciation. Finally, the fluctuations in biodiversity recorded for the chitinozoans at the series level do not fit with any of the abiotic events listed by Morrow et al. (1995) for the Ordovician, Silurian or Devonian periods.

BIODIVERSITY AT THE SCALE OF THE STAGES

The use of a more refined time subdivision, i.e. the global standard stages, for plotting the occurrences of the selected species, provides a more complex curve of chitinozoan biodiversity (Fig. 5). Among the six recorded peaks of biodiversity, some must be regarded of first magnitude, e.g. the Abereiddian and the Sheinwoodian acmes. Others peaks are of second order, e.g. the Streffordian and the Emsian acmes, whereas the highest peak of diversity registered during the Pridoli must be partly regarded as an artefact because this series is not subdivided into stages.

Following the short recovery noted during the Streffordian, a progressive drop of biodiversity is observed during the four Ashgillian stages. This long term-pattern clearly demonstrated that this phenomenon is not exclusively linked to the Hirnantian glacial event, even though the chitinozoan diversity fell to only 9 species during this stage. Three other severe drops of the chitinozoan diversity are recorded during the Gorstian, the Pragian and the Eifelian, respectively. So far, these lowering of diversity cannot be firmly attributed to any well known worldwide abiotic event. The Pragian excursion is mainly due to an important lowering of the late Pragian chitinozoan diversity as already noted by Winchester-Seeto (1993), who suggested that it may be related to marine regression and the reduction of a suitable environment for their preservation. However, the

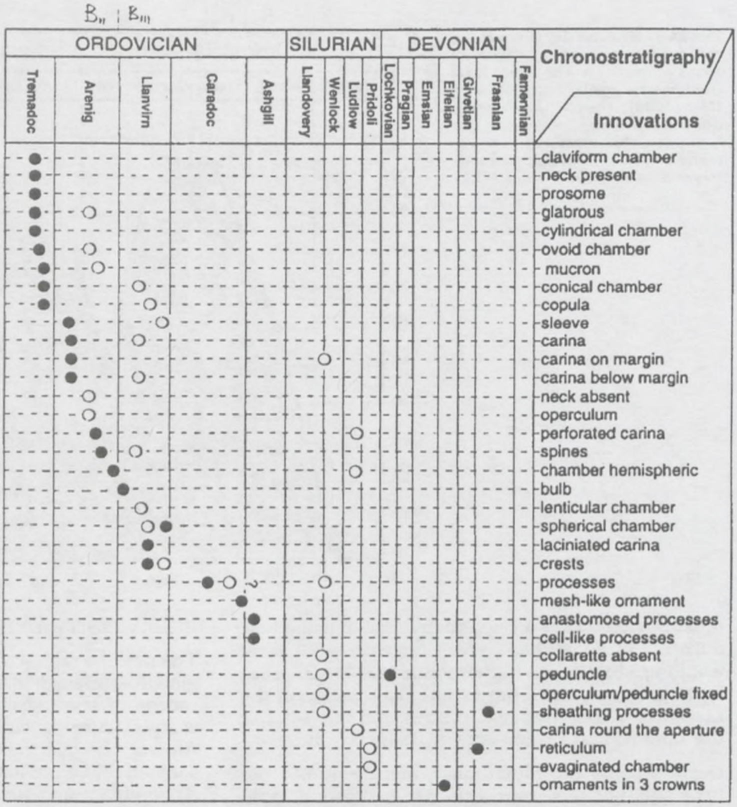

FIGURE 4 - Innovations through time in the chitinozoans from northern Gondwana (solid circles: Prosomatifera; open circles: Operculatifera) (from Paris et al. in press). *Innovations au cours du temps chez les chitinozoaires du Nord Gondwana (ronds pleins: Prosomatifera; cercles: Operculatifera) (d'après Paris et al. sous presse).*

marine regression frequently documented near the Silurian-Devonian boundary in northern Gondwana regions does not coincide with any drop of chitinozoan diversity.

The progressive decline of the group during the Devonian seems to have been only interrupted by two possible brief recoveries during the Emsian and the Givetian. However, it is not clear if these peaks represent artefacts related to the long duration of the Emsian stage, or to sampling problems (e.g. Givetian chitinozoans have been more extensively studied in Laurentia than elsewhere). The progressive loss of specific diversity apparently did not deeply affect the bio-mass production of the "chitinozoan animals", as documented by the 19,000 specimens of chitinozoans per gram of rock recorded in a monospecific assemblage from the base of the Famennian, the highest chitinozoan abundance ever recorded (Paris et al. 1996).

It is likely that by adopting a more refined time subdivision such as the 55 chitinozoan interval biozones identified in the Ordovician, the Silurian and the Devonian, additional signals might be revealed. This will necessitates some deep modifications of the database leading to the elimination of species with poorly constrained biostratigraphic assignments. Moreover, the numeric time calibration of these biozones is less precise than for the stages and their duration is far from constant. A more significant improvement would be to make the evaluations of the chitinozoan diversity at the stage or biozonal levels for each climatic belts, i.e. for the main paleogeographic units.

CONCLUSIONS

Based on the new material extracted from Late Ordovician limestones of Baltica and from unpublished material from the sub-surface of North Africa, we consider that chitinozoans are the eggs of soft-bodied pelagic or nectic animals. We interpret the catenary structures as immature eggs, displaying their original disposition in the oviduct.

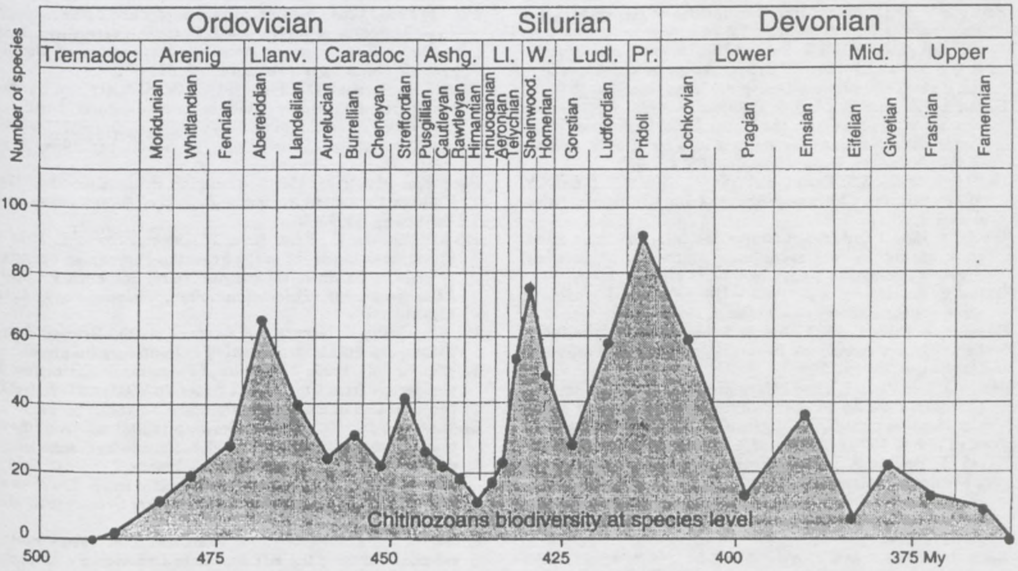

Figure 5 - Biodiversity curve of the chitinozoans (i.e. of the "chitinozoans animals") at the stage level (radiometric scale from Odin & Odin 1990). *Courbe de diversité des chitinozoaires (c'est-à-dire des "animaux-chitinozoaires") à l'échelle de l'étage (échelle radiométrique d'après Odin & Odin 1990).*

These intra-oviduct chain structures were liberated by the decay of the surrounding soft tissues after the death of the "chitinozoan animals". Mature eggs were laid as clusters (cocoons or egg-capsules) or more frequently as single, free vesicles.

The most obvious conclusion concerning the biodiversification of the group is the rapid and important increase in the number of the chitinozoan species, and consequently of the "chitinozoan animal" species, during the early existence of the group, i.e. from the late Tremadoc to the early Llanvirn. This corresponds to a dramatic paleogeographic spreading, such as a major radiation, and to rapid speciation processes introducing numerous innovations. A second major radiation event occurred during the Llandovery-early Wenlock. This event was even more rapid and more important than the early Ordovician one, and can be regarded as the scarring of the dramatic Ashgillian extinction event which was at least partly related to the Late Ordovician glaciation. Following the Silurian acme, the specific diversity declined progressively from the Lochkovian onwards and ended with the extinction of the group in the uppermost Famennian. It is worth noting that the "chitinozoan animals" survived the Frasnian-Famennian mass extinction, despite the fact that the diversity of the group had been declining for a long time.

The comparison of the obtained curves of chitinozoan biodiversity with the excursions recorded in contemporaneous physical and chemical signals does not support a firm relationship between abiotic constraints and the specific diversity of the "chitinozoan animals".

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Paper II

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Ordovician chitinozoan zones from Baltoscandia

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ABSTRACT

Nõlvak, J. and Grahn, Y, 1993. Ordovician chitinozoan zones from Baltoscandia. Rev. Palaeobot. Palynol., 79: 245–269.

On the basis of reports published over the last 60 years, and unpublished data, a chitinozoan zonation is proposed for the Ordovician sequence of Scandinavia, the East Baltic, northeast Poland and Podolia—commonly referred to as Baltoscandia. From the first chitinozoans in the latest Tremadoc to the latest Ashgill (a period comprising ca. 52 My) 15 zones and 8 subzones are defined, each corresponding to some 1–5 My. These are compared with the graptolite and conodont zonations. The greatest precision is achieved in the succession beginning with the conodont Zone of *Amorphognathus tvaerensis*, and ending with the top of the Ordovician, where chitinozoan stratigraphic evolution surpassed that of other faunal elements. Important associated chitinozoan species in each zone and subzone are discussed.

Introduction

Conodonts, ostracodes, graptolites and to some extent trilobites, cephalopods and brachiopods have traditionally been used with varying success in the correlation of Ordovician beds in Baltoscandia (Scandinavia, the East Baltic, northeast Poland and Podolia). Zones have been proposed for all of these fossil groups, except for the brachiopods. The stratigraphic potential of microfossils other than conodonts is insufficiently known. Although chitinozoans have been studied in Baltoscandia for more than sixty years, until now no zonation has been established for the whole Ordovician sequence. Paris (1992) suggested a chitinozoan zonation for the Ordovician of Baltoscandia (Paris and Robardet, 1990), based on published material. However, the zones were not defined, and his scheme is not generally applicable to Ordovician rocks in Baltoscandia. Eisenack's pioneering work on Chitinozoa was based on faunas isolated from glacial erratics originating from Sweden, Estonia and the submarine outcrop area of the Baltic Sea (Fig. 1), and subsequent works have made this region to one of the best investigated in the world. Recently, chitinozoan zones were established for Ordovician

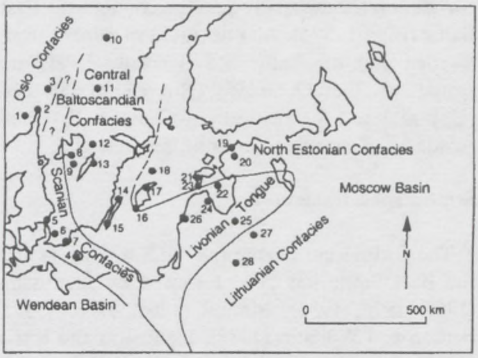

Fig. 1. Map showing approximate boundaries of Ordovician confacies belts in Baltoscandia (slightly modified from Jaanusson, 1982) and localities used in the correlation (Figs. 3–5).

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beds of Québec and western Newfoundland (Achab, 1989) and the northern part of the Gondwana continental plate (Paris, 1990). The Ordovician chitinozoan faunas from these areas are somewhat different from those in Baltoscandia, and most of the zones suggested are not directly applicable to contemporary sequences in Baltoscandia. Some characteristic species appear to have a somewhat different range in Laurentia and North Gondwana than in Baltoscandia. This is in itself not unusual for planktonic organisms. Graptolites, for instance, may have a similar pattern of occurrence. A difficulty in the correlation of Ordovician beds within Baltoscandia is the differentiation of the sequence into at least five composite belts, the Oslo, Scanian, Central Baltoscandian, North Estonian and Lithuanian Confacies belts (sensu Jaanusson, 1976). These are characterized by a single or, more commonly, several contemporaneous lithofacies, but even in cases when differences in lithology between adjacent belts are difficult to define, the benthic faunas remain different (Jaanusson and Bergström, 1980). The main difference in the composition of the fauna of these confacies belts is the dominance of graptolites in the Scanian Confacies belt, in contrast to the four other confacies belts where shelly fossils dominate. It is easy to correlate different areas within each confacies belt, but more problematic between different confacies belts. The aim of this paper is to achieve an Ordovician chitinozoan correlation for especially Scandinavia and the East Baltic (Figs. 1, 3–5). All illustrated specimens from Sweden and the Baltic States (Plates I–VI) are stored at the Geological Survey of Sweden (Uppsala) and the Institute of Geology, Estonian Academy of Sciences (Tallinn), respectively.

Stratigraphic framework

The Ordovician stratigraphy of Scandinavia and the East Baltic has been reviewed by Jaanusson (1963, 1976, 1982), Männil (1966, 1986, 1987), Bruton and Williams (1982), Jaanusson and Karis (1982), Jaanusson and Mutvei (1982), Ulst et al. (1982), Laskov et al. (1984), Kaljo et al. (1988), Kaljo and Nestor (1990), and Owen et al. (1990). In these publications, characteristic macrofossils from Baltoscandia have been discussed, as well as their Ordovician ranges. A tripartite subdivision of the Ordovician sequence, corresponding to a Lower, Middle and Upper Ordovician part, is distinguished in Baltoscandia. The Lower Ordovician (Oelandian) includes strata from the base of the Ordovician (Tremadoc) to the *Didymograptus artus–Didymograptus murchisoni* zonal boundary; the Middle Ordovician (Viruan) ranges from this level to the *Dicranograptus clingani–Pleurograptus linearis* zonal boundary; and the Upper Ordovician (Harjuan) from the base of the *Pleurograptus linearis* graptolite zone to the top of the Ordovician (Jaanusson, 1960). Finds of *Climacograptus spiniferus* (= *C. diplacanthus*) in the Rakvere Stage led Männil (1976) to include this stage in the *Dicranograptus clingani* graptolite zone. However, *C. diplacanthus* is present also in the *P. linearis* Zone, and later investigations place the *clingani*/*linearis* boundary somewhere between upper Oandu and middle Rakvere (cf. Williams and Bruton, 1983). The boundary is probably situated in the lower part of the Rakvere Stage, and corresponds to the base of the Vasagaard Stage in Sweden. It is difficult to define the Middle–Upper Ordovician boundary with fossils in Baltoscandia, but trilobites and graptolites appear to give the best precision at present.

Notes on the Ordovician palaeogeography of Baltoscandia

According to Bergström (1979) and Bruton and Bockelie (1980), Baltoscandia was situated at the South Polar Circle during the Early and Middle Ordovician, while Paris and Robardet (1990) placed Baltoscandia just south of the Tropic of Capricorn. Anyway, Lindström et al. (1991) suggested that an orthoceratitic limestone facies, similar to that in Baltoscandia, occurs when a subtropical or tropical carbonate shelf is drowned, and other evidence of a subtropical to tropical climate, such as "bahamitic" carbonates (Jaanusson, 1973; Stridsberg, 1980), appears for the first time in the Middle Ordovician. The chitinozoan assemblages from North Gondwana differ somewhat from those in Baltoscandia during most of the Ordovician, although they become more similar in the Late Ordovician. More pronounced endemic faunas occurred during Early Ordovician. In the Appalachians (south Laurentia), Middle Ordovician beds show similarities in their chitinozoan faunas with Baltoscandia (Grahn and Bergström, 1984; Hart, 1986), while those from the Avalon Zone become more similar from Caradoc on (cf. Jenkins, 1967). During the Middle and Late Ordovician also a restricted exchange of chitinozoan species between Gondwana and Baltoscandia existed. The chitinozoan genera in Baltica are common to those in Laurentia and Gondwana, but some of the genera, particularly from Gondwana, are absent in Baltoscandia, for instance, *Amphorachitina*, *Ollachitina*, *Plectochitina*, *Siphonochitina* and *Velatachitina*.

Ordovician chitinozoan zones of Baltoscandia

The proposed zones are generally range zones, preferably total range zones. The index species must be common throughout Baltoscandia and have a restricted stratigraphic range. Each zone is defined from the first occurrence of the chitinozoan index species to the first occurrence of the chitinozoan index species in the following zone. Of all the species present in the Ordovician of Baltoscandia, 23 species fulfilled these requirements. We have avoided the use of the genus *Desmochitina* in defining biostratigraphic zonations. This is because of difficulties in distinguishing many of the *Desmochitina* species under the light microscope. The knowledge of phylogenetic lineages for Chitinozoa is still insufficient, and at present the supposed lineages are without any great importance in chitinozoan biostratigraphy. About 100 sections (borings and outcrops) have been investigated. In general 2–3 samples were investigated for each metre, with a sample-weight from ca. 50 g up to about 300–400 g.

Cyathochitina primitiva Zone

Definition: This zone corresponds to the total range of *Cyathochitina primitiva* Szaniawski, 1974 (in Chlebowski and Szaniawski, 1974). This species disappears immediately below the first occurrence of *Conochitina cucumis* Grahn, 1984.

Type interval and locality: The holotype of *Cyathochitina primitiva* was described from the Lower Ordovician conglomerates at Miedzygórz in the Holy Cross Mountains, Poland (Chlebowski and Szaniawski, 1974). As a new type interval and locality for *Cyathochitina primitiva* and the *C. primitiva* Zone we suggest the Leetse and lower Toila formations, Hunneberg to lowermost Volkhov stages (early Arenig), in the Suhkrumägi section (6.3–8.1 m) at Tallinn, North Estonia (Grahn, 1984, fig. 3).

Characteristic chitinozoans: At the type locality *Cyathochitina primitiva* occurs (Plate I, A) together with *Lagenochitina esthonica* Eisenack, 1955, *Eremochitina* sp. (Plate V, C, E) and *Desmochitina* sp. group *minor* (Fig. 6). This is also a typical assemblage in northeast Poland. In Skåne (Scania), south Sweden, *L. esthonica* appears already in the uppermost Tremadoc (Grahn, 1980; see Cooper and Lindholm, 1991), and it is the only species known so far from the Tremadoc beds in Baltoscandia.

Known occurrences outside Baltoscandia: *Cyathochitina primitiva* is known from lower Arenig beds (Obut, pers. commun., 1981; observations by Y.G., 1981) from the Siberian Platform, and the lower Arenigian Hongshiya Formation, Yunnan Province, south China (observations by Y.G., 1983; Gao, 1985).

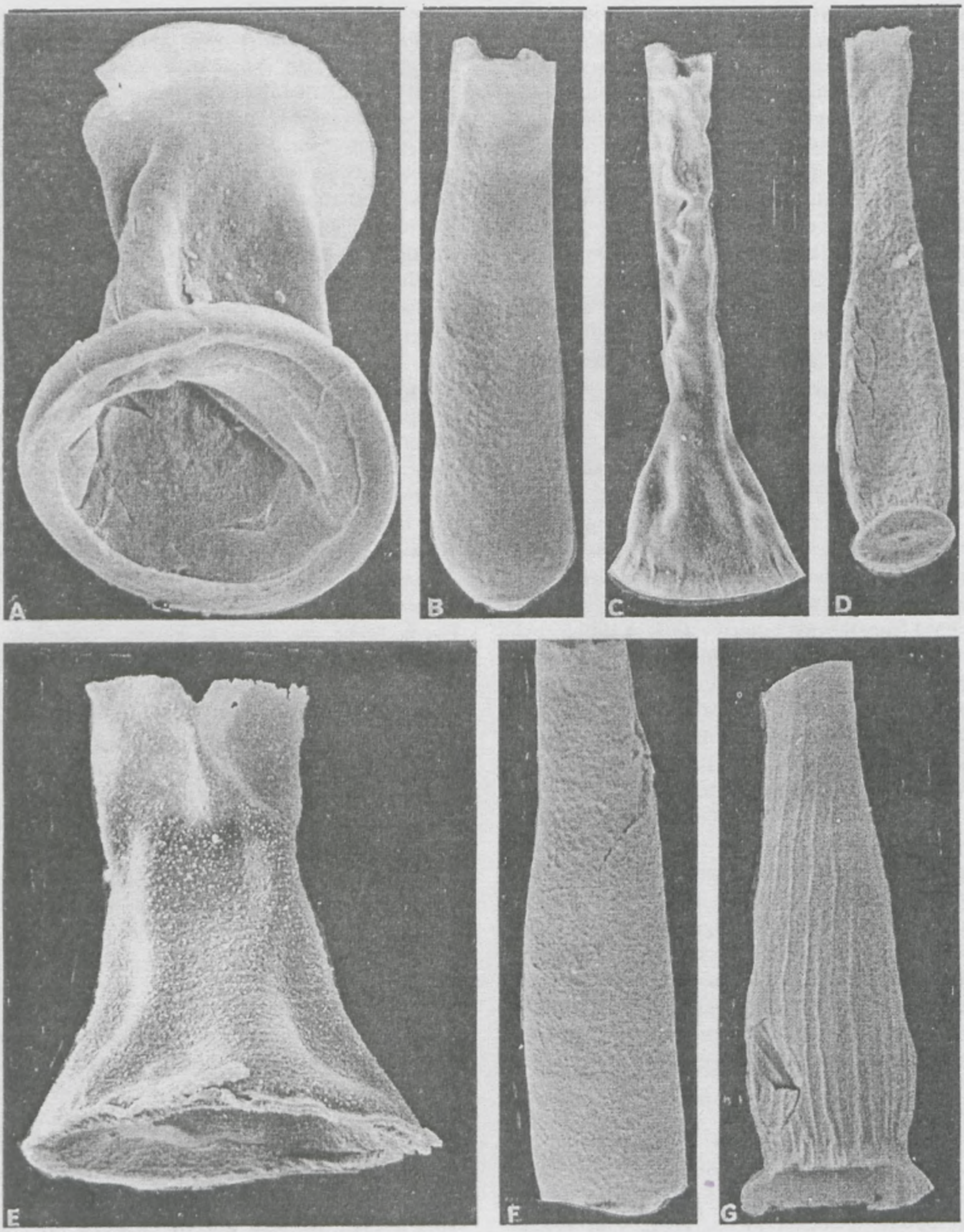
Remarks: The *Cyathochitina primitiva* Zone corresponds to late Hunneberg–late Volkhov time (Fig. 2). SEM-work of the Polish material shows that *Cyathochitina*? *clepsydra* Grahn, 1984 is a junior synonym of *Cyathochitina primitiva*.

Conochitina cucumis Zone

Definition: This zone corresponds to the total range of *Conochitina cucumis* Grahn, 1984, which disappears at the top of Volkhov Stage.

Type interval and locality: The holotype of *C. cucumis* (Plate I, B) is from the upper Toila Formation, Volkhov Stage (late Arenig), in the Suhkrumägi section at Tallinn, North Estonia. This is also the type interval (4.4–6.2 m) and

PLATE I



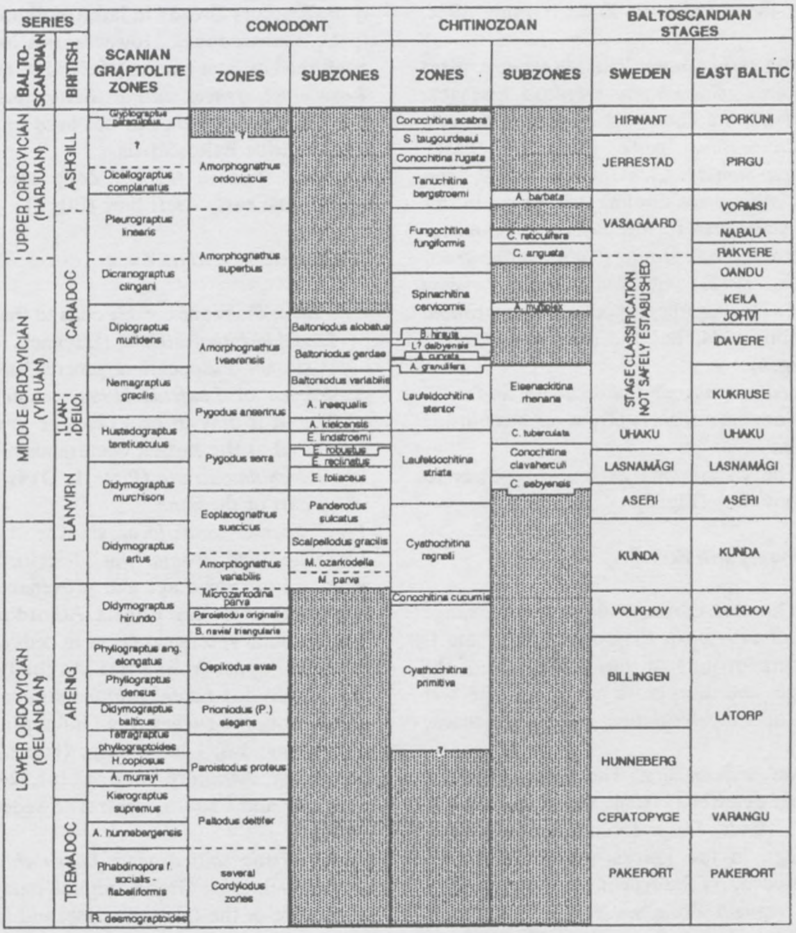


Fig. 2. Correlation table of the Ordovician chitinozoan zones of Baltoscandia. British Series after Fortey et al. (1991). Scanian graptolite zones partly revised by Cooper and Lindholm (1991) and Nilsson (1979).

PLATE I

A. *Cyathochitina primitiva* Szaniawski, 1974. Suhkrumägi, Leetse Formation, Hunneberg Stage, SEM, × 520.
B. *Conochitina cucumis* Grahn, 1984. Suhkrumägi, Toila Formation, Volkhov Stage, SEM, × 340.
C. *Cyathochitina sebyensis* Grahn, 1981. Gammalsby boring 1, Seby Limestone (8.85 m), Lasnamägi Stage, SEM, × 280.
D. *Laufeldochitina striata* (Eisenack, 1937). Persnäs 1, Persnäs Limestone, Uhaku Stage, SEM, × 90.
E. *Cyathochitina regnelli* Eisenack, 1955. Långalvaret 1, Holen Limestone, Kunda Stage, SEM, × 480.
F. *Conochitina clavaherculi* Eisenack, 1959. Böda Hamn boring 1, Källa Limestone (11.30–11.33 m), Uhaku Stage, SEM, × 150.
G. *Laufeldochitina stentor* (Eisenack, 1937). Böda Hamn boring 1, Dalby Limestone (2.18–2.23 m), Kukruse Stage, SEM, × 180.

locality for the *C. cucumis* Zone (Grahn, 1984, fig. 3).

Characteristic chitinozoans: The lowest specimens of, for instance, *Conochitina primitiva* Eisenack, 1939, *Cyathochitina campanulaeformis* (Eisenack, 1931), *Cyathochitina calix* (Eisenack, 1931), *Lagenochitina tumida* s.l. Umnova, 1969, and *Tanuchitina tallinnensis* Grahn, 1984 occur in this zone. This is also true for the *Belonechitina micracantha–wesenbergensis* group, *Rhabdochitina gracilis* Eisenack, 1962, and *Rhabdochitina magna* Eisenack, 1931. Another species, *Desmochitina papilla* Grahn, 1984, has the same range as *C. cucumis* (Fig. 6).

Known occurrences outside Baltoscandia: So far, *C. cucumis* has not been reported from regions outside Baltoscandia.

Remarks: The *C. cucumis* Zone corresponds to latest Volkhov time (Fig. 2).

Cyathochitina regnelli Zone

Definition: This zone corresponds to the total range of *Cyathochitina regnelli* Eisenack, 1955 (Plate I, E), and occurs for the first time at the base of the Kunda Stage, and disappears just before the first occurrence of *Laufeldochitina striata* (Eisenack, 1937).

Type interval and locality: The holotype of *C. regnelli* was described from the lower Holen Limestone, Kunda Stage (Hunderum Substage; latest Arenig), in the Fjäcka section at Dalby, Dalarna, Sweden. As the type interval and locality for the *C. regnelli* Zone we suggest the Pakri, Loobu and lowermost Aseri formations, Kunda and lowermost Aseri stages (latest Arenig–early Llanvirn), in the Suhkrumägi section (3.5–4.2 m) at Tallinn, North Estonia (Grahn, 1984, fig. 3).

Characteristic chitinozoans: In the lower part of the *C. regnelli* Zone *Lagenochitina tumida* s.l. disappears at the Arenig–Llanvirn boundary, but reappears in the upper Llanvirn (Männil, 1986). In the uppermost Kunda Stage *Belonechitina crinita* (Grahn, 1984) and *Linochitina* sp. (Plate VI, C) appear. The latter species disappears, together with *Lagenochitina esthonica*, at the *Didymograptus artus–Didymograptus murchisoni* zonal boundary (Figs. 2, 6). According to Eisenack (1976) *C. regnelli* occurs already in latest Volkhov on Öland (= *C. cucumis* zone). However, this has not been confirmed by later studies (cf. Grahn, 1980, 1982b).

Known occurrences outside Baltoscandia: So far, *Cyathochitina regnelli* has not been reported from areas outside Baltoscandia.

Remarks: The *C. regnelli* Zone corresponds to Kunda and early Aseri time (Fig. 2).

Laufeldochitina striata Zone

Definition: This zone corresponds to the total range of *Laufeldochitina striata* (Eisenack, 1937). The index species disappears in general below the first occurrence of *Laufeldochitina stentor* (Eisenack, 1937), but it may in rare cases be present at the same level as the earliest occurrence of *L. stentor*. *Laufeldochitina striata* (Plate I, D) is rare in the lower part of the zone.

Type interval and locality: The holotype of *Laufeldochitina striata* was described from an erratic of unknown age and provenance found at Samland, former East Prussia. According to Grahn (1984) similar specimens occur in beds corresponding to the upper Aseri Stage. As the type interval and locality for *Laufeldochitina striata* and the *L. striata* Zone we suggest the Gullhögen and lower Ryd formations, Uhaku Stage (late Llanvirn), in the Stora Åsbotorp boring (61.72–74.12 m), Västergötland, south central Sweden (Grahn, 1981b).

Characteristic chitinozoans: *Lagenochitina tumida* s.l. (Männil, 1986; Grahn, unpubl. data) reappears at the base of the *L. striata* zone, and in the upper part appear *Cyathochitina kuckersiana* (Eisenack, 1934). At present two chitinozoan subzones are defined within the *L. striata* Zone. The associated chitinozoans in this part of the zone will be discussed under the subzones of *Cyathochitina sebyensis* and *Conochitina clavaherculi*, respectively (Figs. 2, 6).

Known occurrences outside Baltoscandia: So far, *Laufeldochitina striata* has not been reported from areas outside Baltoscandia.

Remarks: The *L. striata* Zone corresponds to late Aseri–early Uhaku time (Fig. 2). According to Paris and Mergl (1984) *Lagenochitina tumida* s.l. and *Desmochitina bulla* Taugourdeau and

Jekhowsky, 1960 are conspecific. *Laufeldochitina striata* is also known from Poland (R. Wrona, pers. commun., 1991).

Cyathochitina sebyensis Subzone

Definition: This subzone corresponds to the total range of *Cyathochitina sebyensis* Grahn, 1981a. It is more common in the upper part of the subzone, and disappears just below the appearance of *Conochitina clavaherculi* Eisenack, 1959.
Type interval and locality: The holotype of *Cyathochitina sebyensis* (Plate I, C) was described from the Seby Limestone, Lasnamägi Stage (late Llanvirn), in the Gammalsby boring (8.85 m), Öland, south Sweden. As the type interval and locality for the *C. sebyensis* Subzone we suggest units *b* and *c* in the Seby Limestone (Jaanusson, 1960, text-fig. 25) in the Gammalsby boring (8.84–9.03 m).
Characteristic chitinozoans: *Belonechitina pellifera* (Eisenack, 1959) and *Pterochitina retracta* Eisenack, 1955 occur for the first time at the base of this subzone, and *Belonechitina crinita* and *Tanuchitina tallinnensis* (Fig. 6) disappear at the top.
Known occurrences outside Baltoscandia: So far, *Cyathochitina sebyensis* has not been reported outside Baltoscandia.
Remarks: The *C. sebyensis* Subzone corresponds to late Aseri–early Lasnamägi time (Fig. 2).

Conochitina clavaherculi Subzone

Definition: This subzone corresponds to the total range of *Conochitina clavaherculi* Eisenack, 1959.
Type interval and locality: The holotype of *Conochitina clavaherculi* was described from the Källa Limestone at Källa, lowermost Uhaku Stage (late Llanvirn), Öland, south Sweden. As the type interval and locality for the *C. clavaherculi* Subzone we suggest the Väo Formation, Lasnamägi and lower Uhaku stages in the Lasnamägi section (1.4–5.7 m) at Tallinn, North Estonia (Grahn, 1984, fig. 4).
Characteristic chitinozoans: In the lowermost part of the subzone *Sagenachitina* sp. (Plate V, A, B) appears for the first time, and higher up *Belonechitina pellifera* and *Lagenochitina tumida* s.l. disappear. The first specimens of the *Belonechitina cactacea*-group appear in the upper part of the Lasnamägi Stage (Fig. 6). The top of the *C. clavaherculi* Zone coincides with the top of the *Eoplacognathus robustus* conodont zone.
Known occurrences outside Baltoscandia: *Conochitina clavaherculi* (Plate I, F) is reported from the Lenoir Limestone of Alabama and Tennessee, USA (Grahn and Bergström, 1984).
Remarks: The *C. clavaherculi* Subzone corresponds to most of Lasnamägi and early Uhaku time (Fig. 2), but its density varies considerably. It is rare in the middle part of its range including the Lasnamägi–Uhaku transition. *Conochitina clavaherculi* is also known from Poland (R. Wrona, pers. commun., 1991).

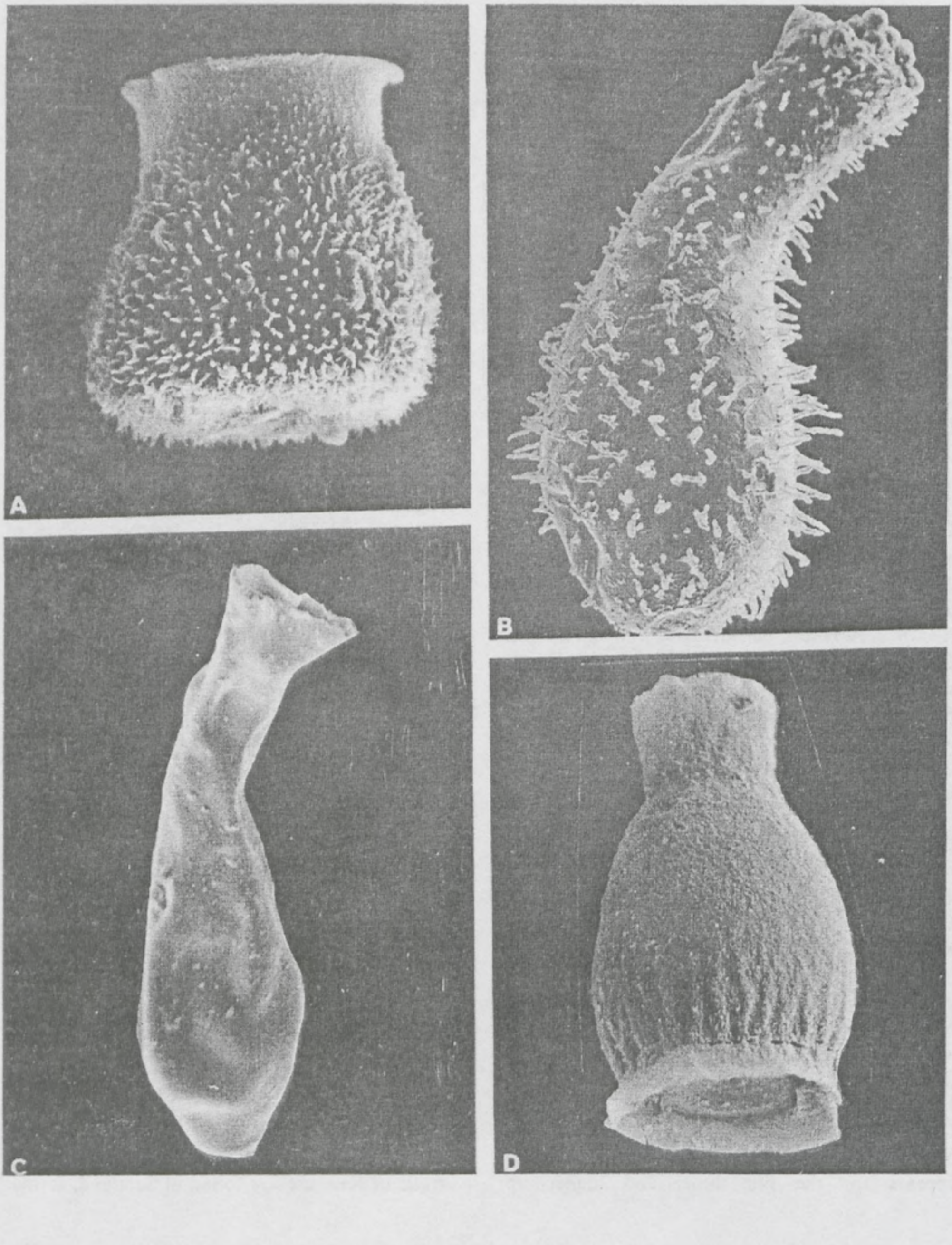
Conochitina tuberculata Subzone

Definition: This subzone corresponds to the partial range of *Conochitina tuberculata* Eisenack, 1962 (Plate V, D), between the last occurrence of *Conochitina clavaherculi* and the first occurrence of *Eisenackitina rhenana* (Eisenack, 1939).
Type interval and locality: The holotype of *Conochitina tuberculata* was described from Kukruse equivalents (Llandeilo–early Caradoc) in the allochtone at Halde bei Metzingen, Württemberg, Germany (Eisenack, 1962). This is also the type interval and locality for the *C. tuberculata* Subzone.
Characteristic chitinozoans: *Laufeldochitina striata* is a characteristic species in the lower part of this subzone.
Known occurrences outside Baltoscandia: Eisenack (1962) described *C. tuberculata* from Kukruse equivalents in Württemberg, Germany.
Remarks: The *C. tuberculata* Subzone corresponds to most of Uhaku time (Fig. 6).

Laufeldochitina stentor Zone

Definition: This zone corresponds to the total range of *Laufeldochitina stentor* (Eisenack, 1937).
Type interval and locality: The holotype of *Laufeldochitina stentor* was described from an erratic of Kukruse age found at Samland, former

PLATE II



East Prussia. As the type interval and locality for *Laufeldochitina stentor* and the *L. stentor* Zone we suggest the upper Ryd and lower Dalby formations, upper Uhaku and Kukruse stages (Llandeilo–early Caradoc), in the Stora Åsbotorp boring (48.52–54.35 m), Västergötland, south central Sweden (Grahn, 1981b).

Characteristic chitinozoans: In the lower part of this zone *Pterochitina retracta* and *Conochitina tuberculata* are characteristic species, and in the uppermost part *Conochitina tigrina* Laufeld, 1967 and, more rarely, *Spinachitina* sp. *Eisenackitina rhenana* (Eisenack, 1939) characterize a subzone within the *L. stentor* Zone (Figs. 2, 6). *Laufeldochitina stentor* (Plate I, G) is rare in the middle part of the zone. However, this middle interval is characterized by *Laufeldochitina* cf. *striata* (see Männil, 1986, fig. 2.1.1) often together with *Desmochitina ovulum* Eisenack, 1962.

Known occurrences outside Baltoscandia: Paris (1990) reported *L. stentor* from the upper part of the Traveusot Formation, Brittany, France. Eisenack (1939) described it from the Herscheider Shale of Westphalia.

Remarks: The *L. stentor* Zone corresponds to late Uhaku and Kukruse time (Fig. 2). Podhalanska (1979) reported *Laufeldochitina stentor* from Poland.

Eisenackitina rhenana Subzone

Definition: This subzone corresponds to the total range of *Eisenackitina rhenana* (Eisenack, 1939). However, it has a discontinous occurrence through the Kukruse Stage.

Type interval and locality: The holotype of *Eisenackitina rhenana* was described from the Rheinischen Schiefergebirges in Germany. The type interval and locality is unknown. As a new type interval and locality for *Eisenackitina rhenana* and the *E. rhenana* Subzone we suggest the Viivikonna Formation, Kukruse Stage (late Llandeilo–early Caradoc), in the Rapla boring (158.4–167.3 m), North Estonia.

Characteristic chitinozoans: The subzone corresponds to the Kukruse Stage where *Conochitina tuberculata* and *Pterochitina retracta* disappear in the lower part, and *Conochitina tigrina* in the uppermost part. The last occurrence of *C. tuberculata* is a good indication of the basal *Amorphognathus tvaerensis* conodont zone. In the lower part the first specimens of *Spinachitina suecica* (Laufeld, 1967) and *Calpichitina complanata* (Eisenack, 1932) occur.

Known occurrences outside Baltoscandia: *Eisenackitina rhenana* (Plate II, A; Plate VI, A) has also been reported from Rheinischen Schiefergebirges in Germany, and the lower part of the Louredo Formation in Portugal (Paris, 1981).

Remarks: The *E. rhenana* Subzone corresponds to the Kukruse time (Figs. 2, 6). Eisenack (1955) described seemingly smooth specimens of *E. rhenana* as *Conochitina oelandica*. SEM-work of "smooth" specimens defined as *C. oelandica* shows that these have traces of spines (cf. Plate VI, D). Hence, *C. oelandica* is here considered as a junior synonym of *E. rhenana*.

Armoricochitina granulifera Zone

Definition: This zone corresponds to the total range of *Armoricochitina granulifera* n. sp. (see Appendix 1).

Type interval and locality: The holotype of *Armoricochitina granulifera* (Plate VI, B, D) is from the Tatruse Formation, lower Idavere Stage (early Caradoc), in the Imavere boring (215.7–215.9 m), North Estonia. This is also the type interval and locality for the *A. granulifera* Zone.

PLATE II

A. *Eisenackitina rhenana* (Eisenack, 1939). Viljandi boring, Viivikonna Formation (365 m), Kukruse Stage, SEM, × 880.
B. *Angochitina curvata* n. sp. Udriku boring, Tatruse Formation, Idavere Stage, SEM, × 620.
C. *Lagenochitina*? *dalbyensis* (Laufeld, 1967). Gotska Sandön boring 1, Dalby Limestone (132.30 m), Idavere Stage, SEM, × 440.
D. *Armoricochitina granulifera* n. sp. Stora Åsbotorp boring 1, Dalby Limestone (47.01 m), Idavere Stage, SEM, × 330.

Characteristic chitinozoans: In addition to the index species only long-ranging species are present.
Known occurrences outside Baltoscandia: So far, *Armoricochitina granulifera* (Plate II, D; Plate VI, B, D) has not been reported outside Baltoscandia.
Remarks: The *A. granulifera* Zone corresponds to the earliest Idavere (Fig. 2).

Angochitina curvata Zone

Definition: This zone corresponds to the total range of *Angochitina curvata* n. sp. (Plate II, B; Appendix 1). In the basal part *Armoricochitina granulifera* may occur together with *Angochitina curvata* (Fig. 6). *Angochitina curvata* disappears immediately below the first occurrences of *Lagenochitina*? *dalbyensis* (Laufeld, 1967).
Type interval and locality: The holotype of *Angochitina curvata* was described from the Tatruse Formation, Idavere Stage (early Caradoc), in the Udriku boring (63.70–63.75 m), North Estonia. As the type interval and locality for the *A. curvata* Zone we suggest the lower Tatruse Formation, basal Idavere Stage in the Äiamaa boring (201.0–201.1 m), North Estonia.
Characteristic chitinozoans: Besides rare occurrences of the index species the zone yields only long-ranging species.
Known occurrences outside Baltoscandia: So far, *Angochitina curvata* has not been reported outside Baltoscandia.
Remarks: The *A. curvata* Zone corresponds to early Idavere time (Fig. 2).

Lagenochitina? *dalbyensis* Zone

Definition: This zone corresponds to the total range of *Lagenochitina*? *dalbyensis* (Laufeld, 1967), which is a common species through the zone, and it disappears just before the first occurrences of *Belonechitina hirsuta* (Laufeld, 1967).
Type interval and locality: The holotype of *Lagenochitina*? *dalbyensis* was described from the upper Dalby Limestone, Idavere Stage (early Caradoc), in the Fjäcka section at Dalby, Dalarna, Sweden. This is also the type interval and locality for the *L.*? *dalbyensis* Zone.
Characteristic chitinozoans: In the upper part of this zone *Belonechitina robusta* s.l. (Eisenack, 1959) and *Desmochitina nodosa* Eisenack, 1931 (Fig. 6) appear.
Known occurrences outside Baltoscandia: *Lagenochitina*? *dalbyensis* (Plate II, C) is also described from the lower Ktaoua Formation, near Zagora, south Morocco (Elaouad-Debbaj, 1986). In central Portugal, Paris (1981) reported it from the topmost Carregueira Formation (= topmost Cacemes Formation) and the basal Louredo Formation. In the Armoricain Massif, *L.*? *dalbyensis* occurs in the basal Kermeur Formation (Paris, 1981).
Remarks: The *L.*? *dalbyensis* Zone corresponds to early–middle Idavere time (Fig. 2). The species was originally described as *Eremochitina dalbyensis* by Laufeld, 1967. Probably this zone is equal to the *L. dalbyensis* Zone of Paris (1990).

Belonechitina hirsuta Zone

Definition: This zone corresponds to the partial range of *Belonechitina hirsuta* (Laufeld, 1967), between the last occurrence of *Lagenochitina*? *dalbyensis* and the first occurrence of *Spinachitina cervicornis* (Eisenack, 1931).
Type interval and locality: The holotype of *Belonechitina hirsuta* was described from the upper Dalby Limestone, Idavere Stage (early Caradoc), in the Fjäcka section at Dalby, Dalarna, Sweden. This is also the type interval (29.8–32.5 m) and locality for the *B. hirsuta* Zone (Laufeld, 1967, fig. 6).
Characteristic chitinozoans: Except for *B. hirsuta* (Plate V, F), the zone is poorly characterized, and the species disappears approximately at the top of the *Baltoniodus gerdae* conodont subzone (Figs. 2, 6).
Known occurrences outside Baltoscandia: *Belonechitina hirsuta* has been reported from the upper Caradocian beds in the Cincinnati Region (Miller, 1976; Knabe, 1980) and New York (Hart, 1986), and from the lower Caradocian in Oklahoma, USA (Grahn and Miller, 1986). Achab (1989) described it from Llandeilian to lower Ashgillian strata in Canada, and Jenkins (1967) from the upper Caradocian in Shropshire, England. This extensive range of *B. hirsuta* suggests that more than one species may be involved. In Baltoscandia the species has a very restricted range in the upper Idavere Stage.

Remarks: The *Belonechitina hirsuta* Zone in Baltoscandia corresponds to middle Idavere time (Fig. 2).

Spinachitina cervicornis Zone

Definition: This zone corresponds to the total range of *Spinachitina cervicornis* (Eisenack, 1931) which is a rare species in the lower part of the zone.
Type interval and locality: The holotype was described from an erratic boulder of unknown age and provenance found at Samland, former East Prussia. To judge from the associated chitinozoans it could be of Keila age. As the type interval and locality for *Spinachitina cervicornis* and the Zone of *S. cervicornis* we suggest the Keila and Oandu equivalents (late Caradoc), in the Fjäcka section (19.5–27.1 m) at Dalby, Dalarna, Sweden (Laufeld, 1967, fig. 6).
Characteristic chitinozoans: In the lower part of this zone *Desmochitina nodosa* and *Spinachitina multiradiata* (Eisenack, 1959) are present. The first-mentioned species disappears at the top and *S. multiradiata* in the middle part of the Keila Stage. *Desmochitina juglandiformis* Laufeld, 1967 first occurs in the lower part of the zone. *Belonechitina robusta* s.l. ranges through the zone. *Cyathochitina calix* s.l. disappears at the top of the Jõhvi Stage. In the uppermost Keila there occurs a characteristic, primarily curved form, transitional between *Belonechitina micracantha* and *Conochitina primitiva*. In the Jõhvi and Keila stages there are typical specimens of *Belonechitina comma* (Eisenack, 1959). At the base of the Oandu Stage a new complex of *Conochitina* and *Belonechitina* species appears: one of these is *Belonechitina villosa* (Grahn, 1984) (Fig. 6).
Known occurrences outside Baltoscandia: So far, *Spinachitina cervicornis* (Plate III, A) has not been reported outside Baltoscandia.
Remarks: The *S. cervicornis* Zone corresponds to late Idavere–early Oandu time (Fig. 2).

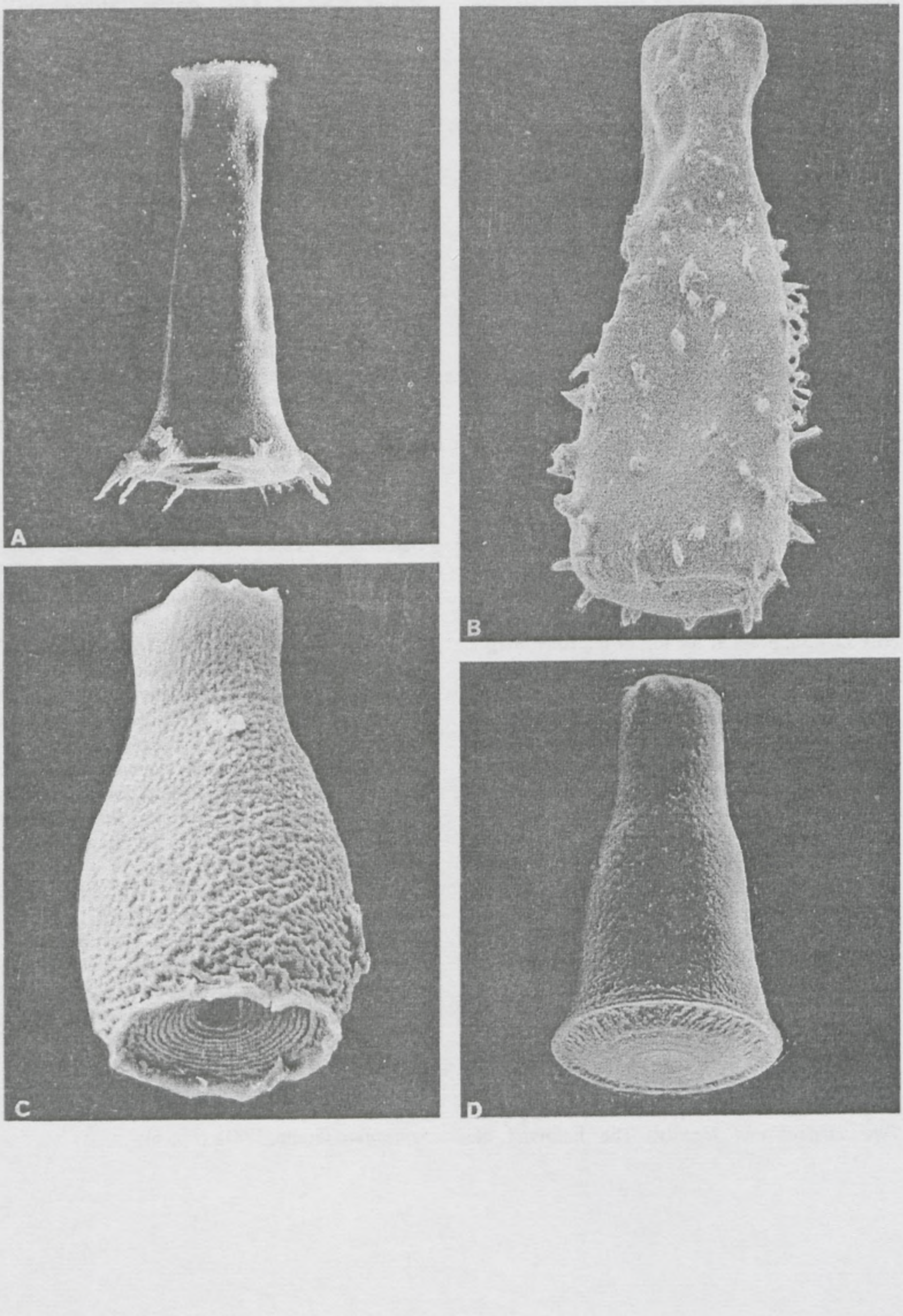
Angochitina multiplex Subzone

Definition: This subzone corresponds to the total range of *Angochitina multiplex* (Schallreuter, 1963).
Type interval and locality: The holotype of *Angochitina multiplex* was described from an erratic dated as Keila age by ostracodes (V. Jaanusson, unpubl. data). It was found at Teschenhagen near Stralsund, Germany. As the type interval and locality for *Angochitina multiplex* and the *A. multiplex* Subzone we suggest the middle Kahula Formation, basal Keila Stage (late Caradoc), in the Rapla boring (145.60–145.70 m), North Estonia.
Characteristic chitinozoans: Only a few long-ranging species are associated with the index species (Fig. 6).
Known occurrences outside Baltoscandia: So far, *Angochitina multiplex* (Plate III, B) has not been reported outside Baltoscandia.
Remarks: *Angochitina multiplex* (= *Illichitina multiplex*) is an excellent index species for a thin succession of beds immediately above the "big bentonite" at the base of the Keila Stage or Skagen Limestone (Figs. 2–6).

Fungochitina fungiformis Zone

Definition: This zone corresponds to the total range of *Fungochitina fungiformis* (Eisenack, 1931). The species is rare in the lower part of the zone (in the Rakvere Stage).
Type interval and locality: The holotype of *Fungochitina fungiformis* was described from an erratic found at Samland, former East Prussia. It is probably from the lower Vormsi Stage. As the type interval and locality for *Fungochitina fungiformis* and the *F. fungiformis* Zone we suggest the Rägavere and lowermost Paekna formations, the Rakvere and lower Vormsi stages (latest Caradoc–early Ashgill), in the Rapla boring (82.0–125.7 m), North Estonia.
Characteristic chitinozoans: At the base *Lagenochitina baltica* Eisenack, 1931 occurs. *Belonechitina capitata* (Eisenack, 1962) disappears in the lower Nabala Stage. At the Nabala–Vormsi boundary the last specimens of the *Belonechitina cactacea*-group occur. In the topmost part of the *F. fungiformis* Zone the first specimens of *Spinachitina coronata* (Eisenack, 1931) are present. The zone comprises the two subzones of *Cyathochitina angusta* n. sp. and *Cyathochitina reticulifera* Grahn, 1981a (Fig. 6).

PLATE III



Known occurrences outside Baltoscandia: So far, *Fungochitina fungiformis* has not been reported outside Baltoscandia.
Remarks: The *F. fungiformis* Zone corresponds to late Oandu–early Vormsi time (Fig. 2). Eisenack (1962) mentioned two subspecies of *Fungochitina fungiformis* from the "Wesenberger Stufe" (= Rakvere Stage). *Fungochitina fungiformis fungiformis* is a smooth subspecies and is more common in the lower Vormsi Stage. The more abundant *Fungochitina fungiformis spinifera* displays small spines at the basal edge.

Cyathochitina angusta Zone

Definition: This zone corresponds to the partial range of *Cyathochitina angusta* n. sp. (Plate III, D; Appendix 1), up to the first occurrence of *Armoricochitina reticulifera*.
Type interval and locality: The holotype is from the lowermost Mõntu Formation in the Eikla boring, North Estonia (see Appendix 1). As the type interval and locality for the *C. angusta* Zone we suggest the Rakvere Stage (latest Caradoc–earliest Ashgill), in the Rapla boring (107.0–123.4 m), North Estonia.
Associated chitinozoans: The lower part of the zone contains *Conochitina incerta* Eisenack, 1962 and *Belonechitina villosa* (Fig. 6). The former species appears in the lower Rakvere Stage, and the latter disappears at this level.
Remarks: The *C. angusta* Zone is equal to the Rakvere time (Fig. 2). Rare specimens may occur in the lowermost Nabala Stage.

Armoricochitina reticulifera Subzone

Definition: This subzone corresponds to the total range of *Armoricochitina reticulifera* (Grahn, 1981).
Type interval and locality: The holotype of *Armoricochitina reticulifera* (Plate III, C) was described from an erratic of late Ordovician age found at Gräsgård, south Öland, Sweden. As the type interval and locality for *Armoricochitina reticulifera* and the *A. reticulifera* Subzone we suggest the Paekna Formation, lower Nabala Stage (late Caradoc), in the Rapla boring (104–107 m), North Estonia.
Characteristic chitinozoans: At the base *Lagenochitina prussica* Eisenack, 1931 appears, and at about the same level *Belonechitina robusta* s.l. and *Belonechitina capitata* disappear (Fig. 6).
Known occurrences outside Baltoscandia: So far, *Armoricochitina reticulifera* has not been reported outside Baltoscandia.
Remarks: The *A. reticulifera* Subzone corresponds to the early Nabala time (Fig. 2). Synonymous to this species is *Cyathochitina dispar* in Estonian literature.

Tanuchitina bergstroemi Zone

Definition: This zone corresponds to the total range of *Tanuchitina bergstroemi* Laufeld, 1967.
Type interval and locality: The holotype of *Tanuchitina bergstroemi* was described from the Fjäcka Shale, upper Vormsi Stage (early Ashgill), at Amtjärn, Dalarna, Sweden. As the type interval and locality for the *T. bergstroemi* Zone we suggest the upper Tudulinna and Moe formations, upper Vormsi and lower Pirgu stages in the Rapla boring (50.0–77.7 m), North Estonia.
Characteristic chitinozoans: *Conochitina incerta*, *Lagenochitina baltica*, *Lagenochitina prussica* and specimens of the *Belonechitina wesenbergensis*-group disappear at the top of the zone. In the uppermost part the first specimens of *Ancyrochitina ancyrea* (Eisenack, 1931) appear. The Subzone of *Acanthochitina barbata* Eisenack, 1931 is defined within the zone (Figs. 2, 6).

PLATE III

A. *Spinachitina cervicornis* (Eisenack, 1931). Rapla boring, Kahula Formation (142.7 m), Keila Stage, SEM, × 440.
B. *Angochitina multiplex* (Schallreuter, 1963). Laeva boring 13, Kahula Formation (237 m), Keila Stage, SEM, × 200.
C. *Armoricochitina reticulifera* (Grahn, 1981a). Gräsgård kanal 1, Erratic boulder, Nabala Stage, SEM, × 365.
D. *Cyathochitina angusta* n. sp. Eikla boring, Mõntu Formation (290.2 m), Nabala Stage, SEM, × 230.

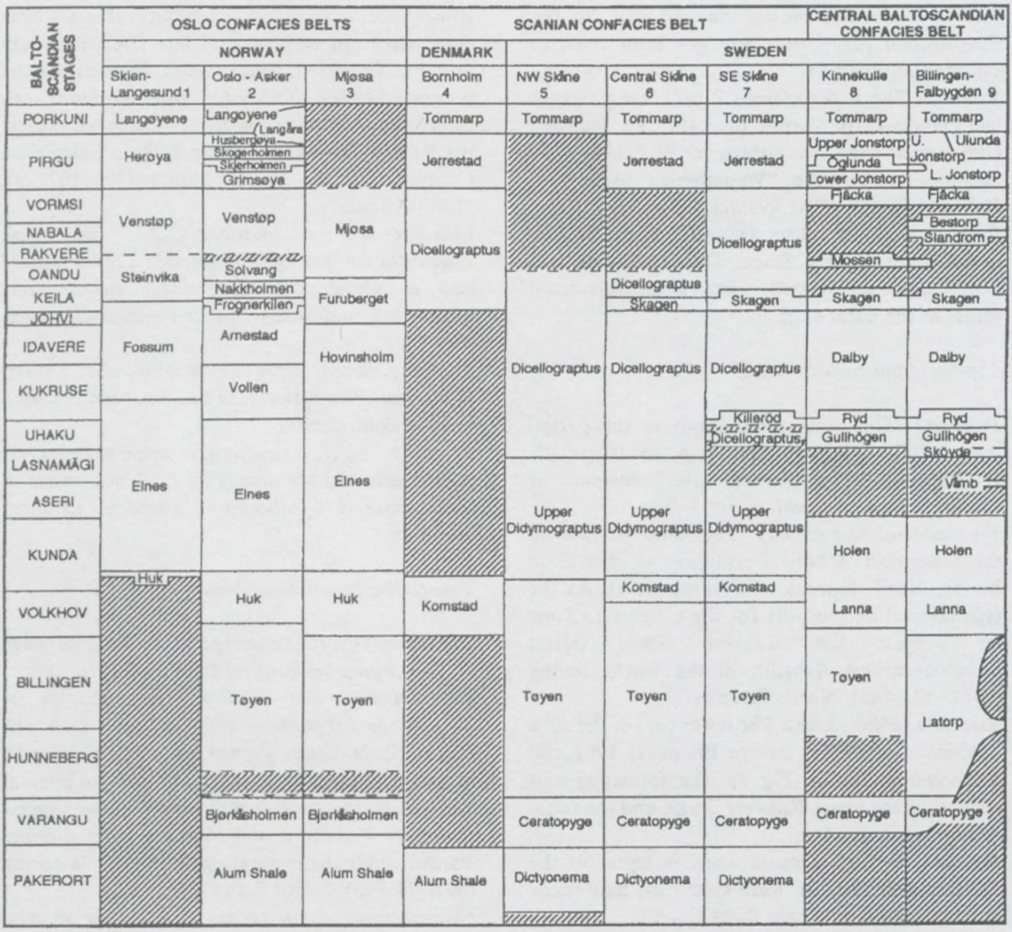


Fig. 3. Correlation table of the Ordovician of the Oslo Region (Owen et al., 1990), Bornholm (Grahn, unpubl. data), Skåne (Bergström, 1982; Grahn, unpubl. data), and Västergötland (Kinnekulle and Billingen–Falbygden; Jaanusson, 1982; Nõlvak and Grahn, unpubl. data). Numbers at the heads of columns refer to Fig. 1.

Known occurrences outside Baltoscandia: According to Paris (1988) *Tanuchitina bergstroemi* (Plate IV, B) occurs in Libya from the Rawtheyan to the Ordovician–Silurian boundary. In Baltoscandia *T. bergstroemi* disappears at about the same level as it should appear in Libya. Later, Paris (1990) considered *Tanuchitina bergstroemi* to be a junior synonym of *Tanuchitina elongata* (Bouché, 1965), and gave the latter species a range corresponding to Rawtheyan and Hirnantian. *Tanuchitina bergstroemi* is also known from the Rawtheyan in Great Britain (Paris, 1990). It is not clear if *T. bergstroemi* and *T. elongata* really are one and the same species. Anyway, *T. bergstroemi* is an excellent index fossil with a restricted range in Baltoscandia.

Remarks: The *T. bergstroemi* Zone corresponds to late Vormsi–early Pirgu time (Fig. 2).

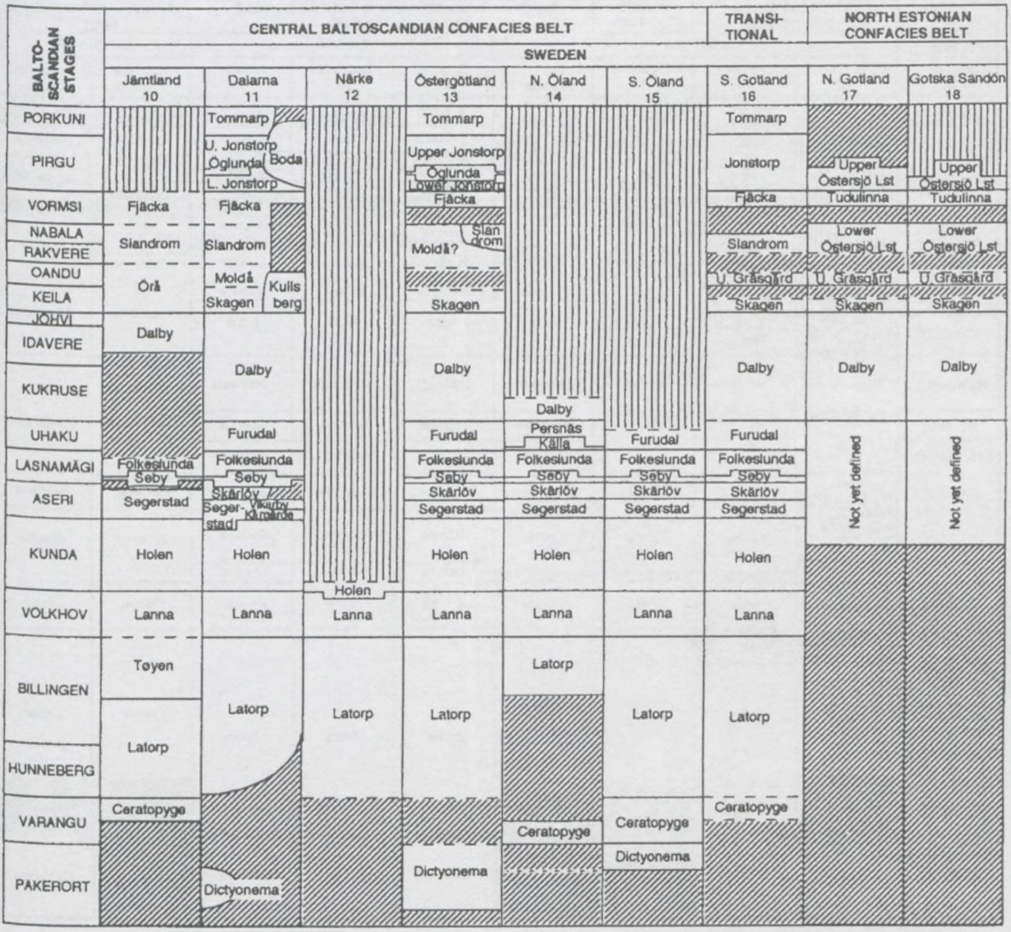


Fig. 4. Correlation table of the Ordovician of the Autochthonous of Jämtland (Bay of Brunflo; Jaanusson and Karis, 1982; Grahn, unpubl. data), Dalarna (Siljan District; Jaanusson, 1982), Närke (Tjernvik, 1956; Grahn, unpubl. data), Östergötland (Fylla Mosse and Smedsby Gård borings; Jaanusson, 1962, for the Smedsby Gård boring; Nõlvak and Grahn, unpubl. data), Öland (Jaanusson, 1982), and Gotland (Jaanusson, pers. commun., 1992; Nõlvak and Grahn, unpubl. data). Numbers at the heads of columns refer to Fig. 1.

Acanthochitina barbata Subzone

Definition: This subzone corresponds to the total range of *Acanthochitina barbata* Eisenack, 1931.
Type interval and locality: The holotype of *Acanthochitina barbata* was described from an erratic found at Samland, former East Prussia. As the type interval and locality for *Acanthochitina barbata* and the *A. barbata* Subzone we suggest the Kõrgessaare Formation, Vormsi Stage (early Ashgill), in the Orjaku boring (79.5–84.0 m), North Estonia (Nõlvak, 1980, fig. 2).
Characteristic chitinozoans: In Baltoscandia *Acanthochitina barbata* (Plate IV, A) is an excellent index species for the upper Vormsi Stage. At the top of the subzone primarily curved specimens of the genus *Rhabdochitina* occur.
Known occurrences outside Baltoscandia:

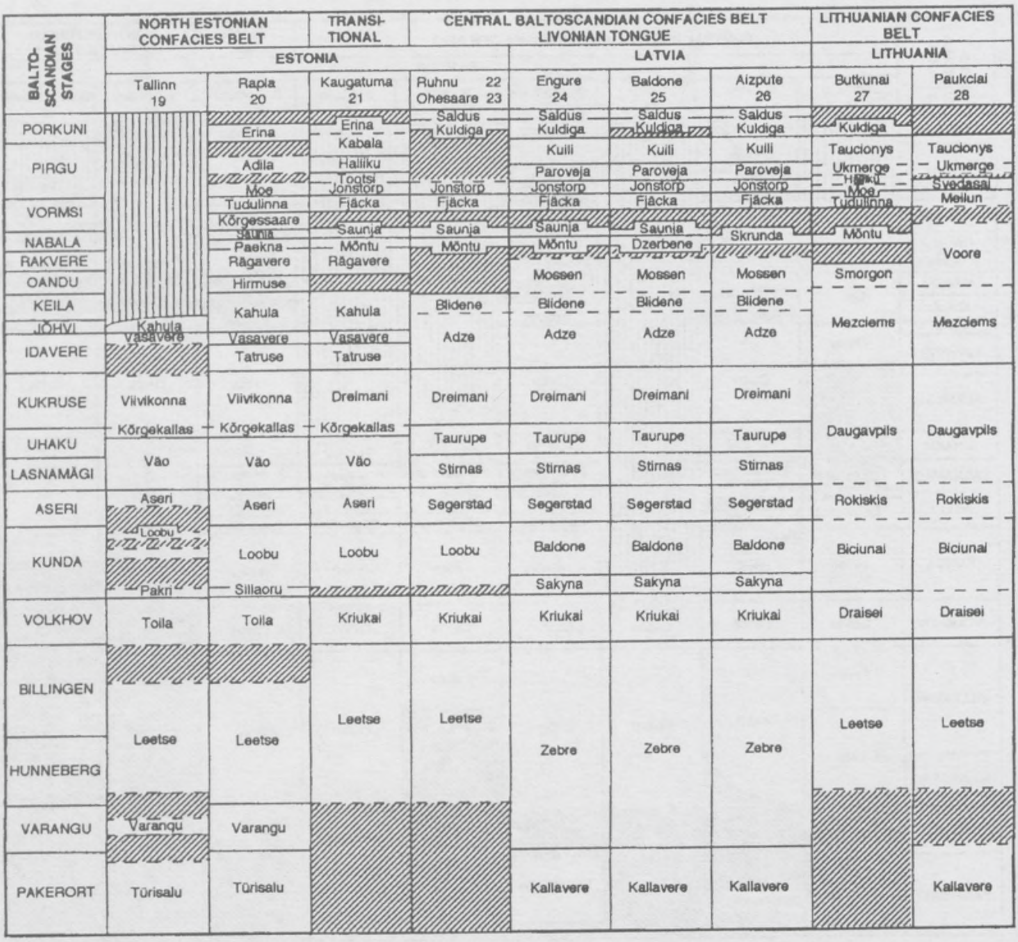


Fig. 5. Correlation table of the Ordovician of Estonia, Latvia and Lithuania (Männil, 1987; Nõlvak, unpubl. data). Numbers at the heads of columns refer to Fig. 1.

Acanthochitina barbata also occurs in northeast Poland (J. Nõlvak, unpubl. data). Jenkins (1967) described *A. barbata* from the Onnian (late Caradoc) of Shropshire, England. This is a somewhat older age than the occurrence in Baltoscandia. However, it should be noted that the Onnian specimens have a much denser ornamentation than the Baltic ones (Paris, pers. commun., 1992). Achab (1989) reported *A. barbata* from Québec and western Newfoundland, and Miller (1976) from the midwest USA. Both these occurrences are from beds of the same age as in Baltoscandia.

Remarks: The *A. barbata* Subzone corresponds to the late Vormsi time (Fig. 2; see Nõlvak, 1980).

Conochitina rugata Zone

Definition: This zone corresponds to the partial range of *Conochitina rugata* Nõlvak (in press),

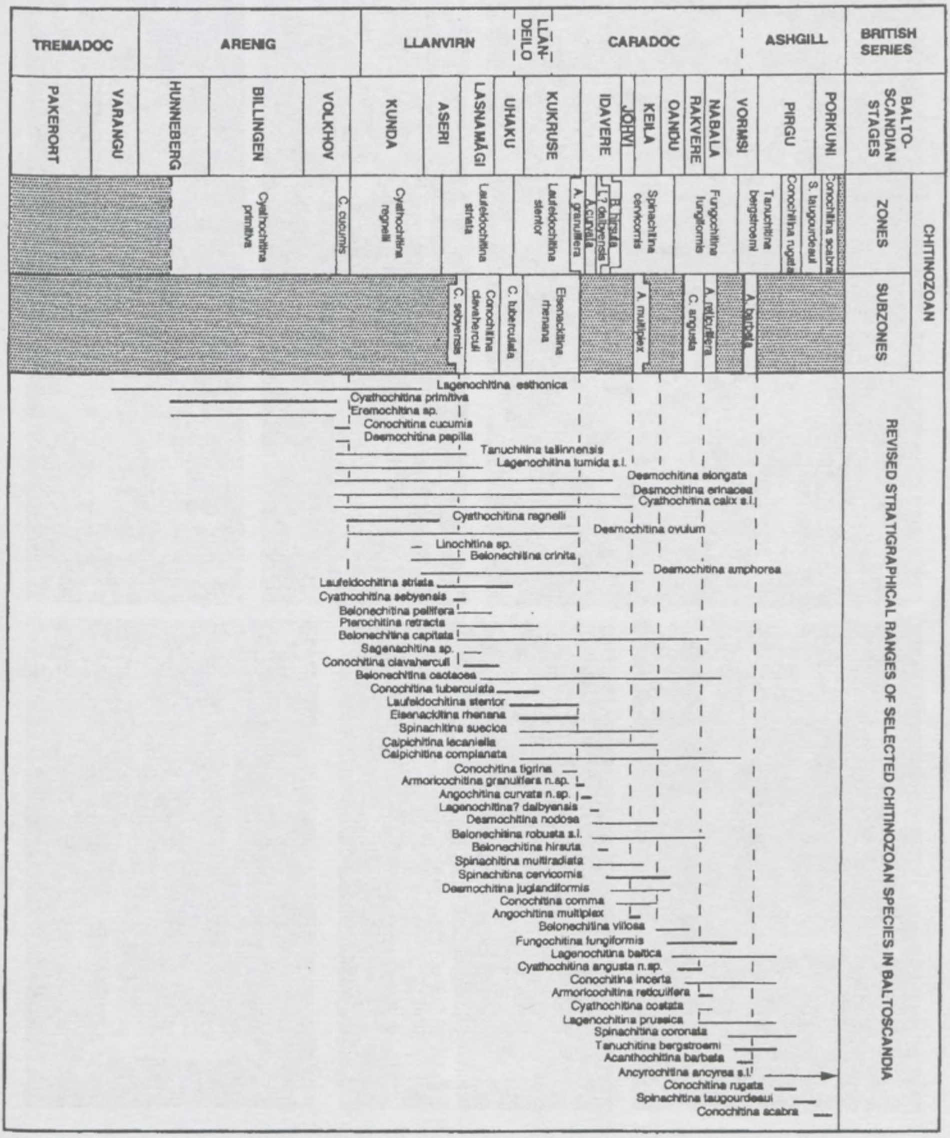


Fig. 6. Revised stratigraphical ranges of selected chitinozoan species in Baltoscandia.

PLATE IV

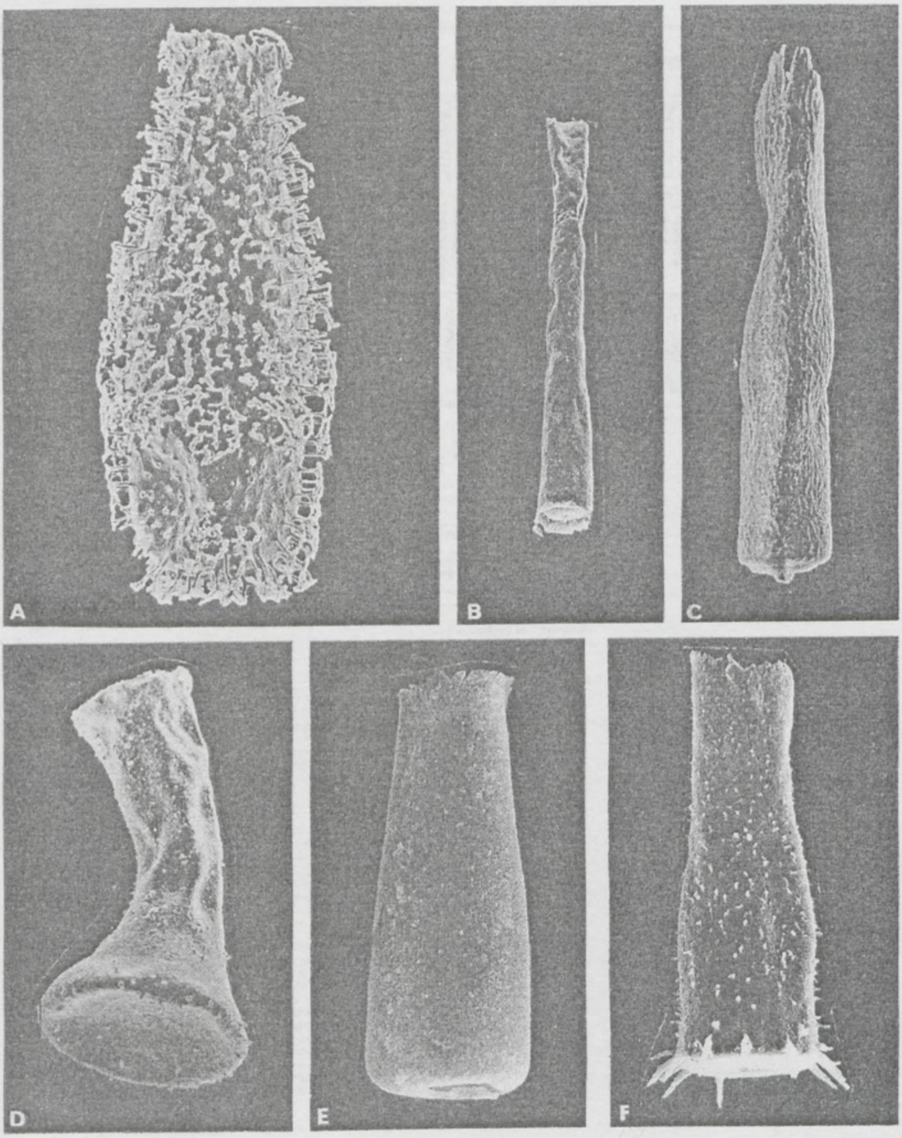
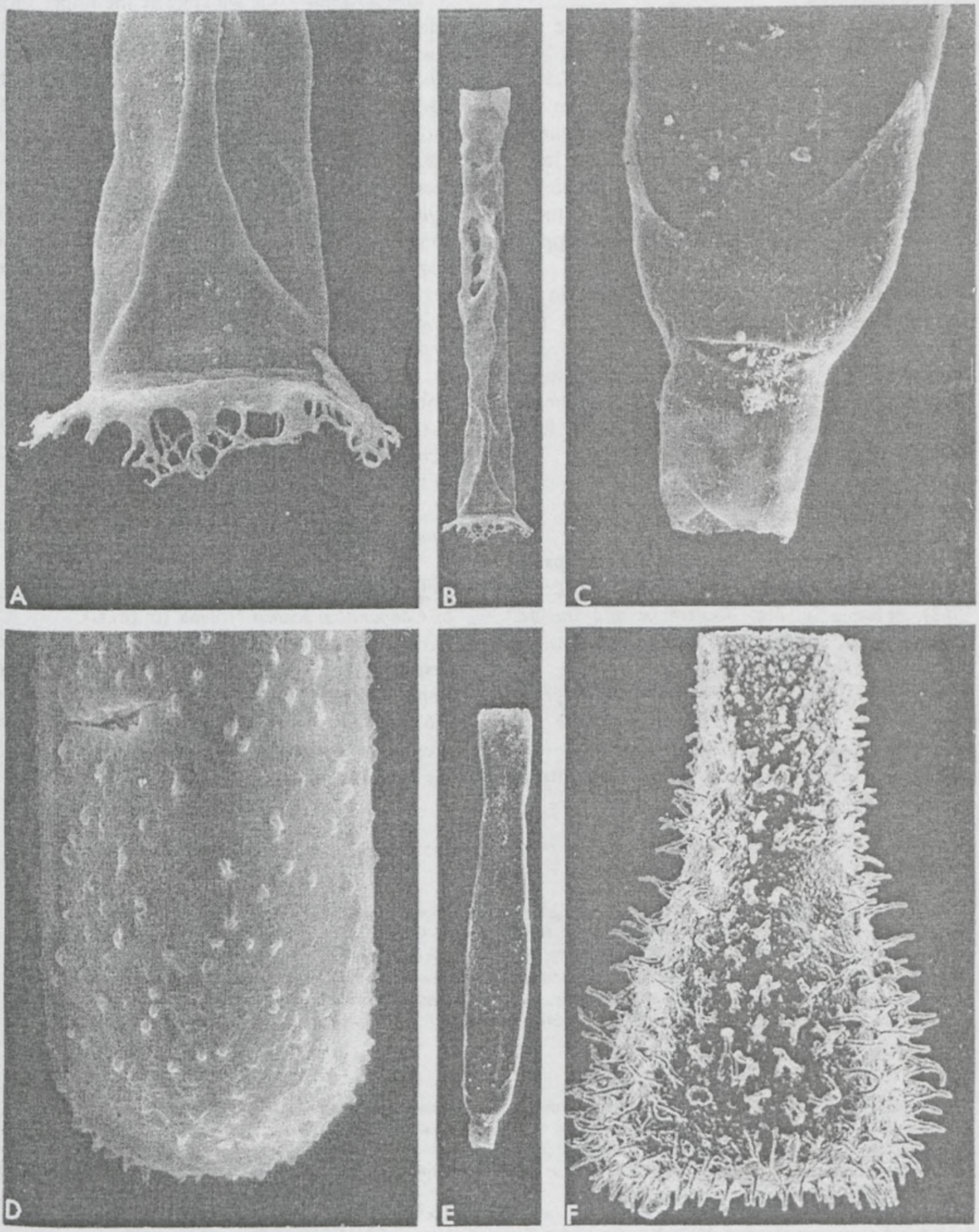


PLATE V



between its first occurrence and the first occurrence of *S. taugourdeaui*.

Type interval and locality: The holotype was described from the Adila Formation, Pirgu Stage (late Ashgill), in the Põltsamaa boring (125 m), Central Estonia (Nõlvak, in press). As the type interval and locality for the *C. rugata* Zone we suggest the Adila Formation, Pirgu Stage in the Hagudi boring (24.3–29.8 m), North Estonia.

Characteristic chitinozoans: At the top of the *C. rugata* Zone, *Spinachitina coronata* disappears (Fig. 6).

Known occurrences outside Baltoscandia: So far, *Conochitina rugata* (Plate IV, C) has not been reported outside Baltoscandia.

Remarks: The *C. rugata* Zone corresponds to late Pirgu time (Fig. 2). This species was assigned to *Conochitina* sp. by Nõlvak (1980, plate 29, fig. 4; 1990, plate 13, fig. 9).

Spinachitina taugourdeaui Zone

Definition: This zone corresponds to the partial range of *Spinachitina taugourdeaui* (Eisenack, 1968), between its first occurrence and the first occurrence of *Conochitina scabra* (Nõlvak, in press).

Type interval and locality: The holotype of *Spinachitina taugourdeaui* Eisenack, 1968 was described from an erratic found at Öjle Myr, Guldrupe, Gotland, Sweden. As the type interval and locality for *Spinachitina taugourdeaui* and the *S. taugourdeaui* Zone we suggest the upper Pirgu and lower Porkuni stages (late Ashgill), in the Kardla boring (181.3–184.2 m), southeast Estonia.

Characteristic chitinozoans: Specimens of the *Belonechitina micracantha*-type disappear at the Pirgu–Porkuni boundary, and a little higher, in the lowermost Porkuni, the specimens of the *Desmochitina minor*-group disappear as well. *Spinachitina taugourdeaui* (Plate IV, F) is a more common species in the East Baltic than in Scandinavia, and some rare specimens may occur together with *Conochitina rugata*.

Known occurrences outside Baltoscandia: *Spinachitina taugourdeaui* has also been described as *Conochitina taugourdeaui* from the Ellis Bay Formation, Anticosti Island, Canada (Achab, 1989).

Remarks: The *S. taugourdeaui* Zone corresponds to the late Pirgu–earliest Porkuni time (Fig. 2).

Conochitina scabra Zone

Definition: This zone corresponds to the total range of *Conochitina scabra* Nõlvak (in press).

Type interval and locality: The holotype was described from the Kuldiga Formation, lower Porkuni Stage (late Ashgill), in the Kardla boring (167.5–181.3 m), southeast Estonia (Nõlvak, in press). This is also type interval and locality for the *C. scabra* Zone.

PLATE IV

A. *Acanthochitina barbata* Eisenack, 1931. Hullo boring, Kõrgessaare Formation (21 m), Vormsi Stage, SEM, ×250.
B. *Tanuchitina bergstroemi* Laufeld, 1967. Moe boring, Moe Formation (11.9 m), Pirgu Stage, SEM, ×110.
C. *Conochitina rugata* Nõlvak (in press). Hagudi boring, Adila Formation (24.3 m), Pirgu Stage, SEM, ×230.
D. *Fungochitina fungiformis* (Eisenack, 1931). Kaugatuma boring, Mõntu Formation (411 m), Nabala Stage, SEM, ×410.
E. *Conochitina scabra* Nõlvak (in press). Kardla boring, Kuldiga Formation (167.5 m), Porkuni Stage, SEM, ×500.
F. *Spinachitina taugourdeaui* (Eisenack, 1968). Tartu boring, Unnamed formation (191.5 m), Pirgu Stage, SEM, ×425.

PLATE V

A. *Sagenachitina* sp. Detail of the basal part. Rapla boring, Väo Formation, Lasnamägi Stage, SEM, ×640.
B. *Sagenachitina* sp. Lateral view of the same specimen as in A, SEM, ×160.
C. *Eremochitina* sp. Detail of the basal part. Suhkrumägi, Leetse Formation, Hunneberg Stage, SEM, ×560.
D. *Conochitina tuberculata* Eisenack, 1962. Detail of the basal part. Rapla boring, Kõrgekallas Formation (170 m), Uhaku Stage, SEM, ×880.
E. *Eremochitina* sp. Lateral view of the same specimen as in C, SEM, ×96.
F. *Belonechitina hirsuta* (Laufeld, 1967). Kuusiku boring, Tatruse Formation (43 m), Idavere Stage, SEM, ×620.

PLATE VI

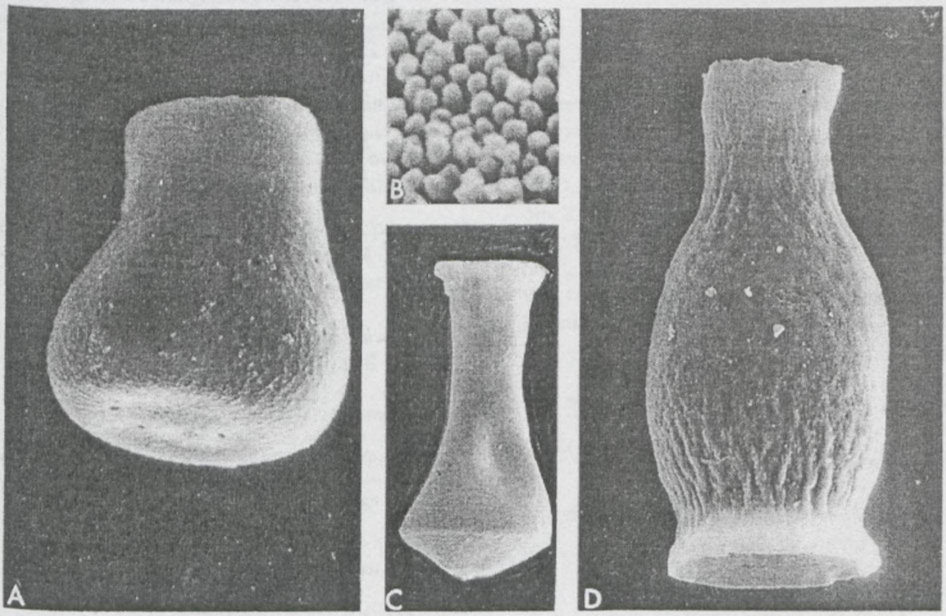


PLATE VI

A. *Eisenackitina rhenana* (Eisenack, 1939). Note the marks of broken spines. Viljandi boring, Viivikonna Formation (365 m), Kukruse Stage, SEM, × 880.
B. *Armoricochitina granulifera* n. sp. Detail of the surface. Imavere boring, Tatruse Formation (215.7 m), Idavere Stage, SEM, × 8000.
C. *Linochitina* sp. Rapla boring, Loobu Formation (182.6 m), Kunda Stage, SEM, × 400.
D. *Armoricochitina granulifera* n. sp. Lateral view of the same specimen as in C, SEM, × 340.

Characteristic chitinozoans: *Rhabdochitina gracilis* disappears just below the lower limit of the *Glyptograptus persculptus* graptolite zone. *Conochitina scabra* (Plate IV, E) disappears in the lower part of this graptolite zone. A few long-ranging species occur together with *Conochitina scabra*, for instance, *Ancyrochitina ancyrea* (Fig. 6) and *Cyathochitina campanulaeformis*. The first specimens of *C. scabra* may occur together with *S. taugourdeaui*.
Known occurrences outside Baltoscandia: So far, *Conochitina scabra* has not been reported outside Baltoscandia.
Remarks: The *C. scabra* Zone corresponds to late Porkuni time (Fig. 2). This species was assigned to *Conochitina robusta*, based on badly preserved specimens, in the uppermost Ordovician of Skåne (Scania), south Sweden (Grahn, 1978, figs. 4A,B,D,G).

Concluding remarks

Ordovician Chitinozoa, common in Baltoscandia, also occur in Belgium (Martin, 1973) and in the Moscow Basin (Umnova, 1969). This suggests that the chitinozoan zones, as defined in Baltoscandia, possibly may have applicability to the whole Ordovician continent of Baltica. However, further investigations of the Russian, Roumanian (Moesian Platform) and Belgian sequences are needed to confirm this. The durations of the chrons corresponding to these zones are some 1–5 My in most cases.

Acknowledgements

We are indebted to the following colleagues for providing us with comments on the manuscript; J. Bergström and V. Jaanusson (Swedish Museum of Natural History, Stockholm), M. Lindström (Stockholm University, Stockholm) and D. Kaljo (Institute of Geology, Tallinn). The work was supported by grants from the Swedish Natural Science Research Council (NFR G-GU 8811-302 for Y.G. and R-RA 8811-306 for J.N.). Our sincere thanks to all.

Appendix 1

***Angochitina curvata* n. sp.** (Plate II, B)

Derivation of name: Latin *curvus*, bent, referring to the overall shape of the vesicle.

Holotype: Ch 0408/6355 (Plate II, B).

Type horizon and locality: Tatruse Formation, Idavere Stage (early Caradoc), in the Udriku boring (63.70–63.75 m), North Estonia.

Diagnosis: Small *Angochitina* species with a primarily curved vesicle. The length is about three times the width. Body ovoid with a subcylindrical neck.

Description: The longitudinal axis is primarily curved which makes the vesicle asymmetrical. A flexure is therefore only present on one of the flanks, and the opposite flank forms a gentle curve from the base to the aperture. The greatest width occur in the middle of the body. Small simple spines randomly covers the whole vesicle, but also small coalescent spines may occur sporadically. The neck comprises about one third to half of the total length, and widens slightly towards the straight aperture.

Dimensions (15 specimens): Length 154–196 µm; greatest width 56–63 µm; width of aperture 28–35 µm; maximal height of ornamentation ca. 20 µm.

Remarks: *Angochitina curvata* is a rare species. Altogether about 100 specimens have been observed. In general 1–3 specimens can be found per 100 g of rock. It was designated as *Angochitina* (sp.) by Männil (1986). It differs from *Angochitina capillata* Eisenack 1937 in its primarily curved vesicle. Furthermore, the ornamentation of *A. capillata* has a tendency to be arranged in densely spaced longitudinal rows. Corroded specimens of *A. curvata* may also be confused with *Lagenochitina*? *dalbyensis* (Laufeld, 1967).

Occurrence: Lower part of the Idavere Stage (early Caradoc). Known from several sections of Estonia and Lithuania.

***Armoricochitina granulifera* n. sp.** (Plate II, D; Plate VI, B, D).

1981b *Cyathochitina reticulifera* Grahn - Grahn, fig. 5 M–P
1986 *Cyathochitina* cf. *reticulifera* Grahn - Männil, fig. 2.1.1.
1987 *Cyathochitina* cf. *reticulifera* Grahn - Männil, scheme 1.
1990 *Cyathochitina* aff. *reticuliferaa* Grahn - Nõlvak, plate 13, figs. 16–17.

Derivation of name: Latin *granulum*, a small grain, referring to the granulated surface of the vesicle.

Holotype: Ch 0774/7422 (Plate VI, A, C).

Type interval and locality: Tatruse Formation, Idavere Stage (early Caradoc), in the Imavere boring (215.7–215.9 m), North Estonia.

Diagnosis: Small *Armoricochitina* species with a short subcylindrical neck. The length is about twice the width. Body truncated ovoid.

Description: The maximum width of the vesicle is below half the length of the body. From this level towards the base occur longitudinal ribs. The body is ovoid to subconical with a rounded flexure and convex flanks. The neck is about a fourth of the total length and widens towards the straight aperture.The vesicle wall is covered with small granules (Plate VI, C). A carina is situated below the maximum width. The basal part is truncated, and on the base occur a few faint concentric ridges and a basal scar.

Dimensions: Length 182–244 µm; greatest width 98–189 µm; width of aperture 56–85 µm.

Remarks: *Armoricochitina granulifera* differs from *Armoricochitina reticulifera* (Grahn, 1981a) in having a granulate surface, larger carina and less developed concentric ridges on the base. The former species also has a somewhat more slender vesicle than *A. reticulifera*.

Occurrence: Lower part of the Idavere Stage (early Caradoc). Known from several sections of Sweden, Estonia and Lithuania.

***Cyathochitina angusta* n. sp.** (Plate III, D)

1962b *Cyathochitina* cf. *calix* Eisenack - Eisenack, plate 44, 16.
1982a *Cyathochitina calix* Eisenack - Grahn, figs. 5, 8.
1990 *Cyathochitina* cf. *calix* Eisenack - Nõlvak, plate 13, fig. 12.

Derivation of name: Latin *angustus*, narrow, referring to the overall shape which is slender for a *Cyathochitina* species.

Holotype: Ch 0481/5394 (Plate III, D).

Type interval and locality: Mõntu Formation, lowermost Nabala Stage (early Ashgill), in the Eikla boring (290.2 m), Saaremaa Island, Estonia.

Diagnosis: *Cyathochitina* species with a slender conical body and a short cylindrical neck. The total length is one and a half to twice the width. Reduced carina. Vesicle wall smooth.

Description: The maximum width of the vesicle is at the basal edge, which is provided with a reduced carina. A constriction is situated immediately above the basal edge towards the straight aperture. The body is conical, and the cylindrical neck comprises one third to half the total length. The vesicle wall is smooth. The base is provided with a basal scar. The flanks are straight to slightly convex, and the flexure is distinct.

Dimensions: Length 154–238 µm; greatest width 105–133 µm; width of aperture 42–49 µm.

Remarks: The species is identical to *Cyathochitina campanulaeformis angusta* in Estonian literature. It differs from *Cyathochitina calix* in having a reduced carina and a somewhat more stouter overall shape. *Cyathochitina campanulaeformis* is a much wider species provided with a larger carina.

Occurrence: Rakvere and the lowermost Nabala stages (latest Caradoc–early Ashgill). Known from many sections of Sweden, Estonia and Lithuania.

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Paper III

Nõlvak, J. 1997. Ordovician. Introduction.
In: Raukas A., Teedumäe A. (eds.). *Geology and Mineral Resources of Estonia.*
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ORDOVICIAN

Introduction

In the Ordovician, epicontinental seas with extensive distribution of carbonate sediments had a greater extent than in any other period. The marine flora and fauna changed markedly in the course of the Ordovician. A number of major taxonomic groups (bryozoans, brachiopods, echinoderms, trilobites, ostracodes, chitinozoans and others) appeared or became common. In this respect, the Ordovician is one of the most interesting periods in the history of marine faunas, and Estonia is among the areas in the world where this fauna is well preserved and studied.

The Ordovician was characterised by an extreme biogeographical differentiation of both planktic and benthic faunas, but with different degree. This makes the worldwide correlation of the Ordovician rocks difficult and has resulted in numerous regional stratigraphic schemes. A series of detailed stratigraphical charts compiled for the East Baltic (see Решения... 1978, 1981, Männil & Meidla 1994, and literature cited in these papers), gave a relatively stable detailed local classification for Ordovician rocks and afterwards obtained the status of a regional standard for most of the East-European Platform (Männil 1990).

The large-scale biogeographical and facies differentiation within the Ordovician Palaeobasin of Baltoscandia is well expressed in the concept of confacies belts (Jaanusson 1976, Fig. 24). The territory of Estonia is divided between the North Estonia and Central Baltoscandian confacies.

An emended version of the correlation chart, presented for the Estonian succession ranging from 70 to 180 m in thickness (Table 7), is based mainly on the above-cited schemes. For practical reasons the present version is simplified and many smaller subdivisions have been omitted. Some subdivisions, defined earlier as formations due to overestimations or difficulties in their specifications, are treated as members. Often the lithounits (Table 7) have diachronous (wavy line) or topical (discontinuous line) boundaries or the unit serves as "topostratigraphic" unit (*sensu* Jaanusson 1976, p. 310) with their boundaries coinciding with the stage boundaries.

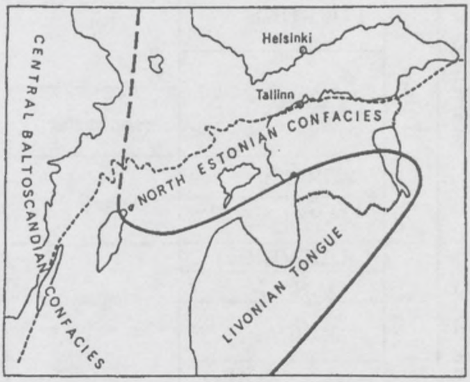
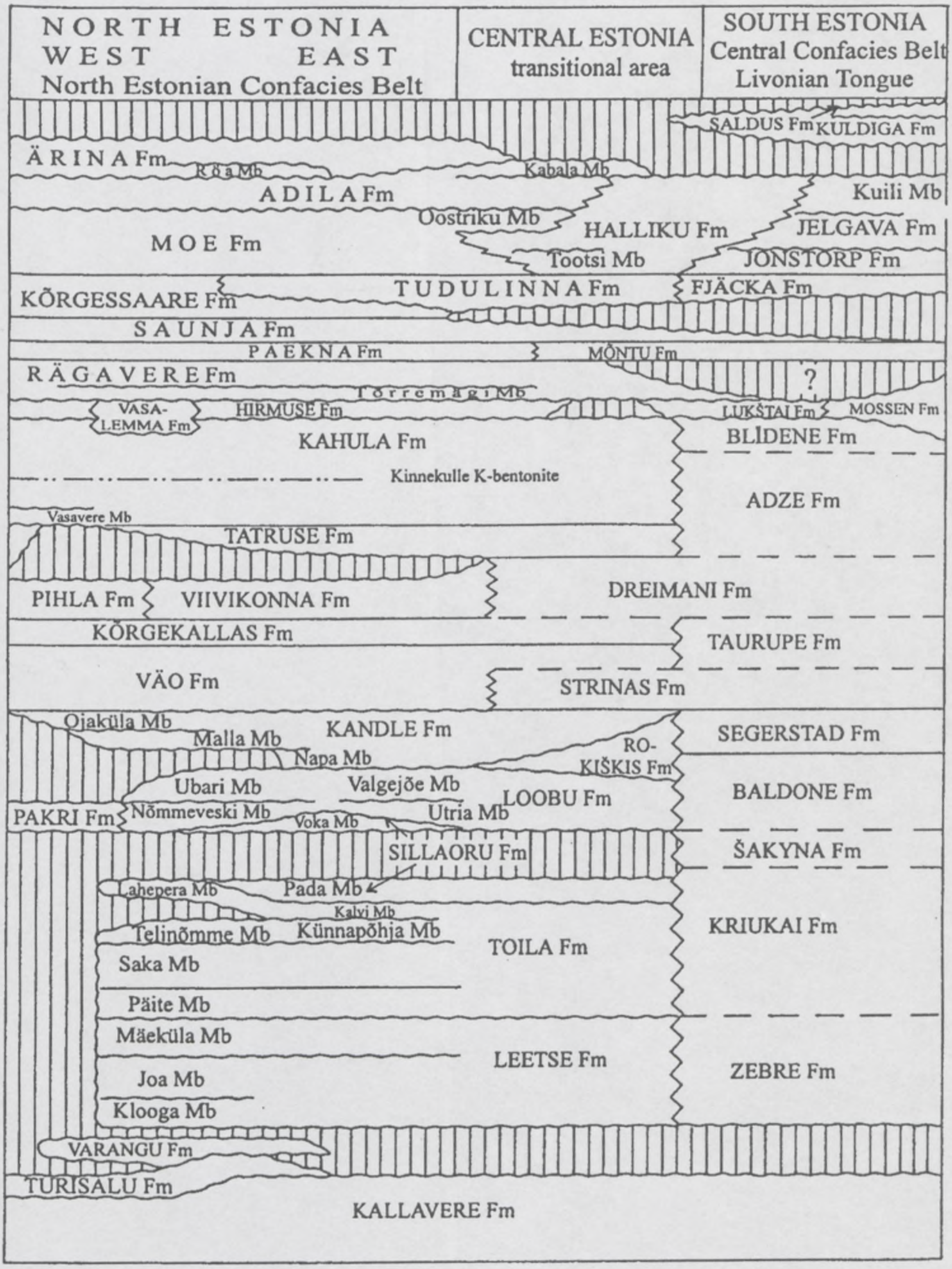


Fig. 24. Approximate boundary of the Ordovician confacies belts. Thin dashed line - the northern limit of the continuous distribution of rocks.

Table 7. The Ordovician of Estonia (*compiled by J. Nõlvak*)

British Series	Regional Series	East Baltic stages substages		Scandinavian graptolite zones	North Atlantic conodont zones and subzones		Baltoscandian chitinozoan zones and subzones	
ASHGILL	HARJU	PORKUNI		*persculptus* ?	*ordovicicus*		*scabra* *taugourdeaui*	
		PIRGU		*complanatus*			*rugata* *bergstroemi*	
		VORMSI		*linearis*				*barbata*
CARADOC		NABALA			*superbus*		*fungiformis*	*reticulifera* *angusta*
	VIRU	RAKVERE		*clingani*				
		OANDU						
		KEILA					*cervicornis*	*multiplex*
		HALJALA	JÕHVI	*multidens*	*tvaerensis*	*alobatus*	*hirsuta*	
			IDAVERE			*gerdae*	*dalbyensis* *granulifera*	*curvata*
LLANVIRN		KUKRUSE		*gracilis*		*variabilis*	*stentor*	*rhenana*
		UHAKU		*teretiusculus*	*anserinus* *serra*	*E. lindstroemi*		*tuberculata*
		LASNAMÄGI		*murchisoni*		*robustus* *reclinatus* *foliaceous*	*striata*	*clavaherculi* *sebyensis*
		ASERI			*suecicus*		*regnelli*	
	OELAND	KUNDA	ALUOJA	*artus*	*variabilis*			
			VALASTE					
ARENIG		HUNDERUM		*austrodentatus*				
		LANGEVOJA			*flabellum parva*		*cucumis*	
		VOLKHOV	VÄÄNA	*hirundo*	*originalis* *navis*		*primitiva*	
			SAKA		*triangularis*			
		BILLINGEN		*elongatus* *densus* *balticus*	*evae* *elegans*			
		HUNNEBERG		*phyllograptoides* *cupiosus* *murray*	*proteus*			
TREMADOC		VARANGU		*supremus* *hunnenbergensis*	*deltifer*		no chitinozoans	
		PAKERORT		*socialis-flabelliformis* *desmograptoides*	*angulatus* *lindstromi*			

Table 7. The Ordovician of Estonia (continued)



Paper IV

Nõlvak, J. 1999. Ordovician chitinozoan biozonation of Baltoscandia. *Acta Universitatis Carolinae. Geologica,* **43**, 1/2, 287-290.

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Ordovician chitinozoan biozonation of Baltoscandia

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INTRODUCTION

The correlation of Ordovician beds within Baltoscandia (the East Baltic, Scandinavia, northeast Poland) is complicated due to the differentiation of the sequence into at least five composite belts: North Estonian, Lithuanian, Central Baltoscandian, Scanian and Oslo Confacies belts (sensu Jaanusson 1976, see Nõlvak and Grahn 1993, Fig. 1). These belts are characterized by several contemporaneous lithofacies and in general the faunas remain different, expressed by the dominance of graptolites in the Scanian Confacies belt, in contrast to the four other belts where shelly fossils dominate. Among other fossil groups chitinozoans, regarded as being least influenced by facies variation, have proved to be very useful in solving some of the correlation problems. Application of chitinozoans in long-distance correlations between the major palaeogeographic domains, i.e. Baltica, Laurentia and North Gondwana (see Paris 1992; Van Grootel *et al.* 1996) will base on well-documented regional zonations. Such zonation has been proposed also for Baltoscandia (Nõlvak and Grahn 1993) and further studies will be focused on these stratigraphical levels which have a greater potential for a long-distance comparisons between different Ordovician basins. On the other hand, a more detailed biozonations are needed for the correlations of varied confacies belts within the Baltic Basin. Below some additional data will be given for some stratigraphical levels (circled numbers on Fig. 1), based on the new material from mainly East Baltic sections. Global subdivisions are given in accordance with Webby (1998), and two new zones and one subzone are proposed.

NEW DATA IN THE CHITINOZOAN DISTRIBUTION

(1) The oldest chitinozoans in the sections of the North Estonian Confacies belt *Lagenochitina esthonica* and *L.destombesi* hav been recorded from the uppermost part of the Varangu Stage, upper Tremadoc. The latter form, despite of its scarcity and often poor preservation, is widespread and has a high stratigraphic value. *L. destombesi* is considered as the oldest zonal form in North Gondwana (Paris 1990, 1992, Fig. 2). This new material was obtained from the stratotype section at Varangu, from the topmost 0.15 m layer. All efforts to find chitinozoans in the Ceratopyge Limestone (Mossebo, Sweden), which is equivalent, or somewhat younger Varangu strata by Erdtmann and Paalits (1995, Fig. 2), were unsuccessful. However, in the upper beds of the Bjorkasholmen Formation, in Roturnet section (Owen *et al.* 1990, Fig. 2) rare *L. esthonica* and *Belonechitina* sp. were identified.

(2) The *primitiva* Zone is poorly fossiliferous and represents a relatively long time. It has a little unclear boundaries caused by the following main reasons. In the sections of North Estonia there are remarkable stratigraphical gaps, which may be proved also by the absence of the conodont *B. triangularis* Zone (V. Viira in Einasto *et al.* 1996, Fig. A15; and not in Meidla 1998, Fig. 31, where are serious drawing mistakes) at the level of a specific discontinuity surface at the top of the Billingen Stage described by Orviku (1961). So, in the Tallinn area the global Lower/Middle Ordovician boundary can be followed on this level, but cannot be defined in chitinozoan record. Another reason is extensive secondary dolomitization in the North Estonian and Lithuanian belts which destroyed all groups of organic-walled microfossils. In the stratigraphically more complete sections of Livonian Tongue barren redbeds are common.

Most probably *C. symmetrica* can also be recognized in the population defined earlier as *Cyathochitina primitiva* in Baltoscandia. This important taxon having a worldwide distribution is reported from Laurentia and North Gondwana plates (Paris 1992). Some badly preserved forms from glauconitic sandstones of the Hunneberg Stage need a revision.

(3) The next important level for long-distance correlations is the uppermost *clavaherculi* Subzone where the conodont *E. robustus* and graptolite *Gymnograptus linnarssoni* also occur in the lowermost Uhaku Stage (see Jaanusson 1963; Männil 1976). However, it is not clear whether this highly diverse and fossiliferous level can be compared with that of Bucaco, Portugal (Henry *et al.* 1977, Paris 1990) or these beds have a good stratigraphical potential only within the Baltic Basin.

(4) Considerable similarity between the Baltica and North Gondwana sections can be followed in the succession associated with the *B. gerdae* conodont Subzone (see also Paris 1992). These relatively short-ranging zones were recently established also in the Fjäcka section (Dalarna, Sweden - Nõlvak *et al.* 1999), which is the type section of the widely distributed *A. tvaerensis* conodont Zone (Bergström 1971). This *L. dalbyensis* - *L.* aff. *capax* - *L. deunffi* assemblage in the lower Idavere Substage of Baltoscandia seems to be the same as in zones 12 – 13 of Paris (1981, Table 42). The latter form was

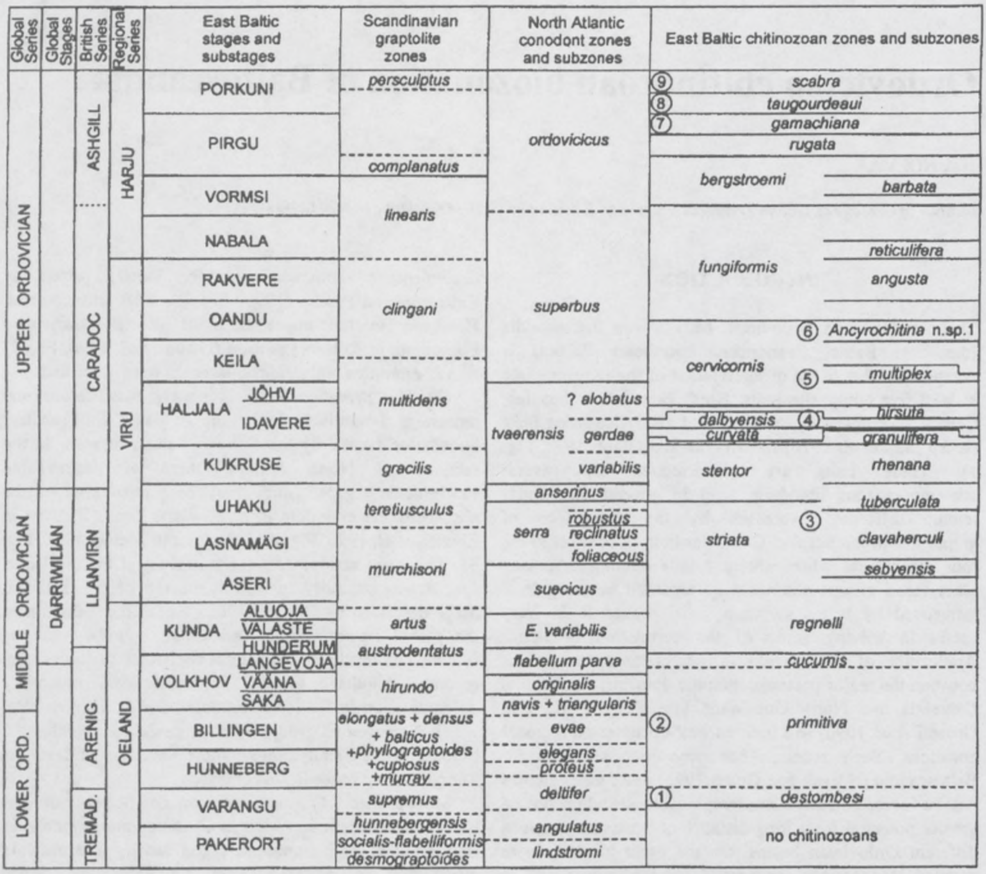

Fig. 1 Revised Ordovician Chitinozoan biozonation.

recently recovered from the Baldone boring, Latvia, in which they occur near by. In addition, the whole assemblage of chitinozoans in these beds is very similar (see faunal log in Grahn *et al.* 1996); in Baltoscandia it is only a little more diverse. During early Caradoc probably direct contact existed between the Baltica and North Gondwana palaeocontinents.

(5) Among chitinozoans this level is marked by rare but widely distributed *L. multiplex.* A thin bed with this species rests just above the well-known Kinnekulle bentonite layer (Bergström *et al.* 1995) together with *Hercochitina* sp. which range is a little wider. Recently the latter was found from the Lindlandveien section (Oslo-Asker, Norway) just above the same bentonite.

(6) In the uppermost part of the *cervicornis* Zone a new tentative subzone(?) is proposed (see below). According to the evidence from chitinozoan distribution no undoubted Mossen equivalents at present can be distinguished. Whether this depends upon a break in the North Estonian sequence, poorly fossiliferous condition of its lower part, represented by dark organic-rich shales or biogeographic differentiation is not known. Somewhat specific assemblage gives a hint to the former.

(7) A new *gamachiana* Zone is defined (see below) in the uppermost Pirgu Stage between the earlier *rugata* and *taugourdeaui* zones. In the Baltic Basin there is a gap in the most areas, except in the sections from the so-called transitional area (or zone) between North Estonian – Lithuanian belts and Livonian Tongue, where sections are stratigraphically more complete. Very important for the long-distance correlation is the fact that on Anticosti (Achab 1981,1989) the ranges of *gamachiana* and *taugourdeau*i are in the same order in the topmost part of Ordovician.

(8) The lower boundary of the Porkuni Stage in Baltoscandia could be defined as the appearance level of the continual distribution of *S. taugourdeaui,* although the lower boundaries of this zone and the Porkuni Stage are difficult to define in some poorly preserved beds in North Estonia, including the stratotype area at Porkuni.

(9) The *scabra* Zone is absent in the Anticosti section (Achab 1989); it is equivalent with the *persculptus* graptolite Zone in Baltoscandia and marks the appearance of the so-called Hirnantian macrofauna in the East Baltic sections (see Kaljo *et al.* 1998). The lower boundary of the

scabra Zone coincides with a drastic decrease in chitinozoan diversity marking the mass extinction level in Baltoscandia.

NEW BIOSTRATIGRAPHIC SUBDIVISIONS

Lagenochitina destombesi Zone

Definition: This zone corresponds to the total range of *Lagenochitina destombesi* Elaouad-Debbaj 1988.

Type interval and locality: The holotype was described from the Lower Fezouata Formation, Anti-Atlas, southern Morocco and is so far the oldest recorded taxon. As the type interval in North Estonia, the upper Varangu Formation is suggested, the uppermost 0.15 m of clays, just below the glauconitic sandstone of the Leetse Formation. It is so far the only locality in Baltoscandia.

Characteristic chitinozoans: only *Lagenochitina esthonica* coexists.

Known occurrence outside Baltoscandia: This biozone is known in Morocco and is the oldest in North Gondwana (Paris 1990, 1992).

Ancyrochitina sp. n. 1 Subzone

Definition: This subzone corresponds to the total range of *Ancyrochitina* sp. n. 1 (*nom. nud.*).

Type interval and locality: The holotype will be described from the Mossen Formation. The type interval is the lower part of the Mossen (Nõlvak and Grahn 1993, Fig. 5), from the boundary beds of the Keila and Oandu stages in the Adze boring (880.4-881.8 m), Latvia.

Characteristic chitinozoans: among more characteristic taxa rare *Spinachitina cervicornis* occur together with *Desmochitina juglandiformis, Belonechitina robusta.*

Belonechitina gamachiana Zone

Definition: This zone corresponds to the partial range of *Belonechitina gamachiana* (Achab 1978), between its first occurrence and the appearance of *Spinachitina taugourdeaui* (Eisenack 1968), which is the index species of the succeeding biozone.

Type interval and locality: The holotype of *Belonechitina gamachiana*, was described as *Conochitina* from the Ellis Bay Formation, Anticosti, Canada (Achab 1978) and is the type taxon of *gamachiana* Zone (Achab 1989). As the type interval for the zone in Baltoscandia the Kabala Formation in Kaugatuma boring (343.2-347.7 m) is suggested, Island Saaremaa, western Estonia, in the upper part of the Pirgu Stage (Fig. 5 in Nõlvak and Grahn 1993). This zone can be followed also in Kardla boring (Central Estonia), in the Taucionys Formation in Iljinskoje boring (Pskov District, Russia; Nõlvak *et al.* 1989) and Butkunai boring (Lithuania; Nõlvak and Grahn 1993, Fig. 5).

Characteristic chitinozoans: A specific form of *Calpichitina complanata* (as *Desmochitina complanata tunica* (nom. nud.). in Nõlvak *et al.* 1989 Fig. 1), *Ancyrochitina ancyrea* s.l. including *A. longispina.*

Known occurrence outside Baltoscandia: It has been described on Anticosti, Canada (see also Paris 1992). In the topmost part of the Langara Formation in the Sandvika section (Oslo-Asker, Norway) in the beds with *Holorhychus giganteus* there occurs together with *Spinachitina? bulmani* and *Rhabdochitina magna* by Grahn *et al.* (1994) also *B. gamachiana*.

Remarks: This zone corresponds to late Pirgu time, but due to its position between the *rugata* and *taugourdeaui* Zones, the range of the former should be specified: the rugata Zone corresponds to the interval between the first occurrence and the appearance of *B .gamachiana* or *S .taugourdeaui* in the Baltoscandia (Nõlvak and Grahn 1993, p.264).

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Paper V

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CHITINOZOANS IN BIOSTRATIGRAPHY OF THE NORTHERN EAST BALTIC ASHGILLIAN. A PRELIMINARY REPORT

NÕLVAK, J.: Chitinozoans in biostratigraphy of the northern East Baltic Ashgillian. A preliminary report. Acta Palaeont. Polonica, 25, 2, 253—260, July 1980.

Acid-resistant microfossils have been studied from serial bore-core samples representing different facies of Ashgillian deposits. Chitinozoans are shown to hold better promise for detailed subdivision and section correlation than other scarce representatives of macro- or microfauna. According to their vertical distribution, three types of chitinozoan species having biostratigraphic significance can be distinguished.

Key words: chitinozoans, Ashgillian, East Baltic, biostratigraphy.

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INTRODUCTION

The last ten years have seen a wide application of acid-resistant microfossils in the biostratigraphic studies of the Baltic Ordovician and Silurian. This is chiefly due to an ever increasing importance of bore-core material. The initial results of acid-resistant Ordovician and Silurian microfossil distribution (chitinozoans in particular) hold a great promise for the section subdivision and correlation (Männil 1970, 1971, 1972; Nestor 1976*a*, *b*). Made below is an attempt to show the importance of individual groups of acid-resistant microfossils for the Ashgillian biostratigraphy on the basis of studying serial samples from 12 core section, the northern East Baltic.

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GEOLOGICAL SETTING

The sections under study are located in the following confacies belts (see Männil 1966; Põlma 1967, Jaanusson 1976): the northern (Estonian) facies belt: the Hagudi, Orjaku, Kiriküla borings; the transition belt: the Eikla, Kingissepa, Kangatuma, Ohesaare borings; the central (Swedish-Latvian) belt: the Kolka, Ruhnu, Ikla, Viesate borings (fig. 1).

In the present report the Ashgillian is tentatively considered to embrace the Vormsi, Pirgu and Porkuni Stages. This is done due to the difficulties in defining the lower boundary of the Ashgillian, especially, after the revision of the British stratotype Ashgillian (see Ingham and Wright 1970; Williams *et al.* 1972).

Discussing the development of the Ordovician basin in the East Baltic, Männil (1966) discriminates the Ashgillian as spanning the Pirgu and Porkuni age, and points out its being fairly distinct. As compared with earlier stages in the basin development the Ashgillian is a regressive one.

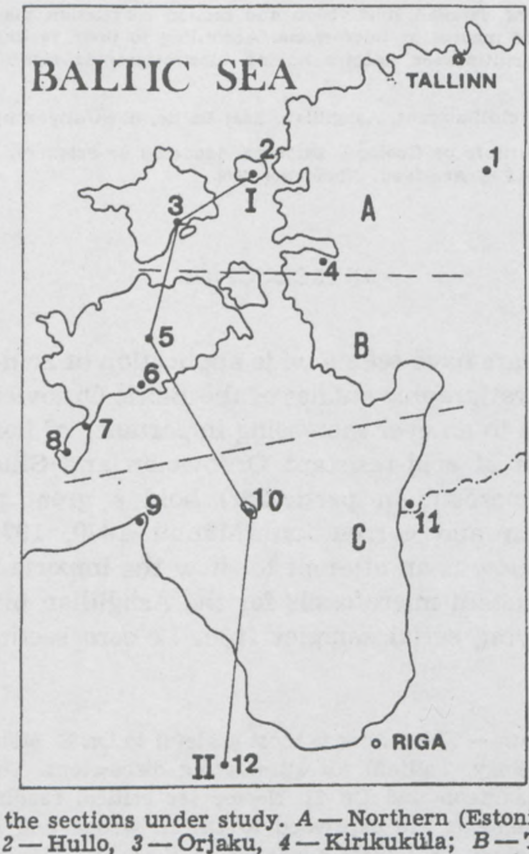

Fig. 1. Location of the sections under study. *A* — Northern (Estonian) structural-facies belt: *1* — Hagudi, *2* — Hullo, *3* — Orjaku, *4* — Kirikuküla; *B* — Transition belt: *5* — Eikla, *6* — Kingissepa, *7* — Kaugatuma, *8* — Ohesaare; *C* — Axial (Swedish-Latvian) belt: *9* — Kolka, *10* — Ruhnu, *11* — Ikla, *12* — Viesate. I—II — The line showing the location of the sections in fig. 2.

The beginning of the Ashgillian is associated with a relative stability of the basin which is accounted for by a distinct succession of lithofacies. Replacing one another towards the open sea are (fig. 2):

1) grey biodetrital carbonate deposits (Kõrgessaare Formation): the Hagudi, Orjaku borings and others;

2) grey carbonate argillaceous deposits (Tudulinna Formation): the Eikla, Kaugatuma borings and others;

3) thin grey argillaceous deposits (Fjäcka Formation): the Ruhnu, Ikla borings and others;

4) dark grey argillaceous deposits enriched with organic material (Fjäcka Formation): the Viesate boring and others.

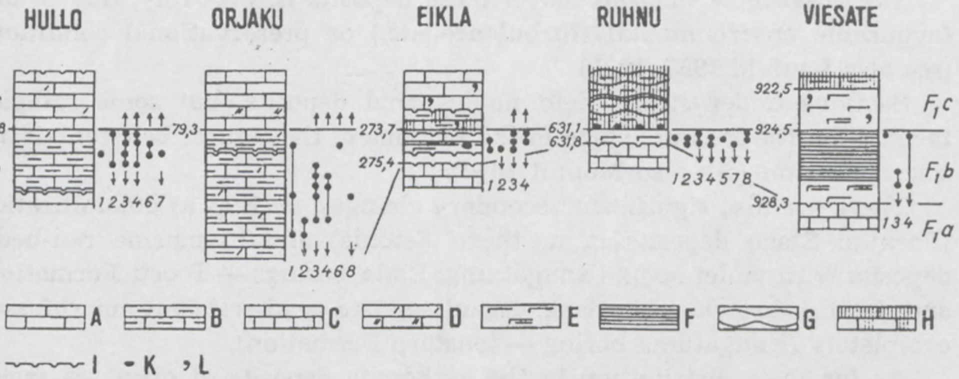

Fig. 2. Chitinozoan distribution in the upper Vormsi Stage of the sections in western North Baltic (the columns are based on the data provided by L. Põlma, E. Kala and the author). F_1a — the Nabalo Stage; F_1b — the Vormsi Stage; F_1c — the Pirgu Stage. *A* — limestone; *B* — clayey limestone; *C* — cryptocrystalline limestone; *D* — fine-biodetritic limestone (a) and coarse-biodetritic limestone (b); *E* — calcitic marl (a) and clayey marl (b), *F* — graptolitic argillite; *G* — medium-nodular limestone; *H* — completely red beds (a) spotted red beds (b); *I* — discontinuity surface; *K* — goethite oolites; *L* — glauconite. 1 — *Acanthochitina barbata*, 2 — *Cyathochitina kuckersiana*, 3 — *Lagenochitina baltica*, 4 — *Lagenochitina prussica*, 5 — *Conochitina* cf. *elegans*, 6 — *Tanuchitina bergstroemi*, 7 — *Desmochitina minor* f. *typica*, 8 — *Desmochitina* sp. The arrow marks the species having a wider range of vertical distribution.

In the northern East Baltic, the boundary between the Vormsi and Pirgu Stages displays a significant change in the nature of sedimentation.

The Pirgu Stage deposits are distinguished from those lying below by their greater differentiation and fair scarcity of macrofauna. This is probably due to a higher sedimentation rate (see Põlma 1973; Oraspõld 1975).

At the boundary level of the Pirgu and Vormsi Stages there also occurs a considerable decrease in various microfossil forms. According to the data available, no zonal forms associated with the above stages have been up till now observed in the northern East Baltic as far as ostracodes (Sarv 1959, 1962, Gailite 1968; and others) and conodonts (Viira 1968, 1974) are concerned.

FO

CHITINOZOANS AND OTHER ACID-RESISTANT MICROSSILS

Of greatest stratigraphic importance for the Ashgillian are graptoids, chitinozoans and acritarchs. Their occurrence depends, to a great extent, on the type of deposits. It appears that those deposits which have formed in fairly shallow waters, that is in lagoonal and shoaly facies belts following the Silurian basin model, contain no microplankton. Among them are:

1. Biohermal limestones of the shoaly facies belt in the Pirgu (Kaugatuma boring — Jonstorp Formation) and Porkuni (Hagudi boring — Törevere Member) Stages. The same is true for Sweden (Laufeld 1967, 1973).
2. Deposits of the Porkuni stage (Kaugatuma, Ruhnu borings — Piltene Member) containing coarse terrigenous material (quartz, carbonate oolites).

The absence of microfossils in these deposits is, probably, due to unfavourable environmental (turbulence etc.) or preservational conditions (see also Laufeld 1967, 1973).

Sections under study yield no lagoonal deposits, but some samples taken from the Porkuni stage in the sections of the central belt contain no microplankton (see also Männil 1970).

There are also significant secondary changes, such as a) dolomitization (Porkuni Stage deposits in northern Estonia) and b) marine red-beds: deposits with violet spots (Kaugatuma, Eikla borings — Toosti Formation) and reddish-brown rocks lack organic-walled acid-resistant microfossils completely (Kaugatuma boring — Ionstorp Formation).

As for their distribution in the carbonate deposits of open sea facies belts, chitinozoans show no basic distinction from representatives of other microfossil groups, for instance, graptoloids. On the basis of graptoloid distribution Männil (1976) recognizes three types of zonation which, probably, are also valid in the case of chitinozoans.

Type I. Zones characterized by the continuous presence of the zonal species; *Cyathochitina* cf. *dispar* (pl. 29: 2; at the lowermost Nabala Stage), *Acanthochitina barbata* (pl. 29: 1; at the uppermost Vormsi Stage); *Conochitina* sp. (pl. 29: 4; in the middle of the Pirgu Stage), *Conochitina taugourdeaui* (pl. 30: 6a, 6b; at the uppermost Pirgu Stage and in the Porkuni Stage). These zones can be very well applied for correlation of deposits developed in different facies.

Type II. Zones which yield the zonal species only within a certain interval (zonule, according to Männil's terminology 1972); for the Ashgillian these are: *Coronochitina coronata* (pl. 30: 1a, 1b), *Cyathochitina campanulaeformis* (pl. 29: 8), *Cyathochitina kuckersiana* (pl. 29: 7) and others. They can be chiefly used for correlation of closely spaced sections and also within certain confacies belt (according to the unpublished data of Männil 1969, and Nõlvak 1972).

Type III. Species having a wide range of vertical distribution: *Conochitina minnesotensis* (pl. 30: 3), *Conochitina micracantha* (pl. 29: 6), *Rhabdochitina gracilis* and others. These species, at the present stage of our knowledge about them, cannot be used for detailed stratigraphic studies, but they are important when deposits are correlated at the stage level.

One of such chitinozoan index species (type I) for the deposits under study is *Acanthochitina barbata* whose stratigraphic significance has been pointed out by Laufeld (1967) and Männil (1971). This zone can be traced in all the known sections in the Baltic irrespective of their facies nature (fig. 2). The occurrence of *Acanthochitina barbata* in association with *Tanuchitina bergstroemi* (pl. 30: 2a, 2b), *Lagenochitina baltica* (pl. 30: 5). *Lagenochitina prussica* (pl. 29: 3) and others, is evidence, on the one hand, of contemporaneity of beds belonging to different facies, while, on the other, it allows a reliable conclusion that the occurrence of some chitinozoan forms shows little dependence on the facies conditions.

Acritarchs (which were picked up with chitinozoans) are confined chiefly to the upper half of the Pirgu Stage where, in some places, they display mass occurrence. The levels at which some of the species make their first appearance can be accepted as secondary criteria for section subdivision.

Graptoloid rhabdosomes indentifiable to the species level are fairly scarce in the Ashgillian (see Männil 1976), while the siculae, which are more common but so far underterminable reveal a sporadic distribution.

The serial samples (about 300 samples with the rock weight of 200—300 grams) yield very few coniform conodonts (*Panderodus, Acodus)* are quite scarce.

Scolecodonts, melanosclerites and foraminifera are fairly common but unevenly distributed; their occurrence is largely accounted for by the facies control.

A similar relationship in the Silurian deposits has been pointed out by Männil (1973). These groups being insufficiently studied, the evidence on their distribution is still lacking but they are of importance for section subdivision at the horizon and stage levels as well as for palaeoecological investigations.

The distribution of acid-resistant microfossils (chiefly chitinozoans) enables us to subdivide the deposits of the Vormsi Stage into two and those of the Pirgu Stage into six biostratigraphic units and to correlate them in the northern East Baltic. Strong dolomitization in the North Estonian sections limits the stratigraphic significance of the Porkuni Stage microfossils to the sections of the transition structural-facies belt (fig. 1)

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JAAK NÕLVAK

BIOSTRATYGRAFICZNE ZNACZENIE CHITINOZOA Z ASZGILU PÓŁNOCNO-WSCHODNIEJ CZĘŚCI OBSZARU BAŁTYCKIEGO. NOTATKA WSTĘPNA

Streszczenie

Omówiono Chitinozoa z systematycznych prób skał aszgilskich różnych facji z wierceń w północno-wschodniej części obszaru bałtyckiego. Stwierdzono większą użyteczność tych mikroskamieniałości dla szczegółowych podziałów stratygraficznych i korelacji profilów, w porównaniu z innymi rzadkimi w tych utworach mikro- i makroskamieniałościami. Na podstawie rozmieszczenia Chitinozoa w profilu, można wydzielić wśród nich trzy grupy gatunków, które mają różne znaczenie biostratygraficzne.

ЯАК НЫЛВАК

ХИТИНОЗОИ В БИОСТРАТИГРАФИИ АШГИЛЛА СЕВЕРНОЙ ПРИБАЛТИКИ. ПРЕДВАРИТЕЛЬНОЕ СООБЩЕНИЕ

Резюме

Изучены кислотоустойчивые микрофоссилии из серийных проб ашгиллских разнофациальных отложений из буровых скважин западной части Северной Прибалтики. Отмечается перспективность хитинозой для дробного расчленения и кор-

реляции разрезов, по сравнению с другими редкими представителями макро- и микрофауны. Среди хитинозой, по вертикальному распределению можно выделять три типа видов, имеющие разное биостратиграфическое значение.

EXPLANATIONS OF THE PLATES 29, 30

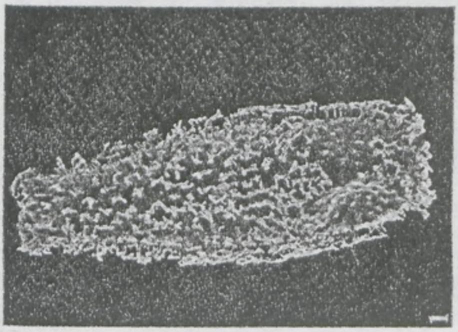
Selected chitinozoans from Upper Caradocian and Ashgillian. The well preserved specimens used for illustrations (SEM-photographs) were selected not only from boreholes and stages discussed in text.

Plate 29

1. *Acanthochitina barbata* Eisenack 1931. Hullo, 21,0 m; Vormsi Stage. Ch 701/5670. Almost complete specimen in lateral view, x 125.
2. *Cyathochitina* cf. *dispar* Benoit et Taugourdeau 1961. Eikla, 289,5 m; Nabala Stage. Ch 702/5395. Complete specimen in oblique lateral view. Note the perforations caused by parasites, x 125.
3. *Lagenochitina prussica* Eisenack 1931. Ruhnu, 631,7 m; Vormsi Stage. Ch 703/5840. Specimen in oblique aboral view. x 220.
4. *Conochitina* sp. Hagudi, 24.3 m; Pirgu Stage. Ch 704/6483. Aboral part of specimen in oblique lateral view, x 310.
5. *Desmochitina minor* Eisenack 1931. Kaugatuma, 404.5 m; Nabala Stage, Ch 705/5791. Specimen in oblique aboral view, x 620.
6. *Conochitina micracantha* Eisenack 1931. Hullo, 35.5 m; Nabal Stage. Ch 706/5650. Specimen in oblique lateral view, x 220.
7. *Cyathochitina kuckersiana* (Eisenack 1934). Hullo, 21.5 m; Vormsi Stage. Ch 707/5669. Specimen in oblique lateral view, x 125.
8. *Cyathochitina campanulaeformis* (Eisenack 1931). Eikla, 275.5 m; Nabala Stage. Ch 708/5412. Specimen in oblique aboral view, x 125.

Plate 30

1. *Coronochitina coronata* (Eisenack 1931). Hagudi, 35.5 m: Pirgu Stage. Ch 709/6478: *a* almost complete specimen in oblique lateral view, x 125; *b* broken appendices at the basal edge in oblique aboral view, x 4800.
2. *Tanuchitina bergstroemi* Laufeld 1967. Hullo, 21.0 m; Vormsi Stage. Ch 710/5670: *a* somewhat flattened complete specimen in oblique lateral view, x 65; *b* aboral part, x 310.
3. *Conochitina minnesotensis* (Stauffer 1933). Hullo, 47.5 m; Rakvere Stage. Ch 711/5634. Somewhat atypical, complete specimen in lateral view. Note the curvature of the vesicle and well developed asymmetrical basal callus at the base, x 65.
4. *Conochitina* cf. *elegans* Eisenack 1931. Hullo, 25.5 m; Vormsi Stage. Ch 712/5661. Somewhat flattened specimen in lateral view, x 125.
5. *Lagenochitina baltica* Eisenack 1931. Orjaku, 96.6 m; Nabala Stage. Ch 713/3082. Specimen in oblique lateral view, x 125.
6. *Conochitina taugourdeaui* Eisenack 1968. Undva, 152.4 m; Pirgu Porkuni? Stage. Ch 714/6186: *a* specimen in oblique aboral view, x 125; *b* aboral part, x 310.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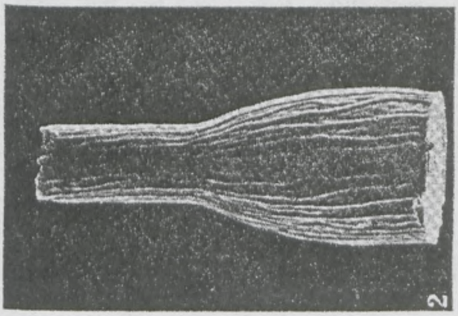
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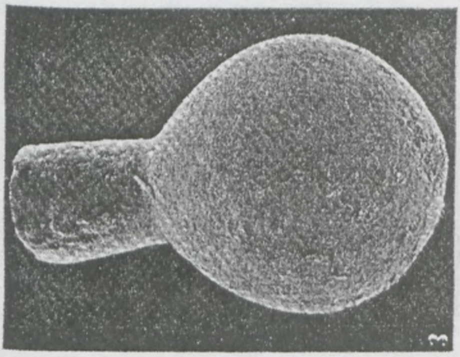
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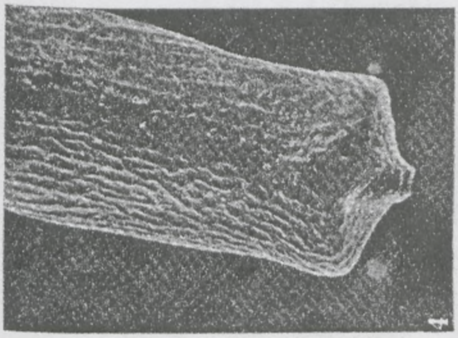
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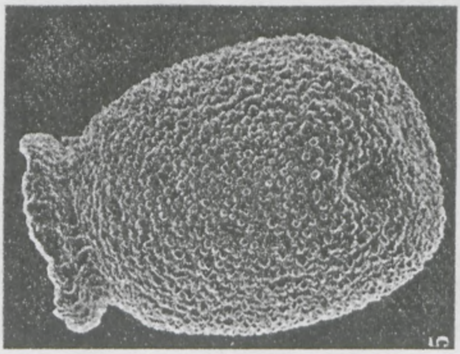
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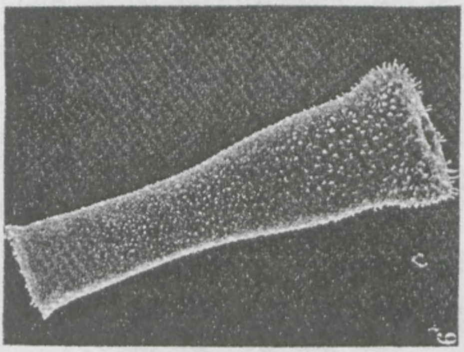
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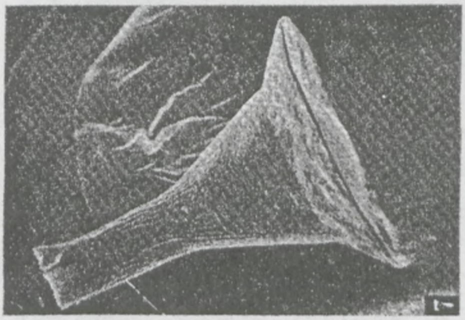
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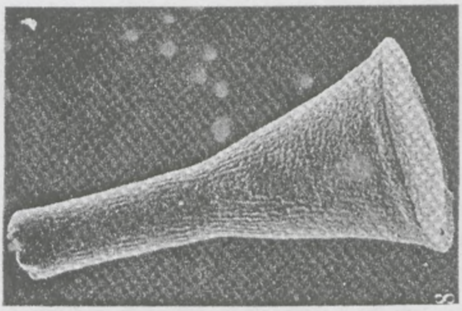
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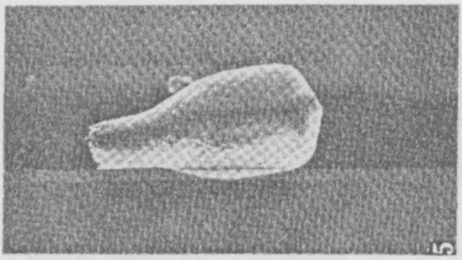
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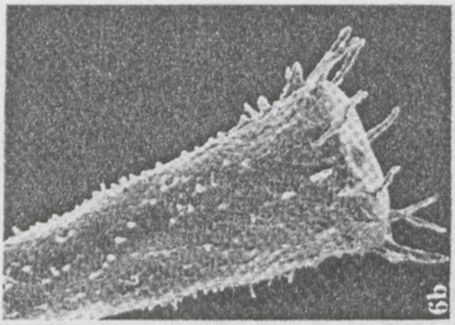
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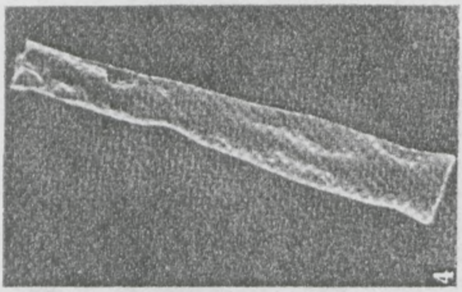
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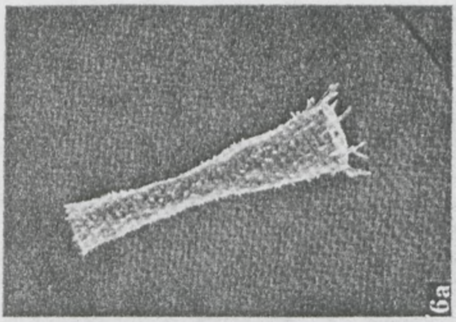
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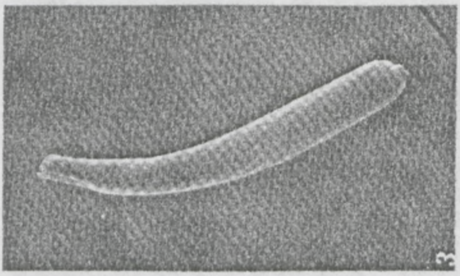
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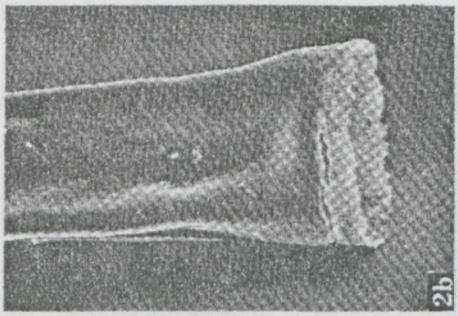
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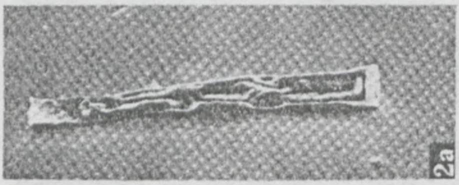
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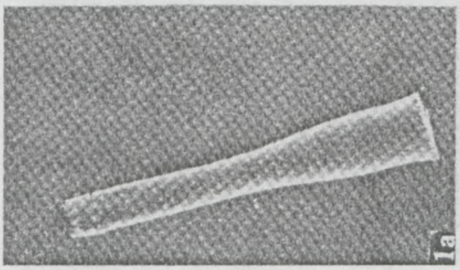
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Paper VI

Nõlvak, J., Meidla T. & Hints L. 1989.
The Taučionys Formation (*Holorhynchus* Beds) in the Iljinskoje boring (Pskov District). *Proceedings of the Academy of Sciences of the Estonian SSR. Geology,* **38,** 88-93.

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THE TAUČIONYS FORMATION (*HOLORHYNCHUS* BEDS) IN THE ILJINSKOJE BORING (PSKOV DISTRICT)

In the South East Baltic the Taučionys Formation, first mentioned as beds with *Holorhynchus* (Пашкевичюс, 1963, 1968; Мянниль, 1966) has been considered to belong to the lowermost part of the Porkuni Regional Stage (Брангулис et al., 1982; Ульст et al., 1982; Решения . . . , 1987) or to the upper part of the Pirgu Regional Stage (Kaljo et al., 1988). So far there have been few special studies on fossils in the *Holorhynchus* beds in the East Baltic. Some brachiopods, corals and echinoderm columnals have been mentioned, mostly in the stratigraphic works.

We report new data on the Upper Ordovician macro- and microfossils in the Iljinskoje boring enabling to specify the stratigraphic position of the Taučionys Formation.

The Iljinskoje borehole, about 90 km south of Pskov, is the northeasternmost section of the Taučionys Formation in the Lithuanian confacies belt of the Baltic basin. The core of the Iljinskoje section has been described by L. Põlma, who divided the Ordovician strata into four parts (beds):

1) 492.0—503.0 m ($F_{I}c$) — light grey medium-bedded argillaceous and organodetritic limestones with dark grey layers of calcareous and argillaceous marls;

2) 479.1—492.0 m ($F_{I}cT\check{c}$) — light grey, slightly brownish, thin- or medium-bedded, partly seminodular cryptocrystalline organodetritic limestones with thin (2—5 cm), rarely thicker (about 10 cm) layers of darkbrown calcareous and argillaceous marls. Limestone-marl contacts sharp. The lower boundary quite distinct;

3) 474.0—479.1 m ($?F_{I}c$) — light grey, slightly brownish seminodular organodetritic limestones. Lower boundary transitional;

4) 469.8—474.0 m ($?F_{II}$) — light grey, partly yellowish seminodular dolomites (above a depth of 471.4 m) and organodetritic limestones. Discontinuity surfaces on the boundaries of different lithotypes.

The distribution of macrofossils in the Taučionys Formation (the 2nd bed) and in the overlying strata will be discussed in some general aspects.

In the lower half of the Taučionys Formation (int. ~485—492.0 m) macrofossils are rare. Besides *Holorhynchus* sp. there are fragments of strophomenids, halysitid *Catenipora tapaensis* Sok. and alga *Vermiporella*. *Ptychopleurella* sp., *Parastrophina*? sp. and *Asteropeltis* sp. (det. by Reet Männil) occur at the lower boundary (Fig. 1). In Norway the last two genera are represented in the *Onniella* association of the Stage 5a (Brenchley, Cocks, 1982).

The fossiliferous upper half of the formation (int. 379.1~485 m) is characterized first of all by *Holorhynchus* and *Vermiporella*. The corals become more frequent. *Holorhynchus* is represented by disarticulated valves and their fragments which are well noticeable on the surface of the core. Almost complete shell of *Holorhynchus giganteus* Kiær has been found at a depth of 480.0 m. *Holorhynchus* is associated with rare strophomenids, dalmanellids (*Isorthis*? sp.), sowerbyellids (*Thaerodonta*? sp.) and atrypids (*Eospirigerina* sp., in the topmost part of the Taučionys Formation).

The low diversity of the macrofossils makes these *Holorhynchus* beds similar to the analogous beds in Norway (Brenchley, Cocks, 1982), but different from the strata with *Holorhynchus* in Sweden (Jaanusson, 1982).

The topmost 9.3 m of the Upper Ordovician section (the 3rd and 4th beds) are characterized by halysitids (*Catenipora tapaensis*, *C.* cf. *tapaensis*), rugose corals and echinoderm columnals (mostly in bed 4). The circular columnals with quadrangular lumen (depth 474 m) are quite similar to columnals in the Röa Member of the Porkuni Regional Stage

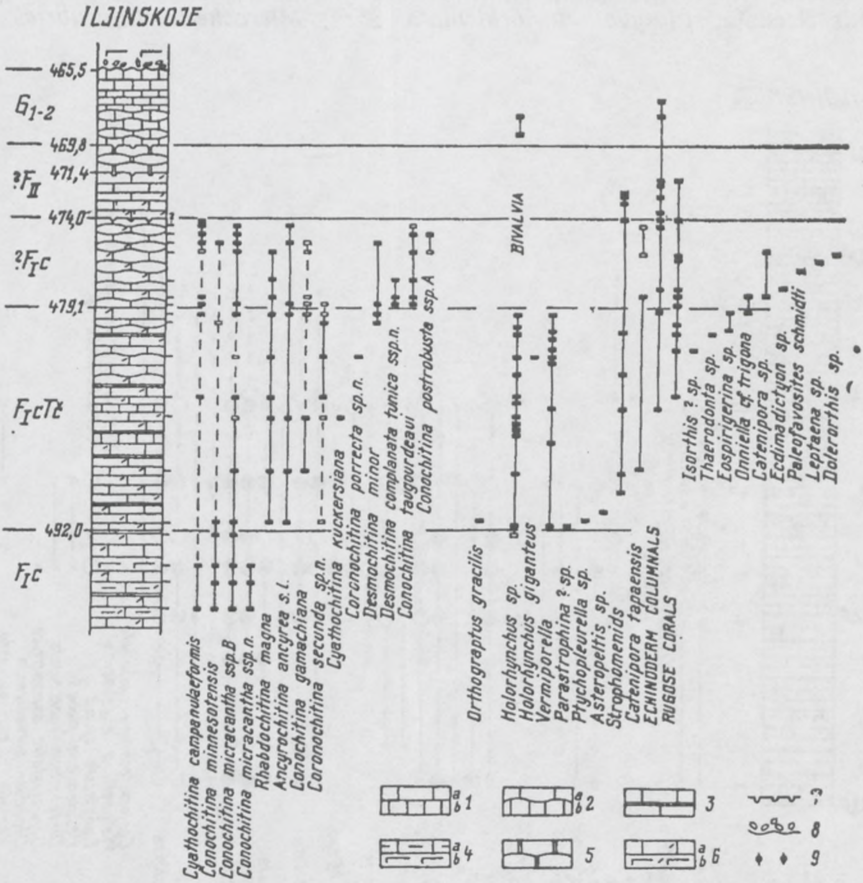

Fig. 1. Distribution of chitinozoans and macrofossils in the Upper Ordovician and Lower Silurian strata of the Iljinskoje core. An empty quadrangle indicates that the identification is conditional (cf.). The core section after L. Põlma. Legend: *1a* — limestone, *1b* — cryptocrystalline (aphanitic) limestone, *2a* — seminodular limestone, *2b* — seminodular cryptocrystalline limestone, *3* — limestone with thin (mostly less than 5 cm) layers of calcareous and argillaceous marls, *4a* — argillaceous limestone, *4b* — calcitic marls, *5* — seminodular dolomite, *6a* — fine organodetritic limestone; *6b* — organodetritic marls; *7* — discontinuity surface, *8* — denudation surface with conglomerate, *9* — silicification. F_{II} — Porkuni Regional Stage; $F_{I}c$ — Pirgu Regional Stage; Tč — Taučionys Formation; G_{1-2} — Juuru Regional Stage (Silurian).

in North Estonia. At the same time *Catenipora tapaensis* has not been found in the strata younger than the Pirgu age. The brachiopods in this part of the section are quite rare. There occurs *Onniella* cf. *trigona* Rubel (Хинтс, 1975) which has a great similarity with the *O. kalvoya* Cocks from some brachiopod associations of Stages 5a and 5b in Norway (Cocks, 1982; Brenchley, Cocks, 1982).

In the Iljinskoje core the Taučionys Formation and the underlying strata (bed 1) contain rich and diverse ostracode fauna (Fig. 2) consisting of more than 50 species. Some species (*Adamczakia holosolenica* Scallreuter, *Easchmidtella* sp. n.) indicate to the faunal connection with the Estonian confacies belt.

The strata below the Taučionys Formation contain an ostracode association represented by two kinds of species: a) ostracodes widely distributed in the Upper Ordovician [*Steusloffina cuneata* (Steusloff), *Medianella blidenensis* (Gailite), *M. intecta* (Stumbur), *Rectella romboformis* Neckaja, *Platybolbina orbiculata* Sarv, *Microcheilinella lubrics*

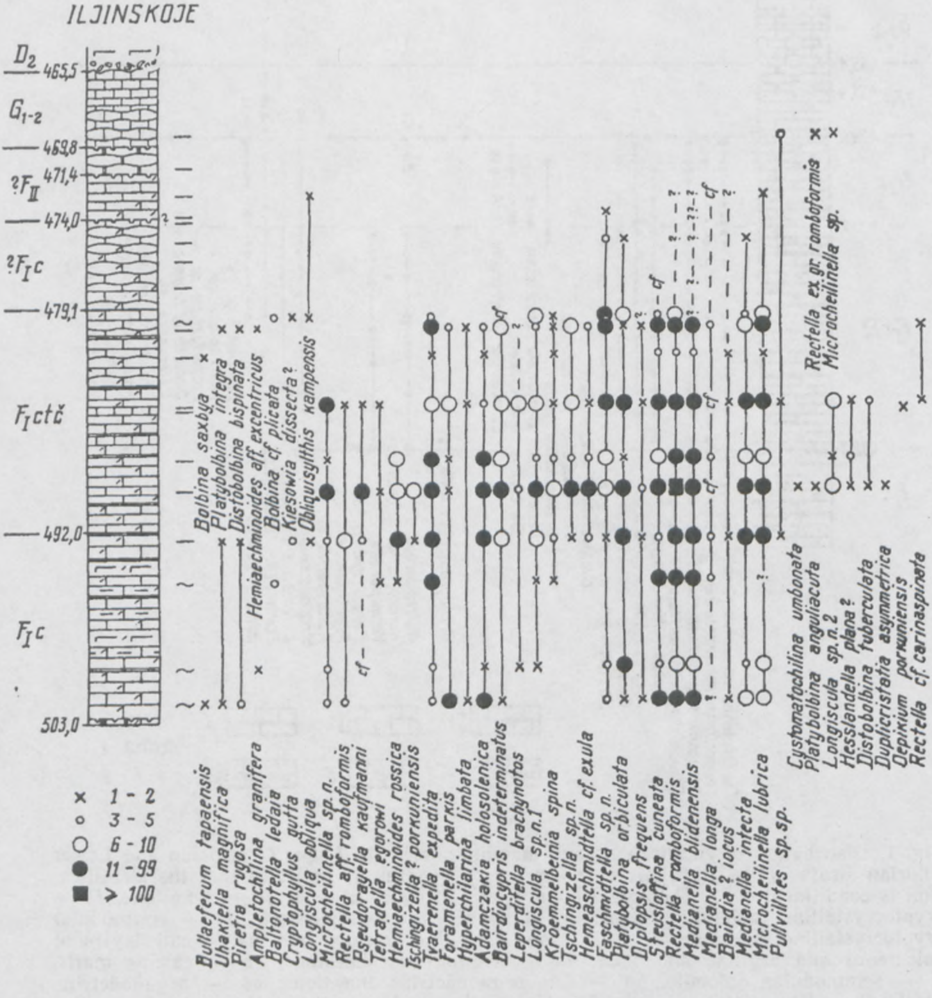

Fig. 2. Distribution of ostracodes in the Upper Ordovician and Lower Silurian strata of the Iljinskoje core.

(Stumbur), *Bairdiocypris indeterminatus* Pranskevicius] and b) a number of species appearing in the Pirgu Regional Stage [*Foramenella parkis* (Neckaja), *Piretia rugosa* (Steusloff), *Adamczakia holosolenica*, *Easchmidtella* sp. n.]. Two species *Baltonotella ledaia* Sidaravičiene and *Adamczakia holosolenica* established in this part of the section have not been known from the Porkuni Regional Stage in the North East Baltic.

The composition of the ostracode fauna in the Taučionys Formation (bed 2) is quite similar to that in the underlying strata (bed 1) undoubtedly of the Pirgu age. Only *Duplicristatia asymmetrica* Scallreuter and *Oepikium porkuniensis* Henningsmoen, represented by a few specimens, are previously known from the rocks of the Porkuni age in North Estonia. At the same time *D. asymmetrica* is established in the Kuldiga Formation of the Porkuni Regional Stage (Сарв, Мейдла, 1984) and *O. porkuniensis* from the Stages 5a and 5b in Norway (Henningsmoen, 1954).

Some differences in the composition and frequency between the lower and upper part of the Taučionys Formation can be outlined. For example, in the upper part of the formation *Hemiaechminoides rossica* Neckaja and *Microheilinella* sp. n. have not been found and *Adamczakia holosolenica* is quite rare in comparison with the lower part..

Above the upper boundary of the Taučionys Formation (at a depth of 479.1 m) the Ordovician ostracodes are identified up to a depth of 472.7 m. Among them we have no species diagnostical for the regional stages, but certainly there are no affinities with the ostracode fauna of the Kuldiga Formation (see Гайлите, 1970). The presumably Silurian forms appear at a depth of 469.3 m.

Chitinozoan fauna is relatively scarce in the lowermost beds below a depth of 492.0 m. Such low diversity assemblage is characteristic to the middle part of the Pirgu Regional Stage. In Lithuania it occurs in the Ukmerge Formation (underlying typical Taučionis beds) and in the lowermost part of the Halliku (in East Estonia) or Adila Formation (in North-West Estonia). *Conochitina micracantha* ssp. n. seems to be a useful biostratigraphic tool (Fig. 1) .

In the Taučionys Formation of the Iljinskoje core the assemblage of chitinozoans resembles that in Lithuania (Нылвак, 1988). The taxonomic composition differs conspicuously from that in the underlying beds (below 492.0 m). Among others, some specific taxa with restricted stratigraphic ranges can be discussed, especially *Conochitina gamachiana* Achab and *Coronochitina secunda* sp. n. The latter is morphologically transitional between *Coronochitina coronata* Eisenack (widely distributed in the middle Pirgu) and *"Conochitina" taugourdeaui* Eisenack. It is based on a gradual change in the nature of ornamentation by which these taxa are distinguished and could be interpreted as evolutionary lineages (see also Melchin, Legault, 1985). It is proposed here that *Conochitina gamachiana* evolved into *C. postrobusta* ssp. a (typical to Kuldiga Formation with *Hirnantian* macrofauna in Latvia) through an intermediate form (*C.* cf. *gamachiana*) near the upper boundary of these beds (Fig. 1).

Among graptolites *Orthograptus gracilis* (Roemer) is distinguished. This species has been also found in the Lithuanian cores from the same unit and it seems to indicate the graptolite zone of *Dicellograptus anceps* (see Skoglund, 1963; Мянниль, 1976).

The chitinozoan assemblage with *"C." taugourdeaui* in the 3rd part is similar in many respect to that of the probably contemporaneous depositis in North-East Lithuania (e. g. in the Butkunai core). These beds were considered comparable with the Kuldiga Formation (Пашкевичюс, 1973; Лашков et al., 1984). No far-reaching conclusion can be drawn on a single section, but it is apparent, however, that approximately at a depth of 479.1 m, there are transitional changes in lithology (see above) and also in the distribution of chitinozoans.

The topmost Ordovician deposits (int. 474.0—469.8 m) are lacking all kinds of organic-walled microfossils. The absence of chitinozoans is more likely due to the lack of suitable sediments for preservation (probably deposited in a high energy environment, and also the dolomitization of the rocks), as it is in the lowermost Porkuni (Rõa Member) in North Estonia.

In the Iljinskoje core the Upper Ordovician macro- and microfossils studied allow to presume the Pirgu age of the Taučionys Formation, which is characterized by the occurrence of the *Holorhynchus*. These *Holorhynchus* beds in the East Baltic are comparable with the analogous beds in Norway, but for their stratigraphic relationship we do not have an unambiguous settlement as yet. In our section the Ordovician strata succeeding the Taučionys Formation are characterized by the transitional fauna between the faunas of the Pirgu and Porkuni age, mostly in the North East Baltic.

The authors are indebted to E. Sammet (Sevzapgeologia, Leningrad) for the possibility to study the core of the borehole. We express gratitude to our colleague L. Sarv for his help in collecting the samples and studying ostracodes, and to H. Nestor for the identification of corals. Thanks are due to D. Kaljo for the critical reading of the manuscript.

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TAUČIONYSE KIHISTU (*HOLORHYNCHUS*'E KIHID) ILJINSKOJE PUURSÜDAMIKUS (PIHKVA OBLAST)

Esitatakse Taučionyse kihistu (*Holorhynchus*'e kihtide), selle lasumi ja lamami paleontoloogiline iseloomustus Iljinskoje läbilõikes (Pihkva oblasti kaguosa). Taučionyse kihistu ostrakoodide ja kitinozoade kompleks, millega kaasneb spetsiifiline makrofossiilide assotsiatsioon, on seotud pirguealise faunaga. Taučionyse kihistul lasuvates ordoviitsiumiealistes kihtides on kindlaks tehtud Pirgu ja Porkuni lademe piirikihtide fauna.

Я. НЫЛВАК, Т. МЕЙДЛА, Линда ХИНТС

ТАУЧЕНСКАЯ СВИТА (СЛОИ С *HOLORHYNCHUS*) В РАЗРЕЗЕ СКВАЖИНЫ ИЛЬИНСКОЕ (ПСКОВСКАЯ ОБЛАСТЬ)

Установленный в таученской свите комплекс остракод и хитинозой, который сопровождается специфической ассоциацией макрофоссилий, связан с фауной пиргуского возраста. Установлена фауна пограничных слоев пиргуского и поркуниского горизонтов в надстилающих таученскую свиту ордовикских отложениях.

Paper VII

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CHITINOZOAN BIOSTRATIGRAPHY OF THE MIDDLE ORDOVICIAN DALBY LIMESTONE IN THE FJÄCKA SECTION, SILJAN DISTRICT, SWEDEN

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IGCP Project 410
The Great Ordovician
Biodiversification Event

Abstract. The main section at Fjäcka, Dalarna, is a classic Ordovician locality in Sweden. It is the type section of three North Atlantic conodont zones and subzones, three Baltoscandian chitinozoan zones, and of several topoformations. Detailed analysis of chitinozoan distribution in the Middle Ordovician Upper Dalby Limestone showed variable thickness of chitinozoan biozones caused by block movements, as well as clear variations in the content of acid-resistant organic-walled fossils in different blocks. Nevertheless, the boundary of the East Baltic Kukruse and Haljala stages was established in one part of the section, and a composite section was worked out in spite of some complications caused by tectonic dislocation of beds.

Key words: Middle Ordovician, Sweden, Dalby Limestone, biostratigraphy, chitinozoans.

INTRODUCTION

The initial purpose of the present study was to ascertain whether any traces of the Lockne impact in Jämtland (Lindström & Sturkell, 1992; Lindström et al., 1996; Grahn et al., 1996; Grahn, 1997) existed in the contemporaneous sedimentary sequence of the Siljan district, some 200 km to the south. The appropriate interval in the main section at Fjäcka (localities 7 and 8 in Jaanusson, 1947, fig. 1; 1963, fig. 11) was closely examined macrolithologically as well as with respect to the succession of acid-resistant organic-walled microfossils, mainly chitinozoans.

No indications of the Lockne impact were found, which were expected to occur in the lithology of beds at locality 8. Chitinozoans, however, yielded new biostratigraphic information, and thus the chitinozoan biostratigraphy is the main subject of the present contribution.

The main section at Fjäcka is a classic Ordovician locality of Sweden and one of the best studied sections of the Central Baltoscandian Confacies. It is situated along the Moldå Stream at Fjäcka in Dalby village, Siljan District, Dalarna. Törnquist (1867) first described the locality as a group of isolated exposures including a small, now abandoned quarry. During 1945–47 the section was excavated, and a continuous outcrop from the upper Furudal Limestone to the Jonstorp Formation was exposed (Jaanusson, 1982, p. 36). Fossils and lithology have been described in numerous papers (Jaanusson, 1947, 1962, 1963, 1976, 1982; Jaanusson & Martna, 1948; Martna, 1955; Laufeld, 1967; Bergström, 1971a, b; Holmer, 1989). In 1976 further excavations were carried out and the section was designated as a nature reserve. Bergström (1971a, b) defined the main section at Fjäcka as the type locality of three North Atlantic conodont zones: *Pygodus anserinus*, *Amorphognathus tvaerensis* (and its three subzones), and *A. superbus*. Nõlvak & Grahn (1993) used it as a reference section for three chitinozoan zones: *Lagenochitina dalbyensis*, *Belonechitina hirsuta*, and *Spinachitina cervicornis*. It is also the type locality of the Dalby, Moldå, and Fjäcka formations (Jaanusson, 1982).

MATERIAL

The main series of samples from locality 8 is from the steep, some metres high bank. It was collected by E. F. F. Sturkell in 1995 along subsections 305 and 306 (Fig. 1). Five samples were provided by H. Bauert, collected during the WOGOGOB excursion in 1990, and additional samples were secured by J. Nõlvak in 1995. The sample intervals were kept particularly small in the levels where the chitinozoan zones are known to have a limited vertical range in other regions (see Nõlvak & Grahn, 1993). The size of the samples varied between 200 and 500 g. All material is deposited in the Institute of Geology, Tallinn.

In sampling the Dalby Limestone sequence at locality 8, the lowermost K-bentonite bed (about 9.5 m below the base of the complex of K-bentonite beds according to Jaanusson, 1963, fig. 9; or 11 m below the Kinnekulle bed after Bergström et al., 1995, fig. 2) was used as an index horizon (level B-1, Figs.1, 2).

Another locality (No. 7 in Jaanusson, 1947, fig. 1; and along subsections 13 and 14, see Figs. 1, 3 herein) was sampled by J. Nõlvak in 1992, just adjacent to

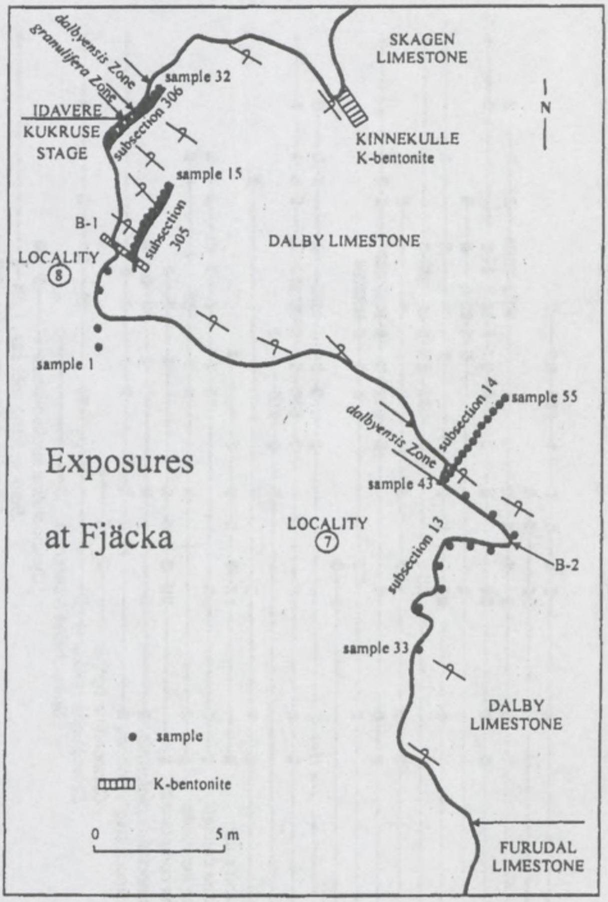

Fig. 1. Sketch-map of the southern part of the exposures at Fjäcka. Localities 7 and 8 according to Jaanusson (1947, fig. 1). The thick line outlines the area of small terraces or acclivities (locality 8) or the edge of the old quarry (locality 7).

the northeastern corner of the abandoned old quarry. In this area the bedrock probably became first exposed in 1976 (V. Jaanusson, pers. comm. 1997). The most argillaceous bed (level B-2, Figs. 1, 3) was used as an index horizon. The beds (subsection 14) with chitinozoans indicating the *L. dalbyensis* Zone rest on limestones (subsection 13, samples 33–42) barren of acid-resistant organic-walled macro- and microfossils, such as acritarchs, scolecodonts, graptoloids, dendroids, organic-walled foraminifera, hydroids, and chitinozoans. The lack of chitinozoans was already noticed by Laufeld (1967, p. 292). Nevertheless, the absence of chitinozoans and the other fossils has no regional significance, since at locality 8, only 15 m to the northwest, the lowermost beds contain all mentioned groups of fossils (see samples 1–4 and higher, Figs. 1, 2). Lithologically these

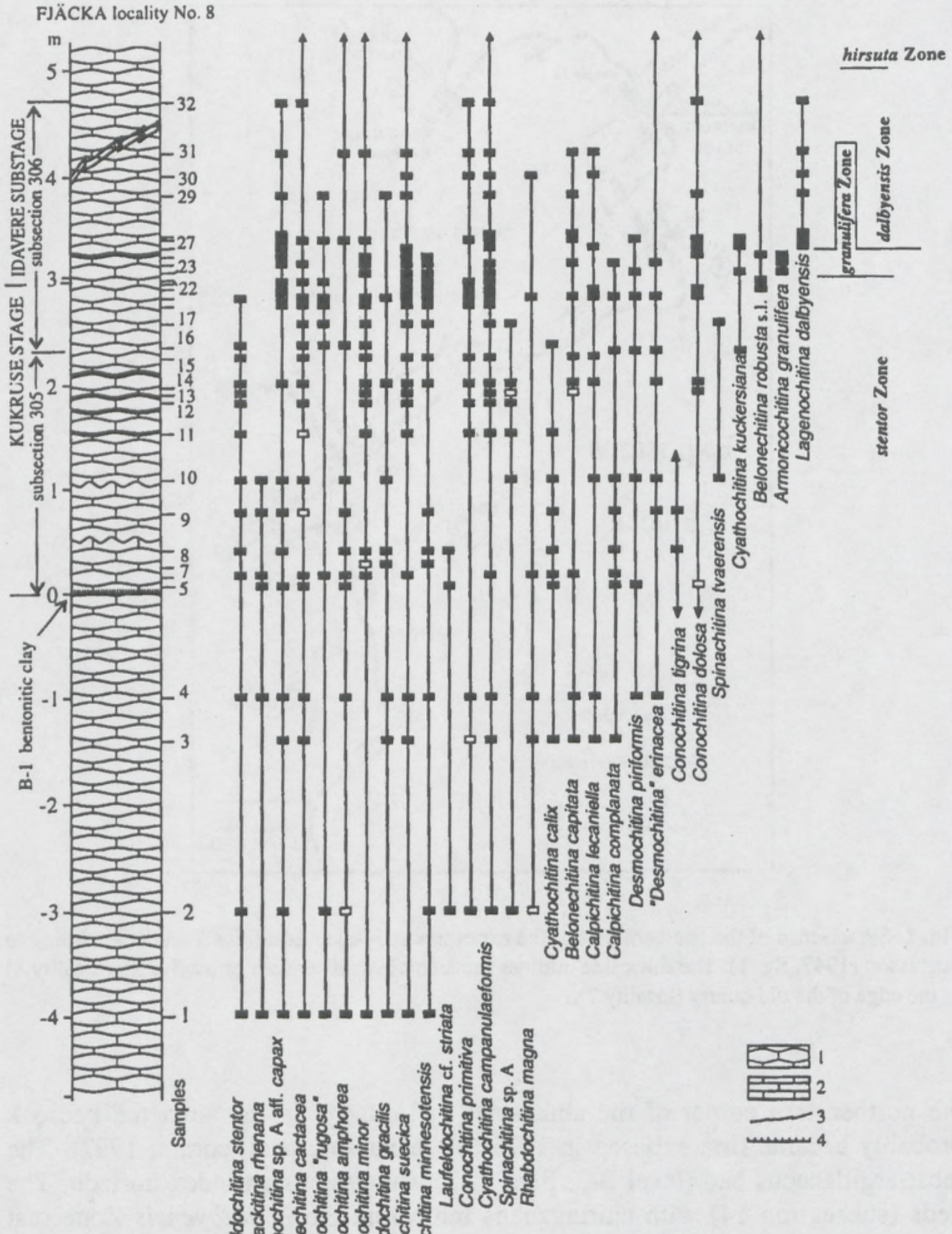

Fig. 2. Faunal log showing the range of chitinozoans in the upper part of the Dalby Limestone (Upper Member in Jaanusson, 1982, p. 38) at Fjäcka locality 8. 1, seminodular limestone (parallel lines mark clayey intercalations); 2, bedded limestone; 3, fault; 4, K-bentonite bed. Open squares represent forms referred to as cf. The lower boundary of the *hirsuta* Zone, about 5 m above the bentonitic clay layer, is shown according to the data from sample 60-166 in Laufeld (1967).

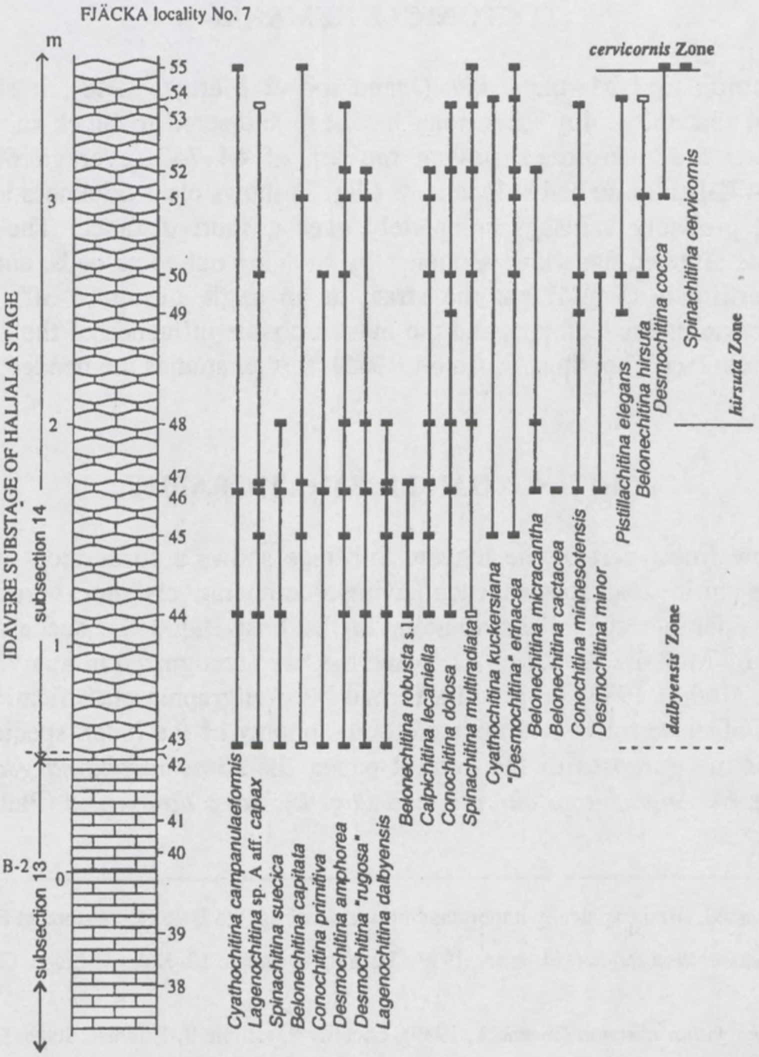

Fig. 3. Faunal log showing the range of chitinozoans in the middle part of the Dalby Limestone at Fjäcka locality 7. Barren samples 33–37 (see Fig. 1) are not shown. The dashed line marks the appearance level of chitinozoan fauna. For legend refer to Fig. 2.

fossiliferous rocks are grey, somewhat nodular limestones with irregular argillaceous intercalations. However, barren limestones in the limits of subsection 13 are clearly more greenish (more weathered?) and less nodular, although in general the same as described by Jaanusson (1982, p. 39).

Subsection 14 continues upwards into the *B. hirsuta* Zone (Plate, fig. 9) and the basal *S. cervicornis* Zone (Fig. 3). The position of these zones is abnormally low and indicates some tectonic dislocation.

TECTONICAL REMARKS

In addition to J. Martna's data (Jaanusson & Martna, 1948), it should be mentioned that the Dalby Limestone has been subjected to block movements. These beds are overturned, having the dip of 64–74° (average 68°). The lowermost K-bentonite bed at locality 8 (Fig. 1) shows clear thickness variations and most probably vanishes completely over a short distance. The tectonic movements affected the whole sequence by wedging out some beds, and several small overthrusts (Fig. 4) cut the strata at an angle of up to 30°. For the reconstruction of the tectonics and the most probable influence of the Devonian Siljan impact (see Thorslund & Auton, 1975), further studies are needed.

CHITINOZOAN BIOSTRATIGRAPHY

The lowermost part of the Idavere Substage shows a succession of several distinctive chitinozoan species, each having a comparatively short vertical range but a considerable spatial distribution. In the East Baltic the succession was described by Männil (1986, fig. 2.1.1) and has been recognized in many sections. Nõlvak & Grahn (1993) proposed a formal biostratigraphic classification based on ranges of these forms and updated the taxonomy of the index species of the zones. For the purposes of the present paper, the *Armoricochitina granulifera* (Plate, fig. 6), *Angochitina curvata*, and *Lagenochitina dalbyensis* (Plate, fig. 8)

PLATE. Selected, stratigraphically important chitinozoans from the Dalby Limestone at Fjäcka.

Fig. 1. *Laufeldochitina stentor* (Eisenack, 1937). Locality 8, sample 10, Kukruse Stage. Ch 2075/10, SEM × 140.

Fig. 2. *Eisenackitina rhenana* (Eisenack, 1939). Locality 8, sample 9, Kukruse Stage. Ch 2242/9, SEM × 515.

Figs. 3, 4. *Conochitina tigrina* Laufeld, 1967. Locality 8, sample 8, Kukruse Stage. Ch 2210/8, SEM × 100, × 405.

Figs. 5, 10. *Spinachitina tvaerensis* Grahn, Nõlvak & Paris, 1996. Locality 8, sample 10, Kukruse Stage. Ch 2083/10, SEM × 300, × 960.

Fig. 6. *Armoricochitina granulifera* Nõlvak & Grahn, 1993. Locality 8, sample 24, Idavere Substage. Ch 2257/24, SEM × 255.

Fig. 7. *Lagenochitina* sp. A aff. *capax*. Locality 7, sample 44, Idavere Substage. Ch 2229/44, SEM × 650.

Fig. 8. *Lagenochitina dalbyensis* (Laufeld, 1967). Locality 7, sample 43, Idavere Substage. Ch 2127/43, SEM × 400.

Fig. 9. *Belonechitina hirsuta* (Laufeld, 1967). Locality 7, sample 51, Idavere Substage. Ch 2113/51, SEM × 480.


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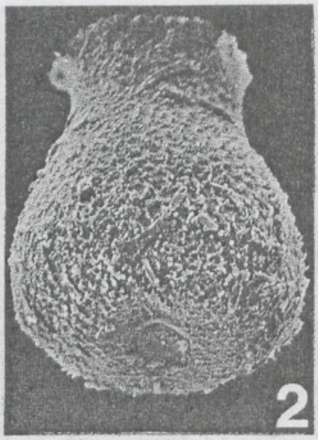
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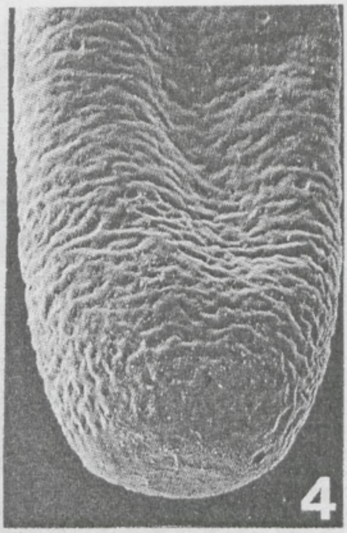
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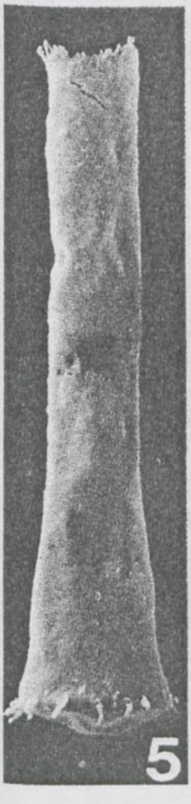
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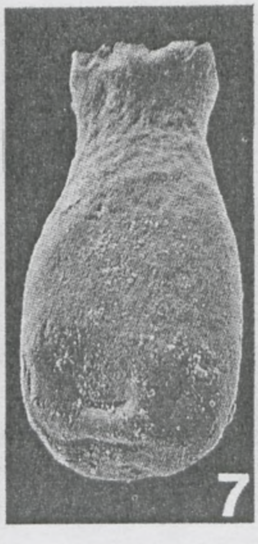
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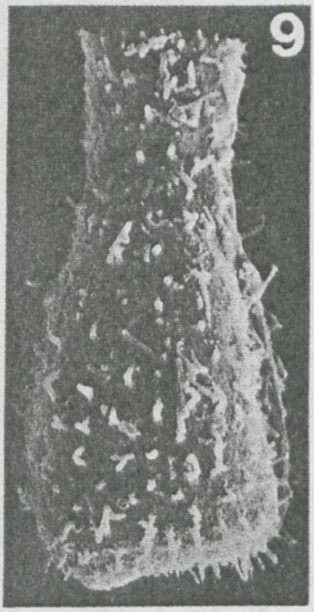
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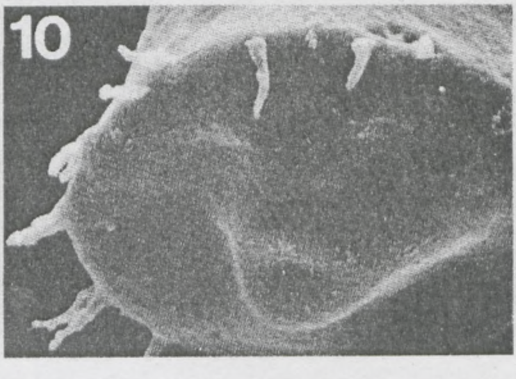
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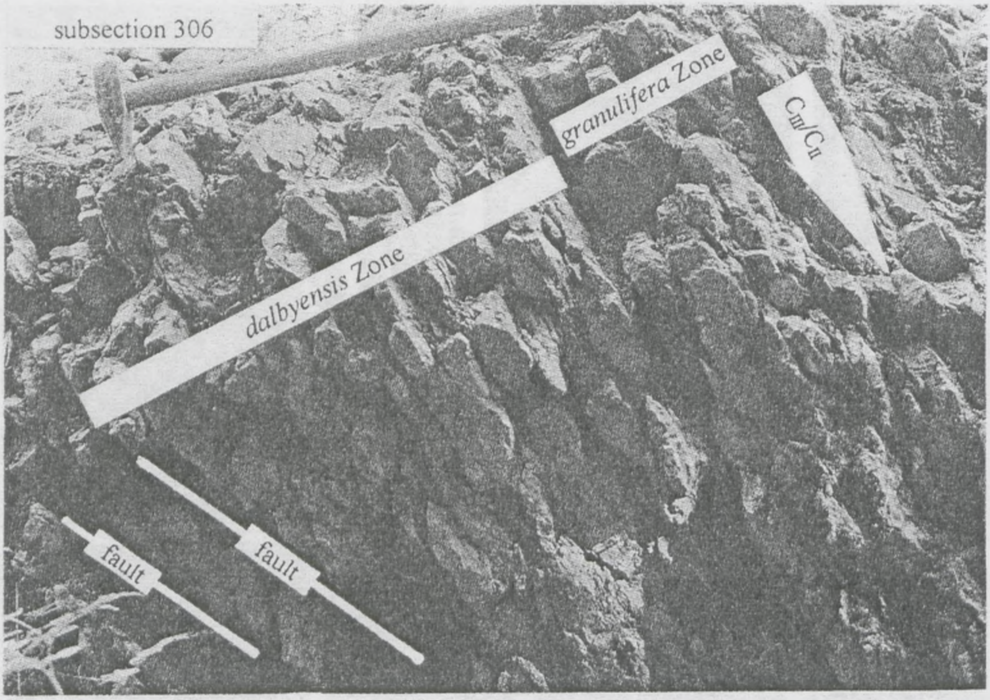

Fig. 4. Overthrusts and thickness of layers with chitinozoan biozones in the lowermost Idavere Substage in subsection 306 (see Fig. 1). C_{II}, Kukruse Stage; C_{III}, Idavere Substage of the Haljala Stage. Note the differences in the thickness of the *dalbyensis* Zone caused by faults.

zones are of particular interest. It should be noted that the base of the Idavere Substage was proposed to specify as that of the *A. granulifera* Zone (Hints et al., 1995).

Previous information on chitinozoans from the main section at Fjäcka is confined to Laufeld (1967), which can be regarded as a pioneer work on the biostratigraphy of Ordovician chitinozoans. There, the *L. dalbyensis* Zone is clearly distinguishable in the faunal log, but the beds containing *A. granulifera* fall within the interval between two successive samples analysed (60-170 and 60-174 in Laufeld, 1967, fig. 6).

Unfortunately, the sample size of 50 g has proved to be too small in the whole Baltoscandia to yield sufficient data on Ordovician chitinozoans. Our study revealed firm presence of the *A. granulifera* Zone (Figs. 2, 4) with a thickness not exceeding 15 cm. The recognition of this chitinozoan zone is of great importance for a precise correlation between the sequences of the North Estonian and Lithuanian Confacies and that of the Siljan district. In the main section at Fjäcka, the level of the base of the Haljala Stage (and Idavere Substage) is situated 2.9 m above the K-bentonite bed (B-1 in Fig. 2) in subsection 306.

Angochitina curvata Nõlvak & Grahn has not been encountered at Fjäcka. The thickness of the corresponding zone in North Estonian sections is often only a few centimetres, so the bed containing the zonal fossils may simply have been missed in the Fjäcka section.

An additional detail which may have biostratigraphic significance is the occurrence of abundant acritarchs *Leiosphaeridia* sp. in a thin bed just below the *A. granulifera* Zone (sample 22, Figs. 2, 4) and in higher levels at Fjäcka. Such a mass occurrence of acritarchs is widespread at the same level also in North Estonian sections, being a useful, and not only a local indication of the lower boundary of the Idavere Substage (Nõlvak, 1972, fig. 2).

The chitinozoan succession of the upper Dalby Limestone below the *A. granulifera* Zone agrees with that in many Estonian and Swedish sections (Grahn et al., 1996). The highest level of *Laufeldochitina stentor* (Eisenack) (Plate, fig. 1) is just below the base of the *A. granulifera* Zone. *Eisenackitina rhenana* (Eisenack) (Plate, fig. 2), which is, most probably, not separated from *Belonechitina cactacea* in samples 60-184, 60-182, 60-178, 60-174 by Laufeld (1967, fig. 6), disappears somewhat lower down. It is also interesting to note that two distinctive species, *Conochitina tigrina* Laufeld (Plate, figs. 3, 4) and *Spinachitina tvaerensis* Grahn, Nõlvak & Paris (Plate, figs. 5, 10; the latter species was described from the Tvären impact crater by Grahn et al., 1996, fig. 6), which are characteristic of the upper part of the *Laufeldochitina stentor* Zone, occur at Fjäcka roughly in the same interval. *Lagenochitina* sp. A aff. *capax* (Plate, fig. 7) has the same vertical range as in East Baltic sections (Grahn et al., 1996, p. 34).

The material from Fjäcka includes also several new chitinozoan species, but their description is postponed to a separate paper.

To sum up, we may note that the succession of comparatively short-range chitinozoans in the main section at Fjäcka is the same as in the earlier studied East Baltic Ordovician sections (Nõlvak & Grahn, 1993, fig. 6). Bed-by-bed sampling of lithologically variable rocks has revealed the presence of some very short-range chitinozoan zones.

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KITIINIKUTE BIOSTRATIGRAAFIA KESKORDOVIITSIUMI FJÄCKA LÄBILÕIKE DALBY LUBJAKIVIDES SILJANI PIIRKONNAS ROOTSIS

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On esitatud andmed kitiinikute leviku kohta Rootsi Siljani piirkonna Fjäcka tüüpläbilõike keskordoviitsiumi kihtides. Uuritud läbilõikest on defineeritud mitmeid Põhja-Atlandi provintsi konodontide biotsoone ja Baltoskandia kitiinikute biotsoone, mis võimaldavad suhteliselt täpseid korrelatsioone ordoviitsiumibasseini osade vahel. Detailsete prooviseeriatega tehti kindlaks Idavere alamlademe varem vaid Baltikumi läbilõigetes määratud alumine piir, mis on põhjendatud 15 cm paksuse *Armoricochitina granulifera* biotsooni levikuga. Raskused ilmnesid Fjäcka läbilõikes levivate fossiilideta kihtide vanuselise järjestuse määramisel paljandi eri osades, sest kihtide lasumus on ilmselt mõjutatud devoniaegse Siljani impaktkraatri tekkeprotsessidest.

БИОСТРАТИГРАФИЯ ХИТИНОЗОЙ В СРЕДНЕОРДОВИКСКИХ ИЗВЕСТНЯКАХ ДАЛБЮ В РАЗРЕЗЕ ФЯКА, РАЙОН СИЛЬЯН, ШВЕЦИЯ

Яак НЫЛВАК, Ингве ГРАН и Эрик Ф. Ф. СТУРКЕЛЛ

Приведены подробные данные о распределении хитинозой в стратотипических известняках Далбю в разрезе Фяка. В этом разрезе обнаружен ряд стратиграфически важных зон конодонтов для Северо-Атлантической провинции и хитинозой для Балтоскандии, по наличию которых можно сравнительно точно сопоставить между собой разрезы в разных регионах ордовикского Балтийского бассейна. По распространению в прибалтийских разрезах узкой (мощностью 15 см) биозоны *Armoricochitina granulifera* установлен уровень нижней границы идавереского подгоризонта. Выяснено, что затруднения при определении возрастных соотношений “немых” слоев в разных частях обнажения обусловлены нарушением их залегания в процессе образования девонского метеоритного кратера в районе Сильяна.

Paper VIII

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MORE ABOUT ORDOVICIAN MICROFOSSIL DIVERSITY PATTERNS IN THE RAPLA SECTION, NORTHERN ESTONIA

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Abstract. Diversity dynamics of acritarchs and chitinozoans was studied in the Rapla borehole section which embraces the Ordovician from the upper Arenig to the topmost Ashgill. The number of taxa per stratigraphic unit, the rate and total rate of appearing and disappearing taxa per unit or per 1 Ma, and diversity changes at the unit boundaries were analysed. Both groups demonstrate an energetic radiation–origination period in the Arenig–Llanvirn, a remarkable extinction event in the late Caradoc (top of the Keila Stage), and a mass extinction in the terminal Ordovician (Pirgu–Porkuni stages). Other diversity changes seem to be local and ecologically controlled. The general pattern of microfossil diversity is similar to that established by Sepkoski (*Ordovician Odyssey*, 1995) for some invertebrates.

Key words: Ordovician, acritarchs, chitinozoans, diversity, extinction, radiation.

INTRODUCTION

In the last decade different events in the Earth's history have aroused wide interest in the world's scientific community. As a result, several IGCP projects have been organized (e.g. 216 "Global Bioevents" and 335 "Biotic recoveries from mass extinctions") to promote detailed study of the most important events, the one occurring at the end of the Ordovician included (Walliser, 1986). In order to understand better the causes and the mechanisms of bioevents, various background studies are needed, especially those describing temporal changes in the taxonomic diversity of different groups of organisms. Diversity dynamics discloses much about evolutionary and environmental biotic changes (radiation, origination, extinction, etc.), provided that adequate methods of description are used.

The objects of this study were acritarchs and chitinozoans, two groups of organic-walled microfossils of obscure origin. The representatives of the first group have very small (diameter 5–150 μm), mostly pilose sphaerical vesicles, those of the second group are much larger (0.5 mm in size) and of different shape (barrel-, bell-, club-shaped, etc.). Such morphological features obviously affected the floating potential of these organisms as planktic particles and consequently some differences in the distribution of the two groups were anticipated but not unambiguously understood.

The main conclusions of our analysis were presented in short at the International Ordovician Symposium in Las Vegas in 1995 (Kaljo et al., 1995). Here we publish a full account of the results obtained, especially the part of the geological background, sampling and analytical methods allowing better understanding of the discussion.

GEOLOGICAL SETTING

The data set analysed was derived from a single borehole section at Rapla. Therefore the study area is geographically very limited, but it covers a considerable part of the Ordovician sequence from the upper Arenig to the topmost Ashgill, in terms of the East Baltic stratigraphic classification from the Volkhov to Porkuni stages. Correlation of this classification with British series is given according to Männil (1990).

The Rapla borehole is situated 60 km south of Tallinn (Fig. 1), within the North Estonian Confacies Belt (Jaanusson, 1976). With regard to the lithological characteristics and subdivision of the section, Põlma (1972) is followed. Only the boundaries of the Uhaku Stage and the lower boundary of the Pirgu Stage have been revised in the light of recent biostratigraphical data (Fig. 2). For the purpose of this study five stages were subdivided into two parts based mainly on the stratigraphic units, currently in use, or lithology, but also for practical reasons. For instance, the Vasavere Formation, which is very thin in the core (150.5–151.5 m), was included into the Jõhvi Stage, the boundary of the Idavere and Jõhvi stages, however, was drawn tentatively at a depth of 151.5 m because a few samples were examined from the Vasavere beds (one for acritarchs and three for chitinozoans).

In the section studied 20 working units were distinguished for the analysis of palaeontological data. Their duration was established by the application of the ages of the main Ordovician boundaries defined by Tucker et al. (1990). The stages within a series were tentatively regarded as being of the same duration. Only in some cases the extent of the stages was somewhat differentiated according to the completeness and thickness of the unit. In the Caradoc part also some gaps were considered.

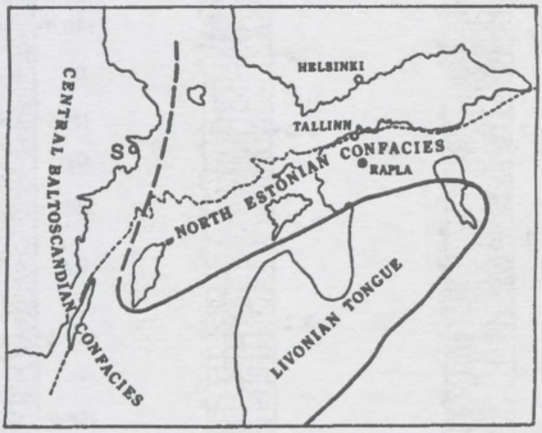

Fig. 1. Location of the Rapla borehole with respect to the Baltic Ordovician confacies belts. S, Stockholm. Thin dashed line – northern limit of the present distribution of the rocks.

A detailed lithological description of the Rapla core section is presented in Fig. 2 compiled according to Põlma (1972). His terminology is given in the legend to the figure. Considering also his data on the content and type of skeletal particles (debris) occurring in the limestone, we can emphasize several more general features and trends in the lithology of the section. For this purpose Dunham's terminology was applied (correlation of both sets of terms was discussed in Nestor, 1990).

In the Arenig and Llanvirn part of the section (Volkhov–Lasnamägi stages) skeletal packstones prevail over wackestones, and oolite intercalations occur. Among skeletal particles trilobites slightly predominate over echinoderms, there is much less brachiopod debris. The Llandeilo and Caradoc (up to the Oandu Stage) are represented by laminar, nodular, and argillaceous wackestones. Trilobites and echinoderms form the bulk of the skeletal debris; in the Idavere–Keila stages also bryozoans are remarkable.

In the upper Caradoc, especially in the Rakvere Stage, calcilutites are most characteristic; skeletal wackestones occur in the lower Nabala Stage. A marked change takes place in the composition of skeletal particles. Algae dominate from the base of the Rakvere Stage to the top of the section. They are less numerous in the Porkuni Stage where echinoderms predominate.

The Ashgill begins with the upper Nabala calcilutites, but higher (Vormsi–Porkuni stages) it is mainly represented by nodular argillaceous

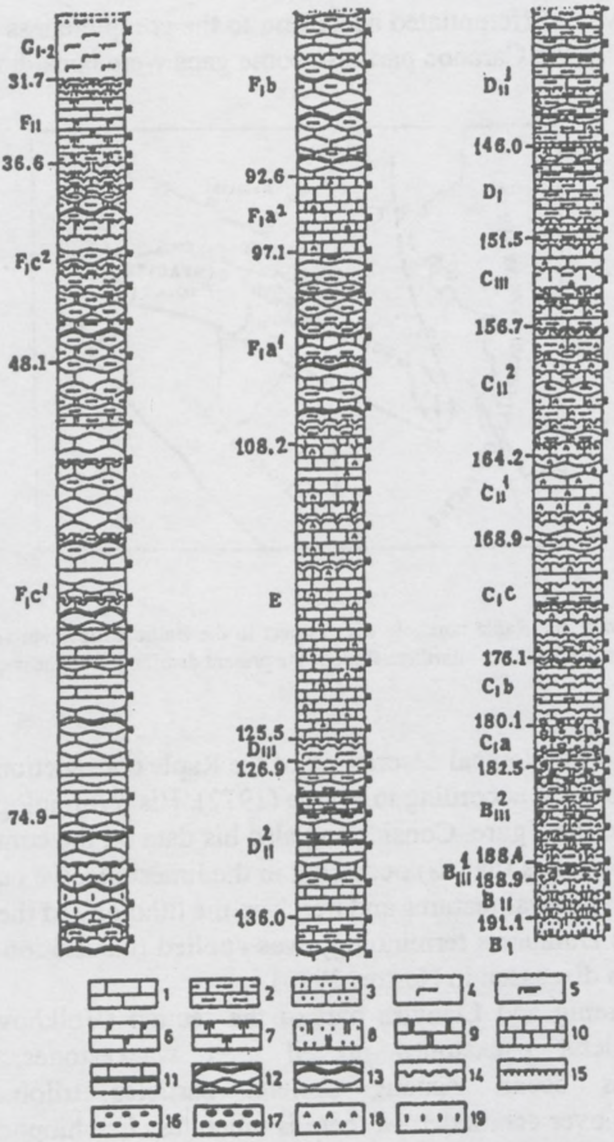

Fig. 2. Lithology and stratigraphy of the Rapla core. Legend: *1*, limestone; *2*, argillaceous limestone; *3*, silty limestone; *4*, calcareous marl; *5*, argillaceous marl; *6*, limestone with dolomite; *7*, dolomitic limestone; *8*, dolomitic siltstone; *9*, dolomite; *10*, cryptocrystalline and microcrystalline limestone; *11*, limestone interbedding with marls; *12*, seminodular limestone interbedding with marls; *13*, nodular limestone interbedding with marls; *14*, discontinuity surfaces; *15*, K-bentonite; *16*, goethite ooids; *17*, calcareous and phosphate ooids; *18*, kukersite (kerogen); *19*, glauconite. Black quadrangles right of the columns denote acritarch sampling places. Units (regional stages, substages): B_I–Latorp; B_{II}–Volkhov; $B_{III}{}^1$, $B_{III}{}^2$–lower, upper Kunda; C_Ia–Aseri; C_Ib–Lasnamägi; C_Ic–Uhaku; $C_{II}{}^1$, $C_{II}{}^2$–lower, upper Kukruse; C_{III}–Idavere; D_I–Jõhvi; $D_{II}{}^1$, $D_{II}{}^2$–lower, upper Keila; D_{III}–Oandu; E–Rakvere; F_Ia^1, F_Ia^2–lower, upper Nabala; F_Ib–Vormsi; F_Ic^1, F_Ic^2–lower, upper Pirgu; F_{II}–Porkuni; $G_{1\text{-}2}$–Juuru (Silurian).

wackestones with a few packstone interbeds in the upper part (upper Pirgu and Porkuni stages; Fig. 2).

The clay content is lowest from the base of the section up to the top of the Llandeilo (lower Kukruse Stage). It is highest (close to 25%) in the lower and partly upper Caradoc (upper Kukruse–Oandu stages), and varies from a few per cent to more than 25% higher in the section.

Most of these rocks formed in open shelf or middle ramp settings, below fair-weather wave base. The high content of skeletal debris in the Volkhov–Lasnamägi packstones, the presence of oolites and many hardground levels etc. are evidence of a shallower environment than that of the wackestones of the Llandeilo, especially of the lower Caradoc.

As usually interpreted, the upper Caradoc–lowest Ashgill (Rakvere–Nabala stages) with calcilutites records the tectonically induced weak influx of terrigenous material into the basin (Põlma, 1972; Hints et al., 1989). In the upper part of the Ashgill (upper Pirgu–Porkuni stages) packstones with abundant echinoderm debris indicate a general shallowing of the sea.

The pattern of sedimentary features more or less coincides with the British series and with the subdivided Caradoc. The varying content of terrigenous material suggests a more detailed cyclicity, especially in the upper Caradoc and Ashgill, in general following the stratigraphic units established for the present study. All this forms a changing environmental background for the acritarch and chitinozoan diversity dynamics discussed in this paper. Many discontinuity surfaces (hardgrounds) recorded in the Rapla section (Põlma, 1972) indicate possible serious gaps which should not be forgotten. In the text below, when using wider time units, local rock units considered are added in brackets.

SAMPLING AND ANALYSE METHODS

The carbonate rocks of the post-Tremadocian Ordovician of the Rapla core (31.7–192.0 m) were sampled for acritarchs (Uutela & Tynni, 1991; 97 samples, mean weight 50 g) and chitinozoans (Nõlvak, unpubl.; 199 samples, mean weight 400 g). The samples were mostly taken at regular intervals (0.5–1.0 m for chitinozoans, 2 m for acritarchs), but close to stratigraphical boundaries. In the case of thin units the sampling density was somewhat higher.

Acritarchs and chitinozoans are absent in the Latorp and in the lower part of the Volkhov stages due to heavy secondary dolomitization of the Rapla section. In the Porkuni Stage these groups, especially chitinozoans, are absent for the same reason and also because of the occurrence of high-energy sediments or reef facies. These local conditions should be considered when evaluating radiation at the very beginning of the section and extinction of Ordovician species at the Pirgu/Porkuni boundary. But in other northern Estonian sections four species of chitinozoans are known from the Porkuni Stage from mudstones of the Siuge Member, three of which have also been recorded from the lowermost Silurian strata. So, if one considers other sections as well, the extinction rate at this level decreases from 100 to 76%. The same is true for acritarchs.

The Rapla core section yields very rich assemblages of acritarchs (319 species, 71 genera) and chitinozoans (110 species). The densely spaced samples and high taxonomic diversity of the assemblages allow the use of this section for examining the diversity patterns and for discussing the possible causes of the microbiotic changes. Most of the numerical data analysed in this paper are presented in three tables.

The identifications on the generic level (*Xus* sp.) have been left out from the number of species (Tables 1, 2). Column 3 shows the number of species recorded in the samples, whereas in column 4 the species are added that were found at higher and lower levels of the section but were missing in samples from an intervening unit under consideration (estimated number of species). The comparison of the data of these two columns (percentage of recorded species of the estimated species, column 5) is an indication of the reliability of the data on observed species.

The data presented in Tables 1, 2 show a good correlation between the number of recorded species and the degree of representativeness of these numbers except for the Arenig and Llanvirn acritarch assemblages. In these assemblages the degree of representativeness is 81–89% though the number of samples per unit is small (1–4). Higher up in the sequence the representativeness of the recorded taxonomic diversity is only 27–35% for one sample per unit, 48% for two samples, and 60–69% for 3–4 samples. The data are more reliable in units with five samples (over 70%, see Table 1).

It should be kept in mind that a very small number of samples per unit (less than three) reduces the probability of obtaining reliable diversity data. On the other hand, we cannot suggest that the estimated number of species (column 4, Tables 1, 2) would be closer to the real diversity state of an assemblage, because most of the species are Lazarus taxa which occur discontinuously (appear–disappear) in a section due to mosaic environmental as well as preservation conditions. The rates of appearing and disappearing species (in per cent, columns 7 and 9) were calculated from the number of taxa recorded from each unit (column 3), but only the real first and last records of a species, not intervening occurrences, were considered as appearances and disappearances.

Table 1

Numerical data for acritarch species and general diversity curves (Figs. 3, 4)

Stratigraphic unit	No. of samples	Species							Genera			Changes at boundaries, %
		Occurring			Appearing		Disappearing		Occurring	Appearing		
		Recorded No.	Estimated No.	%	No.	%	No.	%	No.	No.	%	
1	2	3	4	5	6	7	8	9	10	11	12	13
G_{1-2}	2	29	–	–	4	14	–	–	–	–	–	
												72
F_{II}	4	31	47	66	0	0	18	58	12	0	0	
												63
$F_{I}c^{2}$	7	72	82	88	8	11	44	61	26	0	0	
												46
$F_{I}c^{1}$	13	115	123	93	13	11	46	40	44	6	14	
												32
$F_{I}b$	9	98	122	80	7	7	12	12	34	1	3	
												63
$F_{I}a^{2}$	3	42	122	34	0	0	4	10	33	0	0	
												38
$F_{I}a^{1}$	6	88	127	69	8	9	14	16	39	2	5	
												32
E	9	121	126	96	13	11	17	14	40	2	5	
												76
D_{III}	1	37	137	27	0	0	6	16	14	0	0	
												54
D_{II}^{2}	5	129	160	81	19	15	22	17	38	0	0	
												27
D_{II}^{1}	6	136	157	87	22	16	17	13	47	2	4	
												31
D_{I}	4	107	152	70	20	19	8	7	35	1	3	
												37
C_{III}	3	79	129	61	6	8	7	9	29	2	7	
												22
C_{II}^{2}	4	93	135	69	13	14	9	10	33	0	0	
												27
C_{II}^{1}	3	75	125	60	2	3	4	5	24	1	4	
												26
$C_{I}c$	5	101	140	72	20	20	15	15	38	4	11	
												32
$C_{I}b$	2	106	131	81	32	30	12	11	39	11	28	
												36
$C_{I}a$	3	86	97	89	23	27	7	8	29	5	17	
												34
B_{III}^{2}	4	75	86	87	18	24	6	8	27	5	19	
												27
B_{III}^{1}	1	60	69	87	24	40	3	5	25	9	36	
												44
B_{II}	2	51	(51)	(100)	(51)	(100)	2	4	20	(20)	(100)	
Mean:		85	116	73	13	14	14	17	31	3	9	41

Table 2

Numerical data for chitinozoan species diversity curves (Figs. 4, 5)

Stratigraphic unit	No. of samples	Species							Changes at boundaries, %
		Occurring			Appearing		Disappearing		
		Recorded No.	Estimated No.	%	No.	%	No.	%	
1	2	3	4	5	6	7	8	9	10
F_{II}	5	0	0	0	0	0	0	0	
									(100)
$F_{I}c^{2}$	19	17	17	100	6	35	17	100	
									69
$F_{I}c^{1}$	25	17	20	85	2	12	9	53	
									41
$F_{I}b$	12	24	25	96	6	25	8	33	
									56
$F_{I}a^{2}$	6	15	22	68	2	13	3	20	
									40
$F_{I}a^{1}$	14	21	25	84	9	43	6	29	
									48
E	16	16	21	76	5	31	2	13	
									44
D_{III}	3	9	16	56	4	44	0	0	
									71
D_{II}^{2}	7	13	20	65	1	8	8	62	
									44
D_{II}^{1}	7	15	24	63	2	13	5	33	
									32
D_{I}	8	17	25	68	3	18	2	12	
									44
C_{III}	12	26	29	90	7	27	7	27	
									41
C_{II}^{2}	9	26	28	93	6	23	6	23	
									40
C_{II}^{1}	7	28	29	97	6	21	7	25	
									27
$C_{I}c$	11	24	29	83	3	13	2	8	
									30
$C_{I}b$	9	27	29	93	7	26	6	22	
									40
$C_{I}a$	6	27	28	96	10	37	6	22	
									50
B_{III}^{2}	8	25	25	100	14	56	7	28	
									61
B_{III}^{1}	2	14	14	100	9	64	3	21	
									72
B_{II}	6	9	9	100	9	100	4	44	
	Mean:	19	23	80	6	32	6	30	50

In order to examine the relationship in the diversity curves (Figs. 3, 5) between the duration of each stratigraphic unit and the number of species (or genera) in these units, the data were analysed (Table 3, Fig. 4) by applying the total rate and per taxon rate concepts of Johnson & Kauffman (1990). Because the plots of the total rate and per taxon rate curves agree fairly closely, only the first one is presented.

Table 3

Appearance and disappearance rates of acritarch and chitinozoan taxa in the Rapla section

Stratigraphic unit	Duration Ma	Acritarchs						Chitinozoans			
		Appearance rate				Disappearance rate		Appearance rate		Disappearance rate	
		Total		Per taxon		Total	Per taxon	Total	Per taxon	Total	Per taxon
		spp.	gen.	spp.	gen.	spp.	spp.	spp.	spp.	spp.	spp.
F_{II}	1.6	0	0	0	0	11.3	0.36	0	0	0	0
$F_{I}c^{2}$	1.8	4.4	0	0.06	0	24.4	0.34	3.3	0.20	9.4	0.60
$F_{I}c^{1}$	1.7	7.6	3.5	0.07	0.08	27.1	0.24	1.2	0.07	5.3	0.31
$F_{I}b$	1.6	4.4	0.6	0.04	0.02	7.5	0.08	3.8	0.16	5.0	0.21
$F_{I}a^{2}$	1.3	0	0	0	0	3.1	0.07	1.5	0.10	2.3	0.15
$F_{I}a^{1}$	1.4	5.7	1.4	0.06	0.04	10.0	0.11	6.4	0.31	4.3	0.20
E	1.3	10. 0	1.5	0.08	0.04	13.1	0.11	3.8	0.24	1.5	0.10
D_{III}	1.3	0	0	0	0	4.6	0.12	3.1	0.34	0	0
D_{II}^{2}	1.3	14. 6	0	0.11	0	16.9	0.13	0.8	0.06	6.2	0.47
D_{II}^{1}	1.3	16. 7	1.5	0.12	0.03	13.1	0.10	1.5	0.10	3.8	0.26
D_{I}	1.2	16. 7	0.8	0.16	0.02	6.7	0.06	2.5	0.15	1.7	0.10
C_{III}	1.3	4.6	1.5	0.06	0.05	5.4	0.07	5.4	0.21	5.4	0.21
C_{II}^{2}	1.6	8.1	0	0.09	0	5.6	0.06	3.8	0.14	3.8	0.14
C_{II}^{1}	1.2	1.7	0.8	0.02	0.03	3.3	0.04	5.0	0.18	5.8	0.21
$C_{I}c$	1.8	11. 1	2.2	0.11	0.06	8.3	0.08	1.8	0.07	1.1	0.05
$C_{I}b$	1.5	21. 3	7.3	0.20	0.19	8.0	0.08	4.7	0.17	4.0	0.15
$C_{I}a$	1.5	15. 3	3.3	0.18	0.11	4.7	0.05	6.7	0.25	4.0	0.15
B_{III}^{2}	1.8	10. 0	2.8	0.13	0.10	3.3	0.04	7.8	0.31	3.9	0.16
B_{III}^{1}	1.2	20. 0	7.5	0.33	0.30	2.5	0.04	7.5	0.54	2.5	0.18
B_{II}	2.0	25. 5	10.0	0.50	0.50	1.0	0.04	4.5	0.50	2.0	0.22
Mean:		9.9	2.2	0.12	0.08	9.0	0.11	3.8	0.18	3.6	0.19

Total rate = No. of appearing (or disappearing) taxa per 1 Ma. Per taxon rate = total rate per taxon recorded from a stratigraphic unit.

ACRITARCHS

In the Rapla section the acritarch flora is rich and diverse practically throughout the Ordovician. Each stratigraphic unit is characterized on average by 85 species belonging to 31 genera. Only some minimum levels, for instance the Oandu or Porkuni stages, are nearly 2.5 times less diverse. The maximum diversity in the Keila Stage differs from the average only 1.5 times.

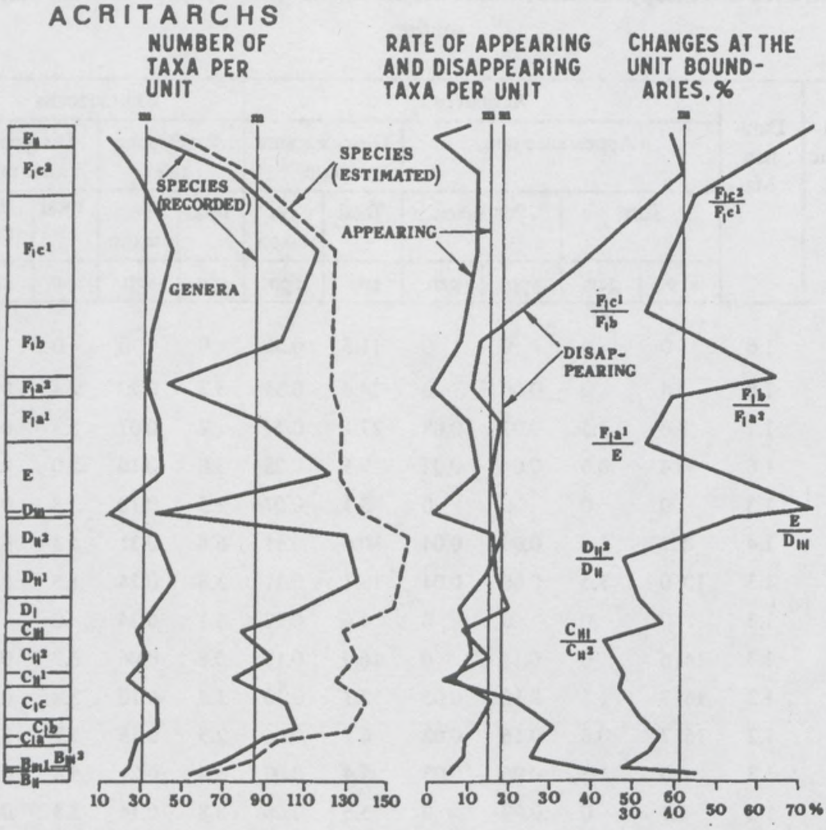

Fig. 3. Acritarch diversity. Stratigraphical column according to thickness. Unit symbols see Fig. 2. m, mean value level. Numerical data in Table 1.

In the Arenig and Llanvirn (the percentage of appearing species is 25–40%, that of new genera is 17–36%) the appearance of new taxa is fairly rapid (Table 1, Figs. 3, 4). Origination intensity decreases in the early Caradoc, falling below the mean level (14%) and, except for the Jõhvi–Keila stages, it is constantly low up to the late Ordovician. The total rate of appearance (Fig. 4) has additional peaks in the curve, but shows the same regularity. An increase in the origination intensity in the Ashgill (early

Pirgu time) is notable with respect to the total appearance rate of the genera exceeding the mean level (Fig. 4). This seems to be connected with the appearance of several new genera (*Estiastra*, *Pulvinosphaeridium*, *Schismatosphaeridium*, etc.) which are more widely distributed in the Silurian.

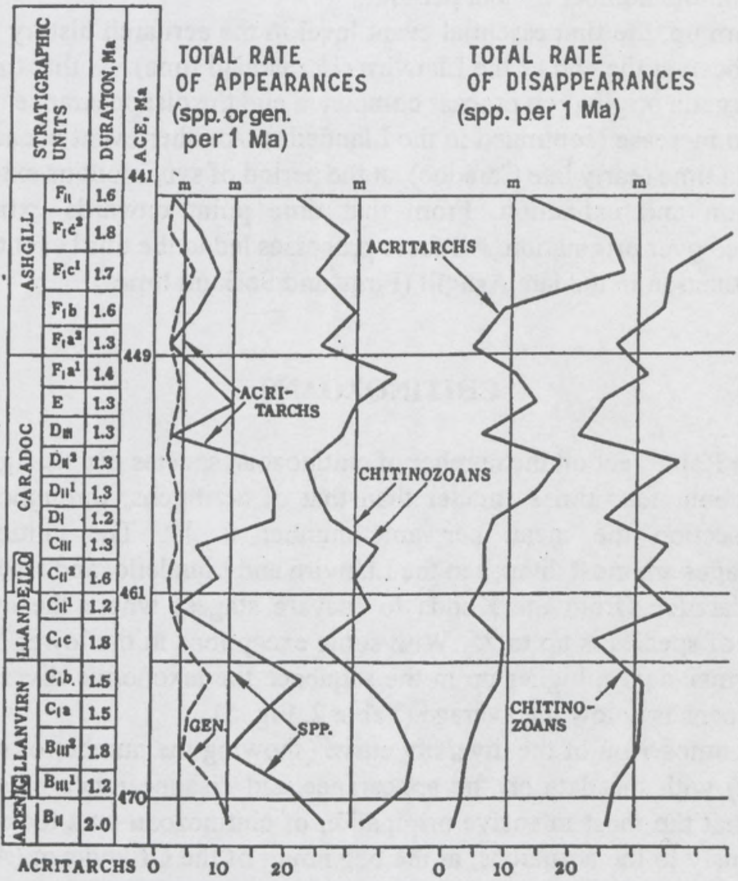

Fig. 4. Total appearance and disappearance rates of acritarchs and chitinozoans. Explanation of unit symbols see Fig. 2. Database in Table 3. Stratigraphical column according to the geochronometric time scale by Tucker et al., 1990.

An opposite tendency can be observed in the change of the disappearance intensity of taxa (Table 1, Fig. 3). The initially very slow process reaches the first low peak (but still below the mean level for the whole Ordovician) in the Llandeilo (Uhaku time); in the late Caradoc (beginning from Keila time) the disappearance intensity reaches almost the average level (17%). In the Ashgill (Pirgu and Porkuni times) it rises to the value of 40–61% (Table 1) and this is close to what can be termed as mass extinction. The process is also seen in Fig. 3 which shows a rapid decrease in the number of taxa. The total disappearance rate curve (Fig. 4) reveals the same tendencies with distinct peaks in Keila and Pirgu times. In Keila time the per taxon rate of disappearances rises abruptly and is there twice as high as in lower units (Table 3). This shows that the intensive extinction which started in the late Caradoc is not related to an increase in the number of taxa per unit.

To sum up, the first essential event level in the acritarch history seems to have been at the end of the Llanvirn (Lasnamägi time). At that time the first energetic origination process completed and the disappearance of taxa started to increase (continued in the Llandeilo). Another event occurred in late Keila time (early late Caradoc), at the period of synchronous extensive origination and extinction. From that time point onwards extinction dominated over origination. All these processes led to the third event level: mass extinction in the late Ashgill (Pirgu and Porkuni times).

CHITINOZOANS

In the Rapla section the number of chitinozoan species per stratigraphic unit is about four times smaller than that of acritarchs; throughout the whole section the mean per unit number is 19. The chitinozoan assemblages are most diverse in the Llanvirn and Llandeilo, and also in the lower Caradoc (from the Kunda to Idavere stages) where the average number of species is up to 25. With some exceptions in the lower Nabala and Vormsi stages, higher up in the sequence the taxonomic diversity of chitinozoans is below the average (Table 2, Fig. 5).

The comparison of the diversity curve (showing the number of species per unit) with the data on the appearance and disappearance of species shows that the most intensive origination of chitinozoan taxa took place, analogously to the acritarchs, at the beginning of the Ordovician (Arenig, Llanvirn) and at the end of the Caradoc, when the number of short-ranging species was relatively large. Origination intensified slightly also at the end of Pirgu time, but most of the Caradoc and Ashgill were characterized by a low appearance rate (Table 2, Fig. 5). The total rate of appearance (Table 3, Fig. 4) confirms the above statements, but shows additionally a relatively high rate during Kukruse–Idavere times and in early Nabala time, but also stresses the minimums in the Llandeilo (Uhaku time) and "middle" Caradoc (Jõhvi–Keila times).

The small origination event in Oandu time (Figs. 4, 5) is associated with a very low number of species. Therefore, the appearance of only four additional species has a considerable impact. Actually, this interval represents the minimum diversity level of chitinozoan assemblages, caused mostly by a substantial extinction event in the latest Keila time when 63% of the species disappeared.

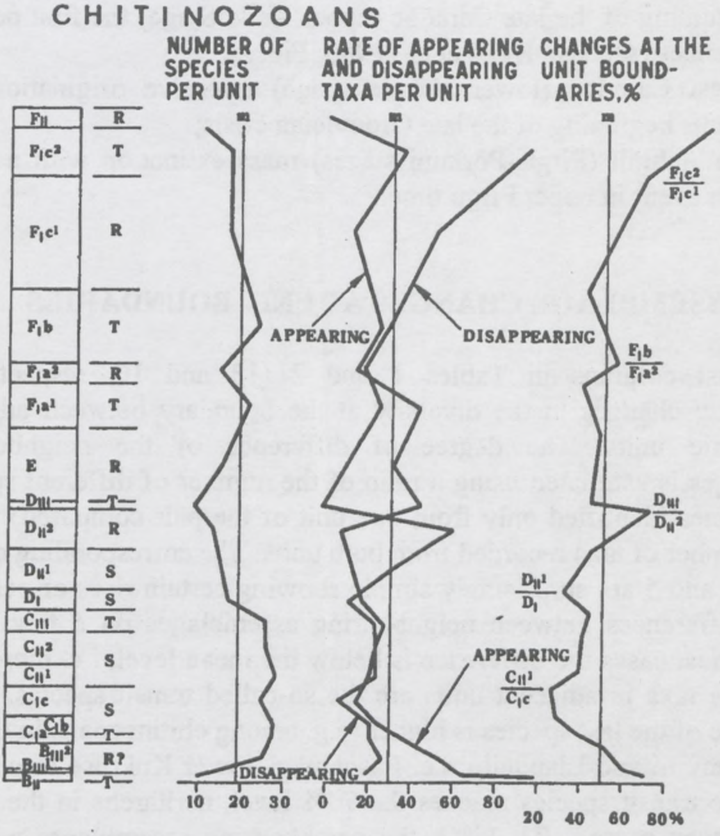

Fig. 5. Chitinozoan diversity. Explanations see Figs. 2 and 3. Numerical data in Table 2. R, S, T, intervals with a prevailing tendency in environmental evolution: R, shallower water rocks, conventionally regression; T, more argillaceous rocks, deepening or conventionally transgression; S, without clear general tendency.

Earlier (from Kunda to lower Keila times) the disappearance rate (Fig. 5) was at the average level. After the low stand during Oandu time it began to increase, reaching the average level in Nabala time and the maximum level at the end of the Ordovician (in the upper Pirgu Stage). The total rate of disappearances (Fig. 4) varies more, but in general it resembles the disappearance curve (Fig. 5). As in the latter, there is a prominent peak in lower Kukruse time and a second one in Idavere time, possibly related to the extinction of a specific but short-term assemblage in the kukersite-bearing beds. Thus, with regard to the chitinozoans, the following main changes in the diversity of the assemblages can be distinguished:

(1) early Llanvirn (Kunda Stage) maximum of radiation;

(2) latest Llandeilo–early Caradoc (Kukruse–Idavere stages) interval of rapid diversity fluctuations;

(3) beginning of the late Caradoc (upper Keila Stage) the first peak of disappearences (see also Hints et al., 1989; Fig. 2);

(4) latest Caradoc (lower Nabala Stage) intensive origination and simultaneous beginning of the late Ordovician crisis;

(5) late Ashgill (Pirgu–Porkuni stages) mass extinction with a small origination event in upper Pirgu time.

ASSEMBLAGE CHANGES AT UNIT BOUNDARIES

The last columns in Tables 1 and 2 (13 and 10, respectively) characterize changes in the diversity at the boundary between adjacent stratigraphic units. The degree of difference of the neighbouring assemblages is estimated using a ratio of the number of different species (i.e. the ones identified only from one unit of the pair compared) to the whole number of taxa recorded from both units. The corresponding curves in Figs. 3 and 5 are surprisingly simple showing certain rises or peaks of greater differences between neighbouring assemblages on a few levels only. In most cases the difference is below the mean level, i.e. more than half of the taxa in adjacent units are the so-called transit species. If the percentage of the last species is higher (e.g. among chitinozoans in the top of the Llanvirn and Llandeilo, i.e. Lasnamägi–lower Kukruse stages, the rate of the transit species reaches the 70% level, acritarchs in the lower Caradoc even more – 73–78%), the neighbouring assemblages are less different.

The greatest differences involve the acritarch assemblages at the beginning of the late Caradoc, at the lower boundary of the Ashgill, and in the late Ashgill. All the other assemblages are poorly distinguished, except that of the Oandu Stage (Figs. 3, 5), as very considerable changes in the chitinozoan and acritarch diversity take place on the lower and upper boundaries of this stage. The distinctiveness of the earliest and latest Ordovician assemblages evidently results from the early Ordovician evolutionary radiation and the late Ordovician mass extinction. The late Caradoc change could be related to a facies change. On the other hand, a small number of abrupt changes of assemblages at the boundaries of stratigraphic units indicates that the Rapla section, despite a large number of discontinuity surfaces, is relatively complete or there are evidently only a few substantial hiatuses in the sequence.

DISCUSSION

The comparison of the total rate of appearance curves by acritarchs and chitinozoans displays a surprising strict regular alternation of their peaks

(Fig. 4). High total rate values of acritarchs often fall into the low stand of the chitinozoan curve. The only coincidence of peaks is observed at the beginning of the Llanvirn, during the first intensive origination of both groups. Later their appearance peaks diverge: chitinozoans prefer rocks (environment) with a higher clay content, acritarchs less argillaceous varieties (Figs. 2, 5). In general, in spite of a very variable shape, both curves have an obvious falling tendency during the whole post-Arenig Ordovician.

The chitinozoan and acritarch disappearance total rate curves are similar (Fig. 4), in particular with major extinctions in the late Caradoc (Keila) and middle Ashgill (Pirgu). The curve for the latter reflects evidently the general late Ordovician extinction, but the reasons of the former are not so easy to understand.

Chitinozoans have a very changeable history from the latest Llandeilo (lower Kukruse) until late Caradoc (upper Keila) with three high disappearance peaks. This time interval is distinguished by a more or less stable environment characterized by the occurrence of laminar and nodular wackestones (Figs. 2, 5). This was also the time of the most intense volcanic activity: the thickest K-bentonite bed is described from the bottom of the Keila Stage (Põlma, 1972). However, on the other hand, there is no evidence that the ashbeds had any influence on the faunal diversity.

Excursions of the total appearance and disappearance rate curves (Fig. 4) are parallel and especially well expressed in the case of chitinozoans. Accordingly, one reason for disappearance peaks was a higher rate of occurring taxa. The same might be concluded from the parallel rise of the disappearance rates of both groups during the Arenig–Llanvirn due to rapid origination and diversity rise. This does not mean that environmental aspects of the origination–extinction processes should be ignored, but it does indicate that these are not the only causes and in different intervals some other factors may dominate.

The late Keila extinction and diversity minimum in the Oandu time, profound assemblage changes on the boundaries of the latter unit coinciding with a considerable sedimentological change in the section evidence more about environmental influence in the biotic process. However, our data are insufficient for evaluating the significance of these events, but the data presented by Sepkoski (1995) confirm that the late Keila bioevent has not only a local but also much wider importance.

CONCLUSIONS

(1) The comparison of the diversity dynamics in the acritarch flora and chitinozoan fauna (?) in an Ordovician section of northern Estonia shows a

significant correlation in the pattern of the most general processes represented by coinciding Arenig–Llanvirn radiation–origination events and mass extinctions at the end of the Ordovician.

(2) Both groups have in the late Caradoc (Oandu time) a very short-term episode (about 1 Ma) that might be called the Oandu crisis. It is characterized by an extreme diversity low, distinct changes in the composition of the microfossil assemblages on its boundaries, and is preceded by a remarkable extinction event at the end of Keila time. This event seems to be caused by environmental agents and is globally observable.

(3) The fluctuation of different curves concurrently with some lithological features (clay content, relations of mud-grain-skeletal particles, etc.) do not explain unambigously the diversity dynamics, especially because the correlation is not consistent with the parallel total appearance and disappearance rate curves. Therefore a higher diversity of a group seems to be one cause for a higher extinction rate.

(4) The terminal Ordovician mass extinction began in both groups earlier than might be expected according to the short-term glaciation event proposed by Brenchley et al. (1994).

(5) The general pattern of microfossil diversity in the Rapla core is the same as established by Sepkoski (1995) for different groups of invertebrates: late Arenig–Llanvirn diversity rise and, with a maximum in the early–middle Caradoc (chitinozoans had a peak earlier), followed by a late Caradoc drop of diversity and a terminal Ordovician mass extinction.

(6) The latter conclusion advocates that general tendencies of biotic diversity might be observed also in only one good section if analysed adequately.

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LISAND RAPLA PUURSÜDAMIKU ORDOVIITSIUMI MIKROFOSSIILIDE MITMEKESISUSE UURIMISELE

Dimitri KALJO, Jaak NÕLVAK, Anneli UUTELA

On uuritud akritarhide ja kitiinikute (*Chitinozoa*) mitmekesisuse muutumist Volhovi lademest kuni Porkuni lademeni. Selgus, et mõlemad rühmad tegid läbi kiire radiatsiooni Arenigis ja Llanvirnis. Hiliskaradokis Keila ea lõpul toimus esimene suurem väljasuremine (mitmekesisuse miinimum Oandu lademes) ning ordoviitsiumi lõpul (Pirgu ja Porkuni eal) massiline väljasuremine. See üldskeem on sarnane J. Sepkoski (1995) kirjeldatud selgrootute mitmekesisuse muutumise seaduspärasusega.

ДОПОЛНЕНИЯ К ИЗУЧЕНИЮ ТАКСОНОМИЧЕСКОГО РАЗНООБРАЗИЯ ОРДОВИКСКИХ МИКРОФОССИЛИЙ ИЗ СКВАЖИНЫ РАПЛА, СЕВЕРНАЯ ЭСТОНИЯ

Димитри КАЛЬО, Яак НЫЛВАК, Аннели УУТЕЛА

Изучены акритархи и хитинозои из скв. Рапла, вскрывшей ордовикские отложения от верхов аренига до верхов ашгилла. Проанализировано количество таксонов в стратоне, скорость появления и исчезновения таксонов в стратоне за миллион лет, а также таксономические изменения на границах стратонов. Обе группы подверглись сильной радиации в арениге–лланвирне, первое их существенное вымирание произошло в позднем карадоке и уже массовое вымирание – в конце ордовика (пиргуское–поркуниское время). Остальные мелкие изменения, видимо, контролировались фациальными условиями. Общая схема изменений таксономического разнообразия микрофоссилий сходна с закономерностью, установленной Дж. Сепкоски (1995) для некоторых групп беспозвоночных.

Paper IX

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Journal of Micropaleontology, **15,** 21-35.

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Precise chitinozoan dating of Ordovician impact events in Baltoscandia

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ABSTRACT - The chitinozoan biostratigraphy of four Ordovician impact craters has been investigated. Three of these (Tvären, Kärdla and Lockne) contain complete sequences of early Caradoc age, while the Granby crater contains rocks of late Arenig age yielding two bentonitic horizons at their top. Chitinozoans, together with graptolites and other planktic organisms, were the first to invade the craters after the impact event. It has therefore been possible to date the impact events with a precision of less than one million years. An immigration of graptolites from Australia during the late Arenig corresponds to an immigration of chitinozoans from Gondwana at this time. Two stratigraphically important taxa, not previously described or discussed, *Lagenochitina* sp. A aff. *capax* and *Laufeldochitina* sp. A aff. *striata* are commented upon. Three species, *Cyathochitina hunderumensis*, *Spinachitina tvaerenensis* and *Tanuchitina granbyensis*, are described as new. *J. Micropalaeontol.* **15**(1): 21–35. April 1996.

INTRODUCTION

Five impact events of supposed Ordovician age have been reported from Baltoscandia (see Henkel & Pesonen, 1992). Four of these were recently dated by Grahn & Nõlvak (1993). A crater below the bottom of the Lumparen Bay in the Åland archipelago, between Sweden and the mainland of Finland, was for a long time considered to be of tectonic origin (see Winterhalter 1982), but Merrill (1980) regarded it to be an Ordovician or younger impact structure. More recent investigations show that the Lumparen Bay structure probably is an impact crater that was already deeply eroded when Ordovician sedimentation commenced in the area during the early Arenig. The Lumparen event might be of early Cambrian age since sandstones of this age, reminiscent of resurge deposits, are distributed in the area (Hagenfeldt, pers. comm. 1993). The oldest Ordovician crater, of late Arenig age, is situated at Granby in south Sweden (Fig. 1). The three other craters, of early Caradoc age, occur at Tvären and Lockne in central Sweden, and at Kärdla on the Island of Hiiumaa in northwest Estonia (Fig. 1).

The maximum thickness of the Ordovician sequences outside the craters is less than 200 m in Sweden and Estonia, and stratigraphical gaps occur frequently through the sequence. In contrast to the sequences surrounding the craters, those within the craters are virtually complete. The crater sequences are therefore excellent models for ecological studies of restricted environments and for high-resolution biostratigraphy. The present paper is a detailed study of the chitinozoan biostratigraphy in the four Ordovician craters investigated. The illustrated specimens are deposited in the collections of the Institute of Geology, Rennes (IGR) under the numbers 58601–58646. The coordinates are those of the England Finder grid.

CHITINOZOAN BIOSTRATIGRAPHY

Nõlvak & Grahn (1993) recently published a chitinozoan biozonation comprising 23 zones and subzones for the Ordovician of Baltoscandia. All eight crater sections investigated from or close to the four craters yielded the index species and/or characteristic species for three of these zones, which allowed a very precise dating. The biostratigraphy from the sections related to the four craters will be described below.

Granby crater

The Granby crater is situated in the subsurface about 4 km southeast of Vadstena, a little town on the east shore of Lake Vättern. It has been discussed by Bruun & Dahlman (1982) and Grahn & Nõlvak (1993). Two cores from the crater were investigated, viz. from the Fylla 9 borehole about 1 km east of the west crater rim, and the Fylla 3 borehole, about 800 m further to the east (Fig. 2). The latter borehole was made near the centre of the crater above the central uplift (Bruun & Dahlman, 1982). The impact sediments in the crater are overlain by calcareous and glauconitic mudstone interbedded by grey limestone. The impact event is placed in the lower part of the *Cyathochitina regnelli* chitinozoan Zone, since a characteristic chitinozoan assemblage from the *Conochitina cucumis* chitinozoan Zone is missing (Nõlvak & Grahn 1993). The index fossil has not been found (Grahn & Nõlvak, 1993), but the presence of for instance *Conochitina decipiens* (Pl. 1, figs 8, 9) together with *Cyathochitina hunderumensis* sp. nov. (see Systematic Desciptions; Figs 3, 4; Pl. 1, figs 1, 10, 11) and *Tanuchitina granbyensis* sp. nov. (see Systematic Descriptions; Figs 3, 4; Pl. 1. figs 2–5) suggests a late Arenig age. In the crater sequence occur also, among others, *Rhabdochitina gracilis* (Pl. 3, fig. 8), *Conochitina* sp. 1 (Pl. 3, fig. 5) and *Lagenochitina esthonica* (Pl. 3, fig. 9). Two bentonitic horizons found in the Fylla 9 core (Fig. 3) are probably of the same age, as evidenced from the presence in these levels by *Clavachitina poumoti* (Pl. 1, figs 6, 7), a species reported from Upper Arenig strata in Australia (Combaz & Peniguel,

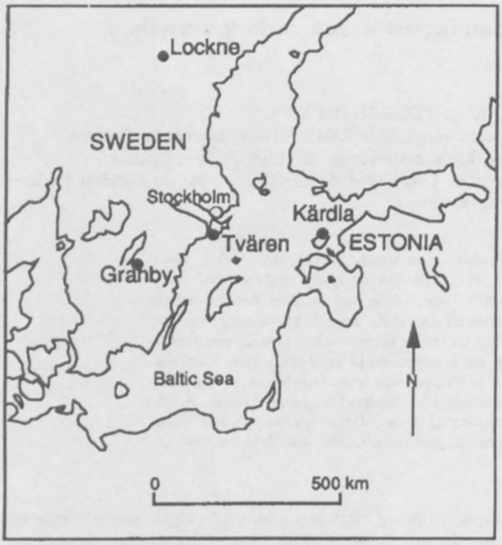

Fig. 1. Map showing the sites of the investigated Ordovician impact craters in Baltoscandia.

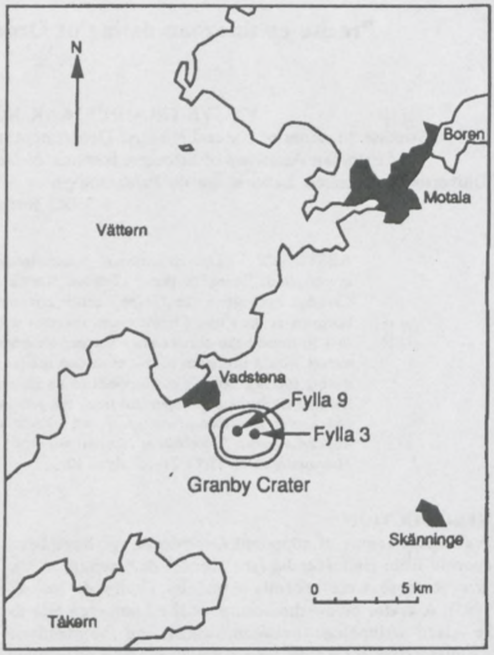

Fig. 2. Map showing the sites of the investigated boreholes in the Granby crater, Östergötland, Sweden.

1972), Svalbard (Bockelie 1980) and eastern Canada (Achab, 1986). It should be noted that in the graptolite Zone of *Didymograptus hirundo* (late Arenig) there is also evidence of an immigration of graptolites from Australia into Baltoscandia (Nilsson, 1983).

Tvären crater

The Tvären crater is situated below the bottom of the Tvären Bay, about 72 km south–southwest of Stockholm. It has been discussed by Flodén *et al*. (1986) and Lindström *et al*. (1994). Two boreholes, Tvären 1 and 2, were drilled in the structure (Fig. 5). The former borehole was placed in the crater rim, while the latter (Fig. 6) penetrated the whole sedimentary sequence ending in the basement breccia. In the first post-impact sediments, consisting of dark grey calcareous mudstone with interbedded limestone, the index species *Laufeldochitina stentor* (Pl. 2, fig. 6) occur together with *Lagenochitina* sp. A aff. *capax* (Pl. 3, figs. 1, 2) and *Laufeldochitina* sp. A aff. *striata* (Pl. 2, figs 9, 10). The presence of *L.* sp. A aff. *striata* is characteristic for a short interval in the upper part of the *L. stentor* Zone, where the index species is rare or absent (Nõlvak & Grahn, 1993). *Laufeldochitina* sp. A aff. *striata* was also found in a coarse breccia formed by the resurge turbidite immediately after the impact. This places the base of the Tvären post-impact sequence, and the impact event, within beds corresponding to the lower Peetri Member of the Viivikonna Formation

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Explanation of Plate 1

Selected chitinozoans from the Granby crater. Late Arenig. **Fig. 1.** *Cyathochitina hunderumensis* sp. nov. Holotype. Fylla 3 borehole, core sample at 254.30 m. Specimen in lateral view, SEM ×300. IGR 58614 (L. 39/4). **Fig. 2.** *Tanuchitina granbyensis* sp. nov. Holotype. Fylla 9 borehole, core sample at 272.20 m. Specimen in lateral view, SEM ×75. IGR 58601 (O. 43). **Fig. 3.** *Tanuchitina granbyensis* sp. nov. Holotype. Fylla 9 borehole, core sample at 272.20 m. Detail of the base showing the carina, SEM ×500. **Fig. 4.** *Tanuchitina granbyensis* sp. nov. Fylla 9 borehole, core sample at 270.20 m. Specimen in lateral view, SEM ×75. IGR 58601 (O. 40/1). **Fig. 5.** *Tanuchitina granbyensis* sp. nov. Same species as in fig. 4. Detail of the base showing the broken carina, SEM ×500. **Fig. 6.** *Clavachitina poumoti* (Combaz & Peniguel 1972). Fylla 9 borehole, core sample at 199.35 m. Specimen in lateral view, SEM ×150. IGR 58607 (O. 34/2). **Fig. 7.** *Clavachitina poumoti* (Combaz & Peniguel, 1972). Fylla 9 borehole, core sample at 199.35 m. Specimen in lateral view, SEM ×100. IGR 58607 (R. 36). **Fig. 8.** *Conochitina decipiens* Taugourdeau & Jekhowsky, 1960. Fylla 3 borehole, core sample at 254.30 m. Specimen in lateral view, SEM ×300. IGR 58614 (N. 36). **Fig. 9.** *Conochitina decipiens* Taugourdeau & Jekhowsky, 1960. Fylla 9 borehole, core sample at 228.10 m. Specimen in lateral view, SEM ×200. IGR 58604 (Q. 41/2). **Fig. 10.** *Cyathochitina hunderumensis* sp. nov. Fylla 9 borehole, core sample at 265.30 m. Specimen in lateral view, SEM ×300. IGR 58604 (Q. 37/1). **Fig. 11.** *Cyathochitina hunderumensis* sp. nov. Fylla 9 borehole, core sample at 265.30 m. Specimen in lateral view, SEM ×300. IGR 58604 (P. 37/3).

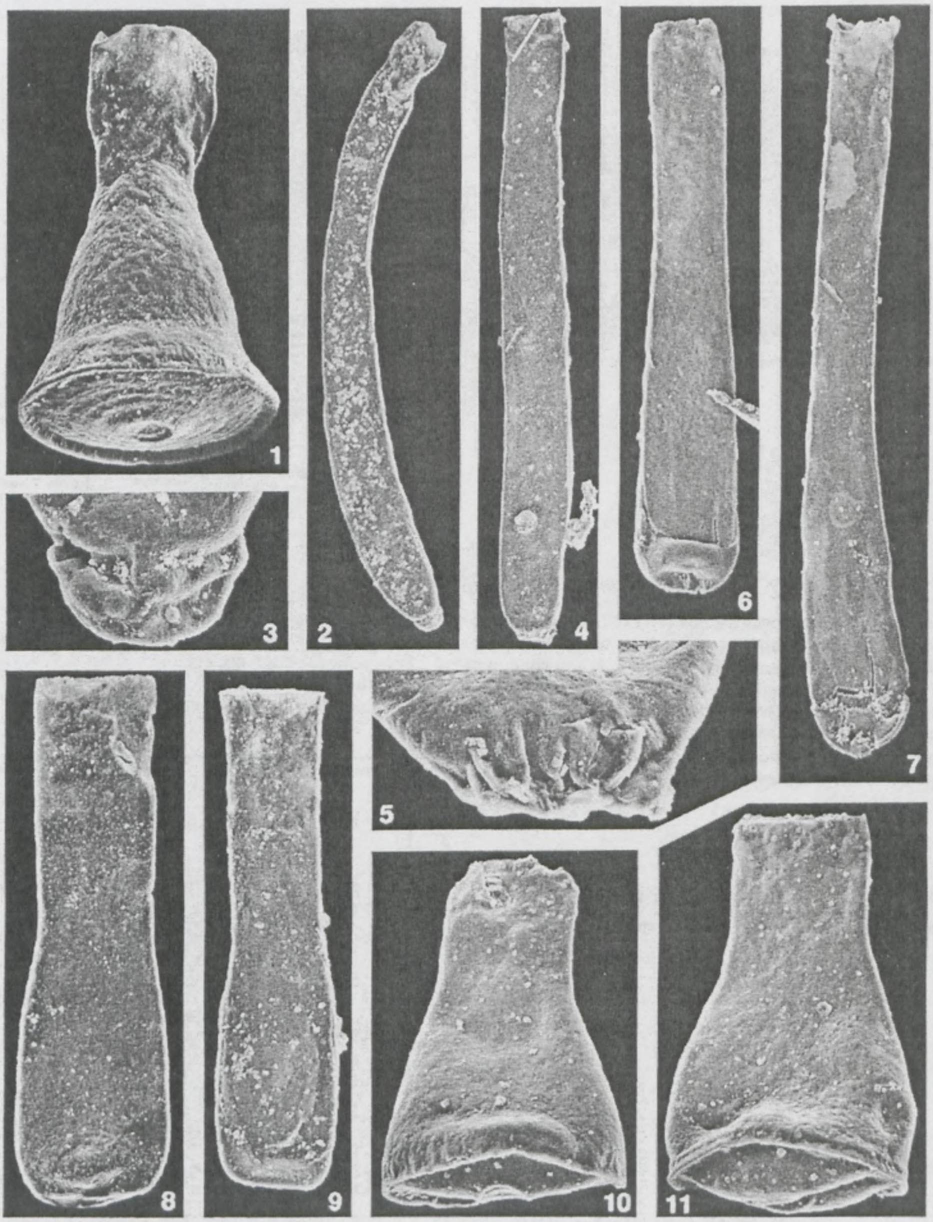

Plate 1

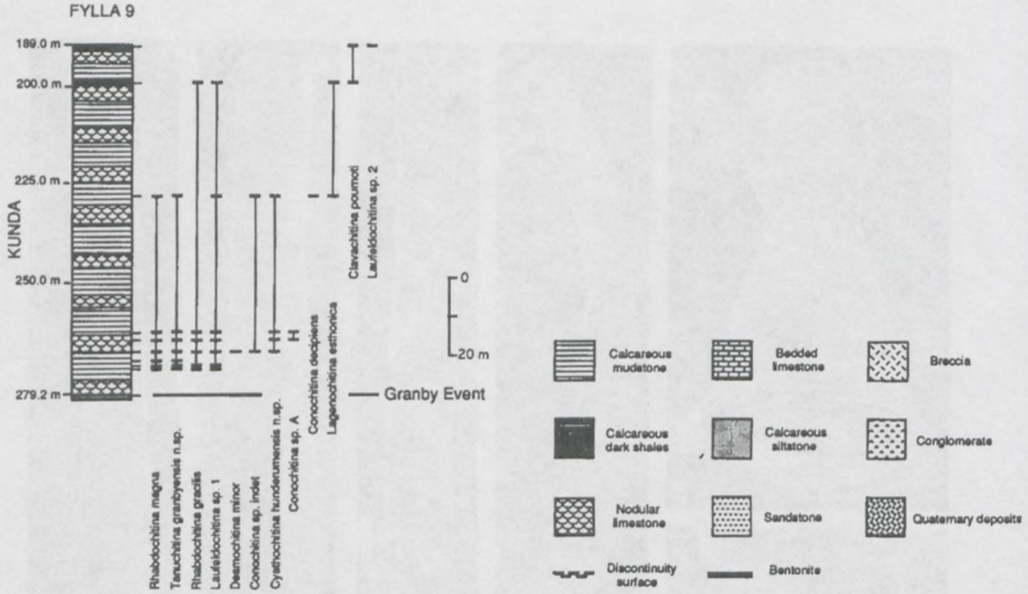

Fig. 3. Faunal log of the Fylla 9 borehole, Granby crater, with sedimentary legend.

(Kukruse Stage) in North Estonia (Figs 6, 14). An important taxon *L.* sp. A aff. *capax* (ranging from upper Kukruse to lower Idavere; Nõlvak, unpublished data), is also confirming an early Caradoc age (Fig. 14). Other common species are *Conochitina minnesotensis* (Pl. 3, fig. 7), *Calpichitina complanata*, *Calpichitina lecaniella* (Pl. 2, figs 3, 5), *Cyathochitina kuckersiana* and *Desmochitina ovulum*. In the topmost layer of the pre-Quaternary rocks in the Tvären 2 borehole *Conochitina tigrina* (Pl. 2, figs 1, 2, Pl. 3, fig. 4) occurs with *Spinachitina tvaerenensis* sp. nov. (see Systematic Descriptions; Fig. 6; Pl. 2, figs 4, 7, 8, Pl. 3, fig. 6, 11). These species appear in a short interval at the top of the Kukruse Stage in Baltoscandia.

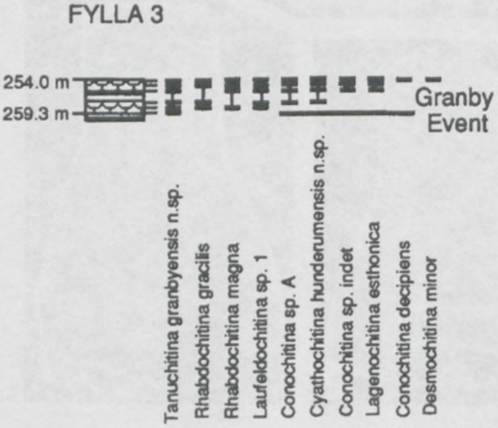

Fig. 4. Faunal log of the Fylla 3 borehole, Granby crater.

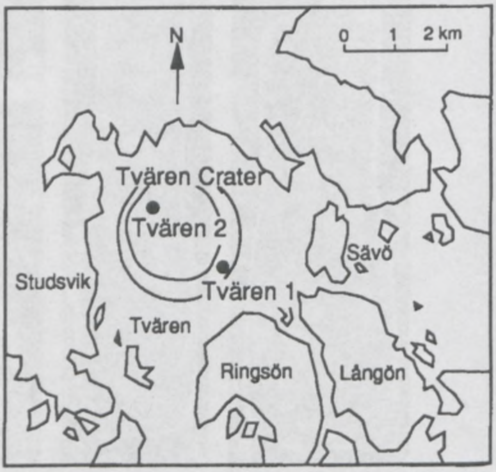

Fig. 5. Map showing the site of the Tvären 1 and 2 boreholes in the Tvären crater, Södermanland, Sweden.

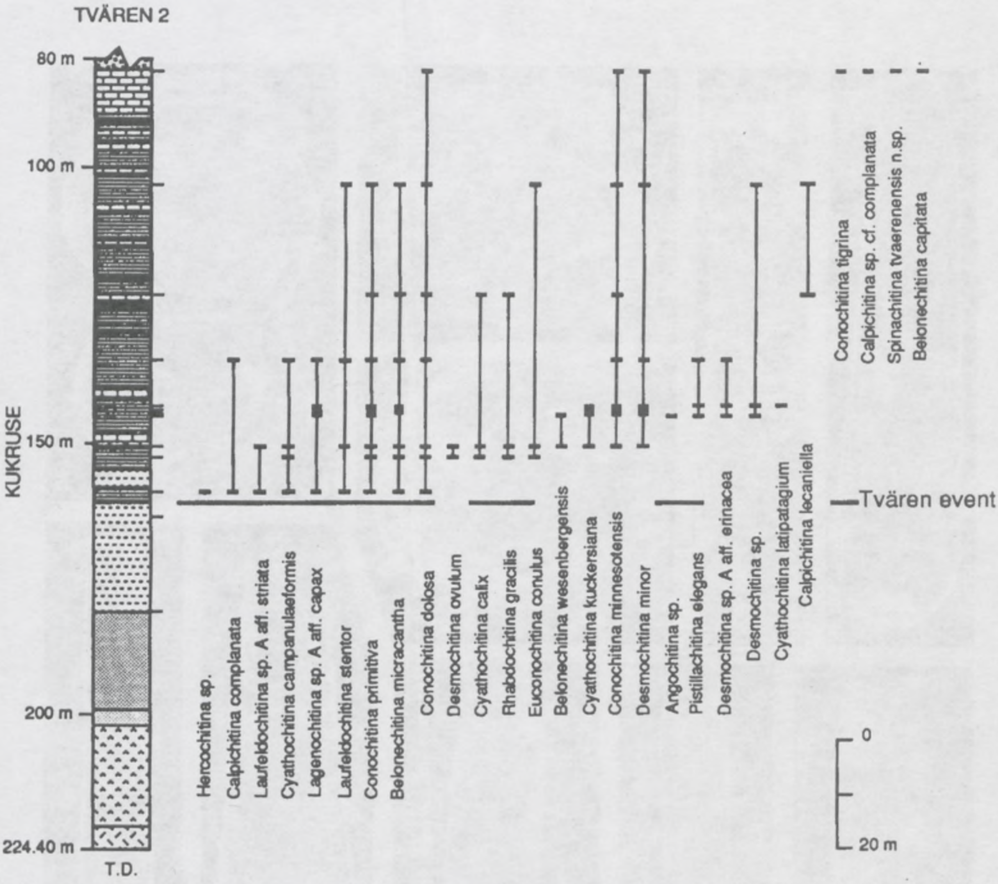

Fig. 6. Faunal log of the Tvären 2 borehole, Tvären crater.

Kärdla crater

The Kärdla crater is situated in the subsurface, just east of the Kärdla city, and on the north coast of the Island of Hiiumaa. It has been discussed by Puura & Suuroja (1992) and Grahn & Nõlvak (1993). The chitinozoan biostratigraphy in two boreholes drilled in the structure has been investigated. Paluküla 383 (Fig. 8) is situated within the crater near the northeastern rim, and Männamaa (Fig. 9) about 20 km southwest of the crater (Fig. 7). The Paluküla 383 borehole terminated just above the first post-impact sediments, which consist of calcareous mudstone with rare limestone intercalations. However, ejecta from the impact are spread over large areas in northwest Estonia. They consist of a quartz-rich limestone, known as the Kisuvere Member of the lower Tatruse Formation (Põlma *et al.* 1988). This layer is present at 164.82–164.93 m and has been dated in the Männamaa borehole. It gives a precise dating of the first post-impact sedimentation in the crater, as well as the impact event (Grahn & Nõlvak, 1993). In both the Paluküla 383 and Männamaa boreholes *Lagenochitina* sp. A aff. *capax* and *Spinachitina multiradiata* appear above the first occurrence of the index species *Lagenochitina dalbyensis*. This suggests that the impact event cannot be much younger than the middle part of the *L. dalbyensis* Zone, since *S. multiradiata* has its first occurrence in the middle part of this zone. In the Männamaa borehole the Kisuvere Member is situated above the last occurrence of *Angochitina curvata*, and before the first occurrence of *L. dalbyensis*, which means that the impact event, and the first post-impact sedimentation, took place in the early Caradoc and corresponds to the transition between the chitinozoan Zones of *A. curvata* and *L. dalbyensis* (Grahn & Nõlvak, 1993; Nõlvak & Grahn 1993; Fig. 14).

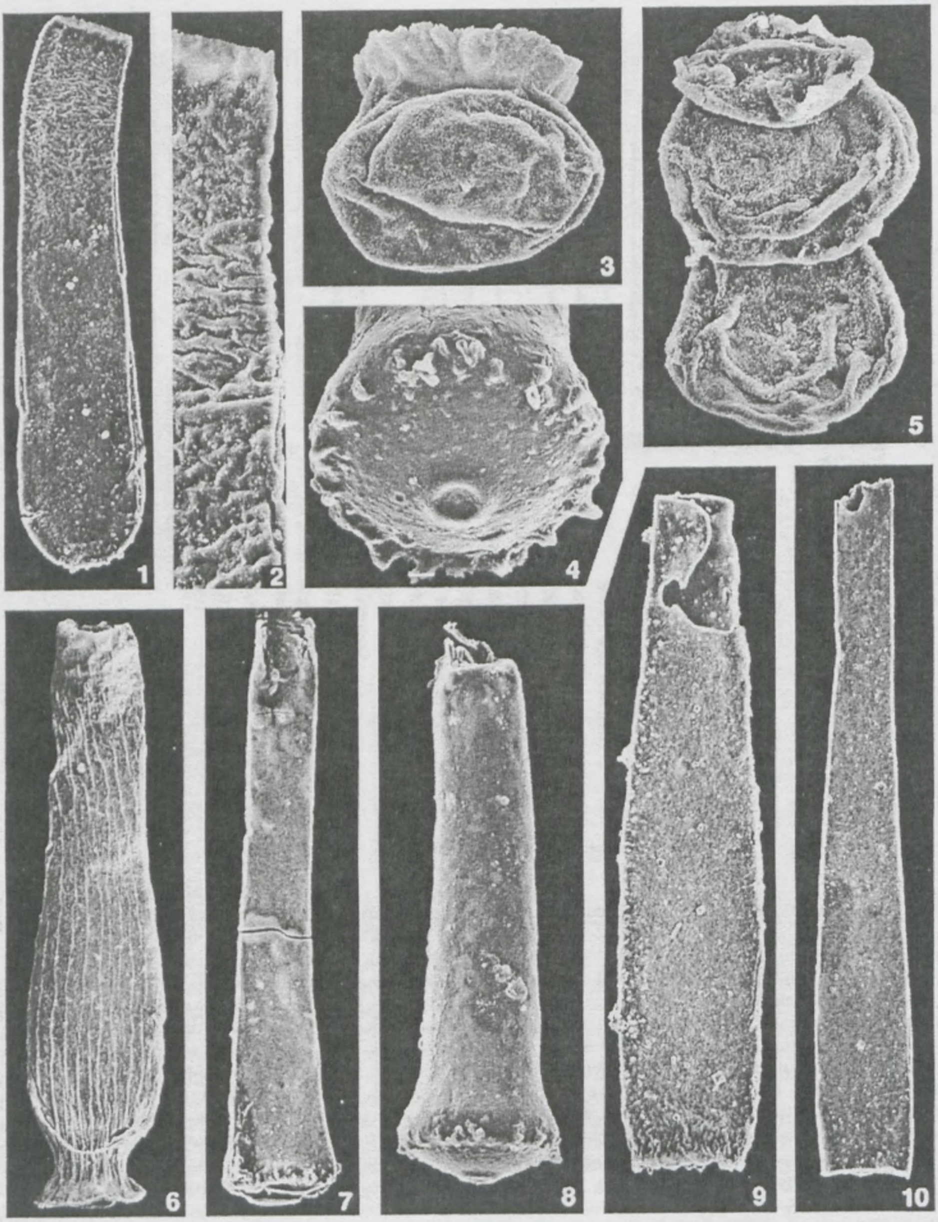

Plate 2

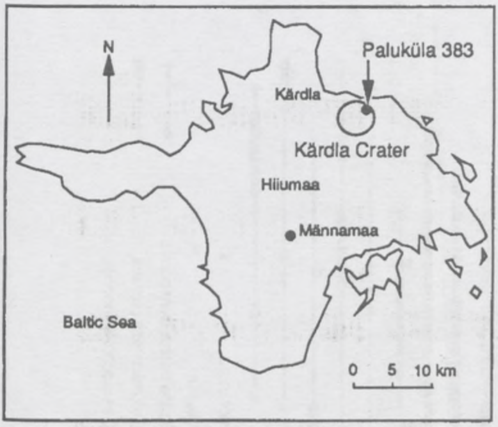

Fig. 7. Map showing the sites of the investigated boreholes in connection with the Kärdla crater, Hiiumaa, Estonia.

Lockne crater

The Lockne crater is situated at Lake Lockne, about 20 km southeast of the town of Östersund in the province of Jämtland. The crater has been described by Simon (1987), Lindström & Sturkell (1992) and Grahn & Nõlvak (1993). Three sections outside the west rim of the crater were investigated southwest of the village of Tandsbyn (Fig. 10). The first section, described by Thorslund (1940, fig. 21), is a railway-cut at Lappgrubban about 1.1 km southwest of the Tandsbyn church (Fig. 11). The second section is along the stream Ynntjärnsbäcken, about 400 m east of Lake Ynntjärn (Fig. 12), and the third section is a railway-cut, described by Simon (1987; Fig. 23), situated about 400 m west of Lake Ynntjärn (Fig. 13). All three localities exhibit the sandy resurge deposit locally known as 'Loftarstone', which is followed upwards by the first post-impact sediments, a limestone with interbedded calcareous shales. Locally the limestone contains patch reefs that grew on the crater rim. Characteristic of the chitinozoan fauna is the presence of the index species *Lagenochitina dalbyensis* (Pl. 3, fig. 3) together with *Lagenochitina* sp. A aff. *capax* in two of the localities (Lappgrubban and Ynntjärnsbäcken). This means that the Lockne event (Grahn & Nõlvak, 1993), and the first post-impact sedimentation in the crater, took place in the early Caradoc, and correspond to the lower part of the *L. dalbyensis* chitinozoan Zone. The Lockne event may be coeval with the Kärdla event, but most probably is slightly younger. The occurrence of *Belonechitina hirsuta* in the section at Ynntjärnsbäcken (Fig. 12) suggests that beds younger than those corresponding to the *L. dalbyensis* Zone are also present at this locality. Other chitinozoan species present are, for instance *Cyathochitina campanulaeformis* characteristic of the earliest Caradoc, *Calpichitina lecaniella, Desmochitina ovulum*, and *Belonechitina capitata* (Pl. 3, fig. 10). It should be noted that one boulder from the impact-related coarse breccia at Ynntjärnsbäcken yielded a chitinozoan assemblage from the upper? part of the Kukruse Stage (Zone of *Laufeldochitina stentor*, i.e. *Laufeldochitina stentor, Conochitina primitiva, C.* cf. *minnesotensis, Belonechitina capitata, Desmochitina erinacea*, and *D. minor*). Previously beds of this age had not been reported from the autochthon in Jämtland (Jaanusson & Karis 1982, fig. 1).

CONCLUSION

The immediate appearance of chitinozoans in the craters after the impact events, and their rapid evolution, have made it possible to achieve a high-resolution biostratigraphy of virtually complete sequences related to the impact craters. The planktic nature of the chitinozoophorans (Grahn, 1981) may also make it possible to date other types of natural hazards in marine environments (e.g. volcanic eruptions, earthquakes, etc.), from Ordovician through Devonian, sometimes with a precision of less than one million years. Chitinozoans occur in most lithologies, except for coarse sandstones, reef limestones, carbonate mounds and also marine redbeds and dolomites.

The environments within the craters were restricted. The first Ordovician reefs known from Baltoscandia started to grow on the rim of the Lockne crater during early Caradoc (lower Idavere Stage) times (Fig. 14). The climate was obviously warm enough for the formation of reefs, and the absence of other contemporary reefs in Baltoscandia indicates that the seas were probably deeper (Lindström 1971) than generally believed earlier. (For a summary see Jaanusson, 1982.)

Explanation of Plate 2

Selected chitinozoans from the Tvären crater. Early Caradoc. **Fig. 1.** *Conochitina tigrina* Laufeld, 1967. Tvären 2 borehole, core sample at 82.15 m. Specimen in lateral view, SEM ×150. IGR 58646 (Q. 36). **Fig. 2.** *Conochitina tigrina* Laufeld, 1967. Same specimen as in fig. 1. Detail showing the corrugations on the neck, SEM ×750. **Fig. 3.** *Calpichitina lecaniella* (Eisenack, 1965). Tvären 2 borehole, core sample at 123.02–123.05 m. Specimen in lateral view, SEM ×500. IGR 58643 (M. 36/1). **Fig. 4.** *Spinachitina tvaerenensis* sp. nov. Holotype. Tvären 2 borehole, core sample at 82.15 m. Detail of the base showing the processes. Note the mucron, SEM ×600. **Fig. 5.** *Calpichitina lecaniella* (Eisenack, 1965). Tvären 2 borehole, core sample at 123.02–123.05 m. Chain with two specimens in lateral view, SEM ×500. IGR 58643 (L. 39/4). **Fig. 6.** *Laufeldochitina stentor* (Eisenack, 1937). Tvären 2 borehole, core sample at 102.97 m. Specimen in lateral view, SEM ×100. IGR 58644 (Q. 40). **Fig. 7.** *Spinachitina tvaerenensis* sp. nov. Tvären 2 borehole, core sample at 82.15 m. Specimen in lateral view, SEM ×200. IGR 58645 (R. 38/3). **Fig. 8.** *Spinachitina tvaerenensis* sp. nov. Holotype. Tvären 2 borehole, core sample at 82.15 m. Specimen in lateral view, SEM ×350. IGR 58645 (R. 40). **Fig. 9.** *Laufeldochitina* sp. A aff. *striata*. Tvären 2 borehole, core sample at 158.77–158.80 m. Specimen in lateral view, SEM ×200. IGR 58634 (R. 37/2). **Fig. 10.** *Laufeldochitina* sp. A aff. *striata*. Tvären 2 borehole, core sample at 150.78–150.81 m. Specimen in lateral view, SEM ×100. IGR 58636 (S. 41/1).

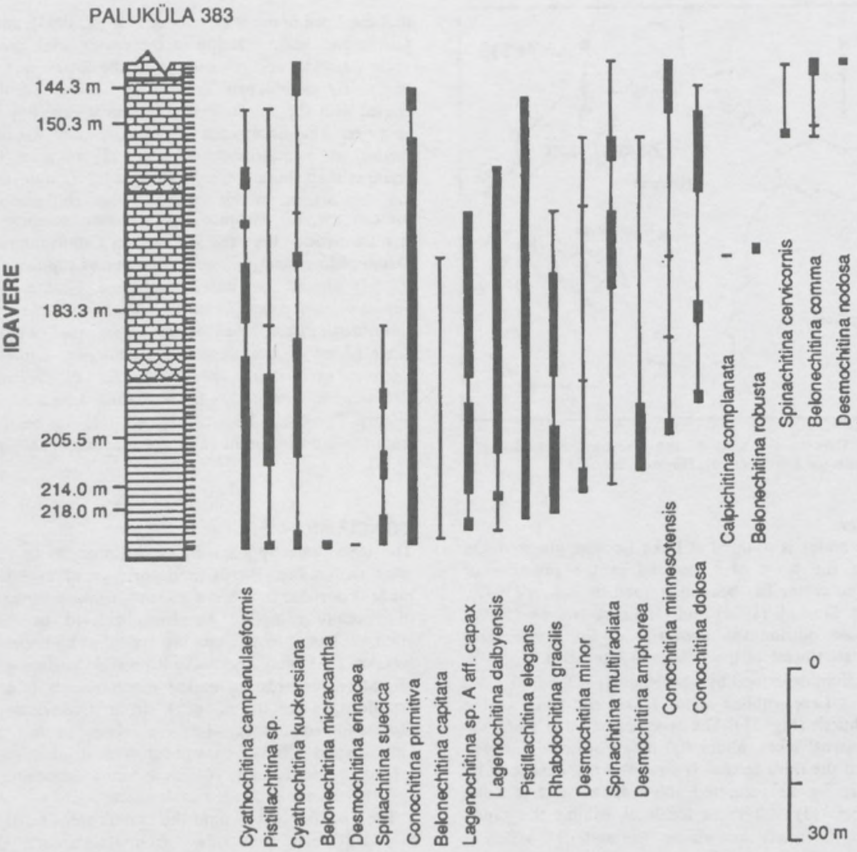

Fig. 8. Faunal log of the Paluküla 383 borehole, Kärdla crater. The depths on the left side of the column indicate the levels of discontinuity surfaces.

SYSTEMATIC DESCRIPTIONS

Cyathochitina hunderumensis sp. nov.

(Pl. 1, figs 1, 10, 11)

?1967 *Cyathochitina campanulaeformis* Eisenack; Jenkins: 456–458, pl. 71, figs 8–11.

1976 *Cyathochitina campanulaeformis* Eisenack: 187, pl. 2, fig. 4

1980 *Cyathochitina* cf. *campanulaeformis* Eisenack; Grahn: 25–27, pl. 15, figs A–D.

1984 *Cyathochitina campanulaeformis* Eisenack; Grahn: 16–17.

Derivation of name. Latin, *hunderumensis*, from the late Arenig substage of Hunderum where the species is common.

Diagnosis. Small *Cyathochitina* with a short thickened carina and a basal scar on the apex of the vesicle.

Holotype. Pl. 1, fig.1. IGR 58614 (L. 39/4).

Type locality. Fylla 3 borehole, core sample at 254.30 m, Granby crater, Östergötland, south Sweden.

Description. A small species of *Cyathochitina*. Vesicle smooth with a characteristic bell-like to conical shape. The maximum width is at the margin. The margin is provided with a short thickened carina. A basal scar is present. The neck is cylindrical, and shorter than half the total length. Aperture straight.

Dimensions. The dimensions given by Grahn (1980, p. 27) are characteristic also for the specimens from the Granby crater. He noted that also the main parts of the populations fall within the length:width ratio 1.25–2:1. From the type level 30 flattened specimens were measured. A coefficient of 0.7 was used to restore the diameter of chamber and neck. The total length is 156–280 μm (holotype 233 μm, mean value 199 μm),

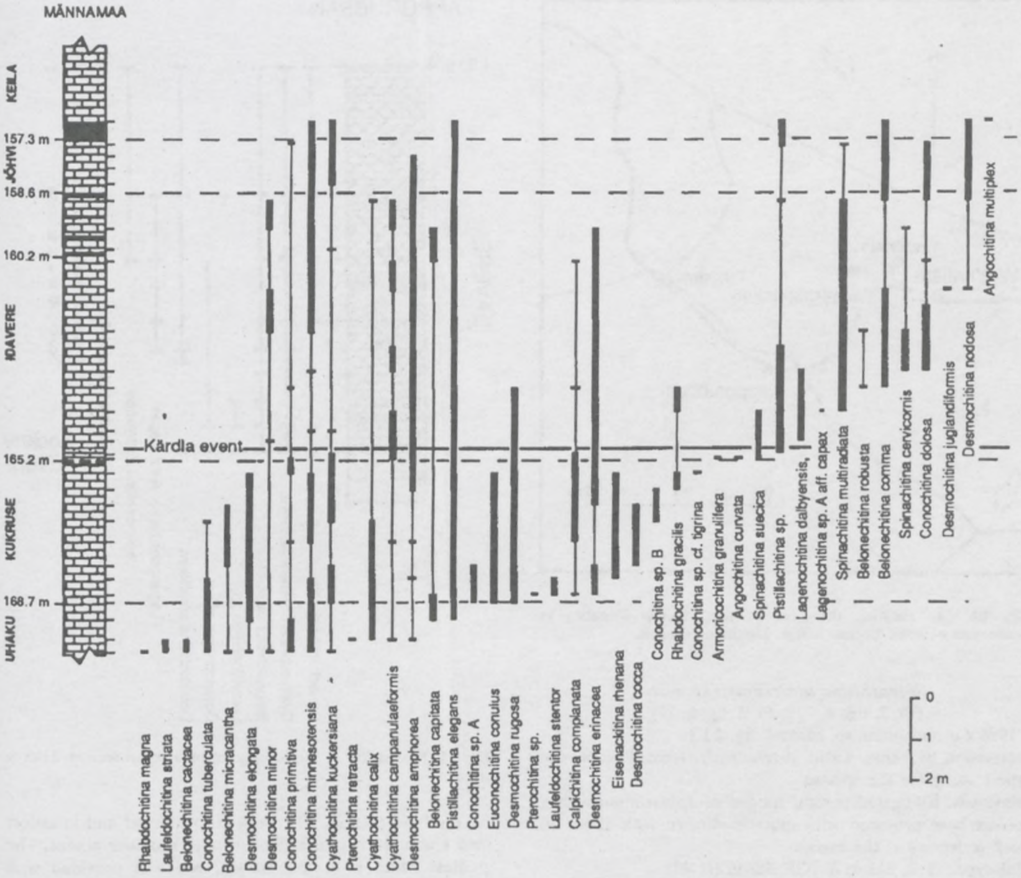

Fig. 9. Faunal log of the Männamaa borehole, Kärdla crater. The depths on the left side of the column indicate the levels of discontinuity surfaces.

maximum width 73–124 μm (holotype 123 μm, mean value 94 μm), width of neck 42–53 μm (holotype 48 μm, mean value 47 μm) and the length of the neck 50–94 μm, (holotype 90 μm, mean value 70 μm). The ratio of vesicle length/chamber diameter for specimens from the type level is shown in Fig. 15, and the ratio of chamber diameter/neck diameter in Fig. 16.

Discussion. The small *C. hunderumensis* sp. nov. are easily distinguished from typical specimens of *Cyathochitina campanulaeformis* as defined by Eisenack (1931, 1962) through its small size, and because of the short and thickened carina. *Cyathochitina varennensis* Paris 1981, from the early Llanvirnian of western France, has a similar size. However, the ratio of length of the neck/length of the vesicle is significantly different.

Occurrence. *Cyathochitina hunderumensis* sp. nov. ranges from the upper Volkhov Stage (chitinozoan Zone of *Conochitina cucumis* Nõlvak & Grahn, 1993) to the top of Kunda Stage (upper Arenig–lower Llanvirn). The species is known from Öland (Eisenack, 1976; Grahn, 1980) and Närke (Grahn, unpublished data), Sweden and from the Granby crater (this paper). It occurs also in North Estonia (reported but not illustrated from the Suhkrumägi section in Tallinn by Grahn 1984 as *Cyathochitina campanulaeformis*). It should be noted that specimens assignated to *Cyathochitina campanulaeformis* first occur in the Aseri Stage, after the last occurrence of *C. hunderumensis* sp. nov. Specimens similar to *C. hunderumensis* sp. nov. have been reported from the lower Hope Shales (early Llanvirn) in Shropshire, England (Jenkins, 1967).

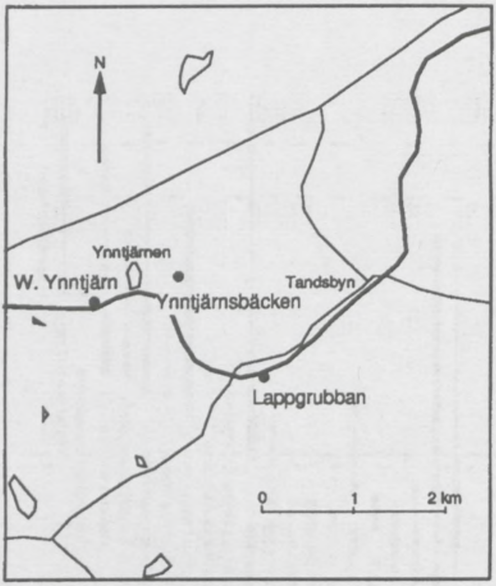

Fig. 10. Map showing the sites of the outcrop localities in connection with the Lockne crater, Jämtland, Sweden.

Spinachitina tvaerenensis sp. nov.

(Pl. 2, figs 4, 7, 8; Pl. 3, fig. 6, 11)

? 1986 *Coronochitina* sp. Männil, fig. 2.1.1.

Derivation of name. Latin, *tvaerenensis*, from Tvären, the type locality for the species.

Diagnosis. Elongated conical species of *Spinachitina* with a convex base provided with mucron. Crown with about 20 thick processes at the margin.

Holotype. Pl. 2, figs 4, 8. IGR 58645 (R 40).

Type locality. Tvären 2 borehole, core sample at 82.15 m, Tvären crater, Södermanland, south Sweden.

Description. This elongated conical *Spinachitina* species is characterized by the conical expansion of the chamber close to the margin. The neck is cylindrical and indistinct with a straight aperture. Flexure and shoulder absent. The greatest width is at the margin, which is provided with about 20 robust short and simple conical processes, elongated parallel to the vesicle axis with their proximal end (insertion zone). A mucron is present. The vesicle wall is smooth aperturewards the margin.

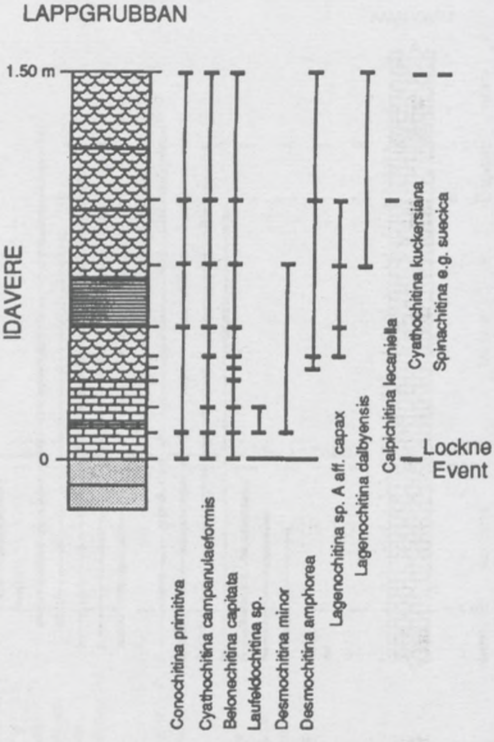

Fig. 11. Faunal log of the Lappgrubban outcrop section, Lockne crater.

Explanation of Plate 3

Selected chitinozoans from the Granby (Late Arenig), Tvären (Early Caradoc) and Lockne (Early Caradoc) craters. **Fig. 1.** *Lagenochitina* sp. A aff. *capax*. Tvären 2 borehole, core sample at 144.99–145.00 m. Specimen in lateral view, SEM ×500. IGR 58637 (L. 32/4). **Fig. 2.** *Lagenochitina* sp. A aff. *capax*. Tvären 2 borehole, core sample at 144.99–145.00 m. Specimen in lateral view, SEM ×500. IGR 58637 (P. 35/3). **Fig. 3.** *Lagenochitina dalbyensis* (Laufeld, 1967). Ynntjärnsbäcken, basal 7 cm of the upper Dalby Limestone. Specimen in lateral view, SEM ×300. IGR 58632 (L. 37/3). **Fig. 4.** *Conochitina tigrina* Laufeld, 1967. Tvären 2 borehole, core sample at 82.15 m. Specimen in lateral view, SEM ×200. IGR 58646 (R. 38/1). **Fig. 5.** *Conochitina* sp. 1. Fylla 3 borehole, core sample at 258.00 m. Specimen in lateral view, SEM ×350. IGR 58610 (Q. 38). **Fig. 6.** *Spinachitina tvaerenensis* sp. nov. Tvären 2 borehole, core sample at 82.15 m. Detail of the basal part showing the processes, SEM ×350. IGR 58646 (O. 34/3). **Fig. 7.** *Conochitina minnesotensis* (Stauffer, 1933). Tvären 2 borehole, core sample at 144.40 m. Specimen in lateral view, SEM ×150. IGR 58639 (N. 45/2). **Fig. 8.** *Rhabdochitina gracilis* Eisenack, 1962. Fylla 3 borehole, core sample at 258.00 m. Specimen in lateral view, SEM ×100. IGR 58610 (O. 38). **Fig. 9.** *Lagenochitina esthonica* Eisenack, 1955. Fylla 3 borehole, core sample at 254.30 m. Specimen in lateral view, SEM ×80. IGR 58614 (L. 40/3). **Fig. 10.** *Belonechitina capitata* (Eisenack, 1962). West Ynntjärn, basal 10 cm of the upper Dalby Limestone. Specimen in lateral view, SEM ×300. IGR 58622 (M. 36). **Fig. 11.** *Spinachitina tvaerenensis*? sp. nov. Tvären 2 borehole, core sample at 82.15 m. Specimen in lateral view, SEM ×300. IGR 58646 (O. 37/3).

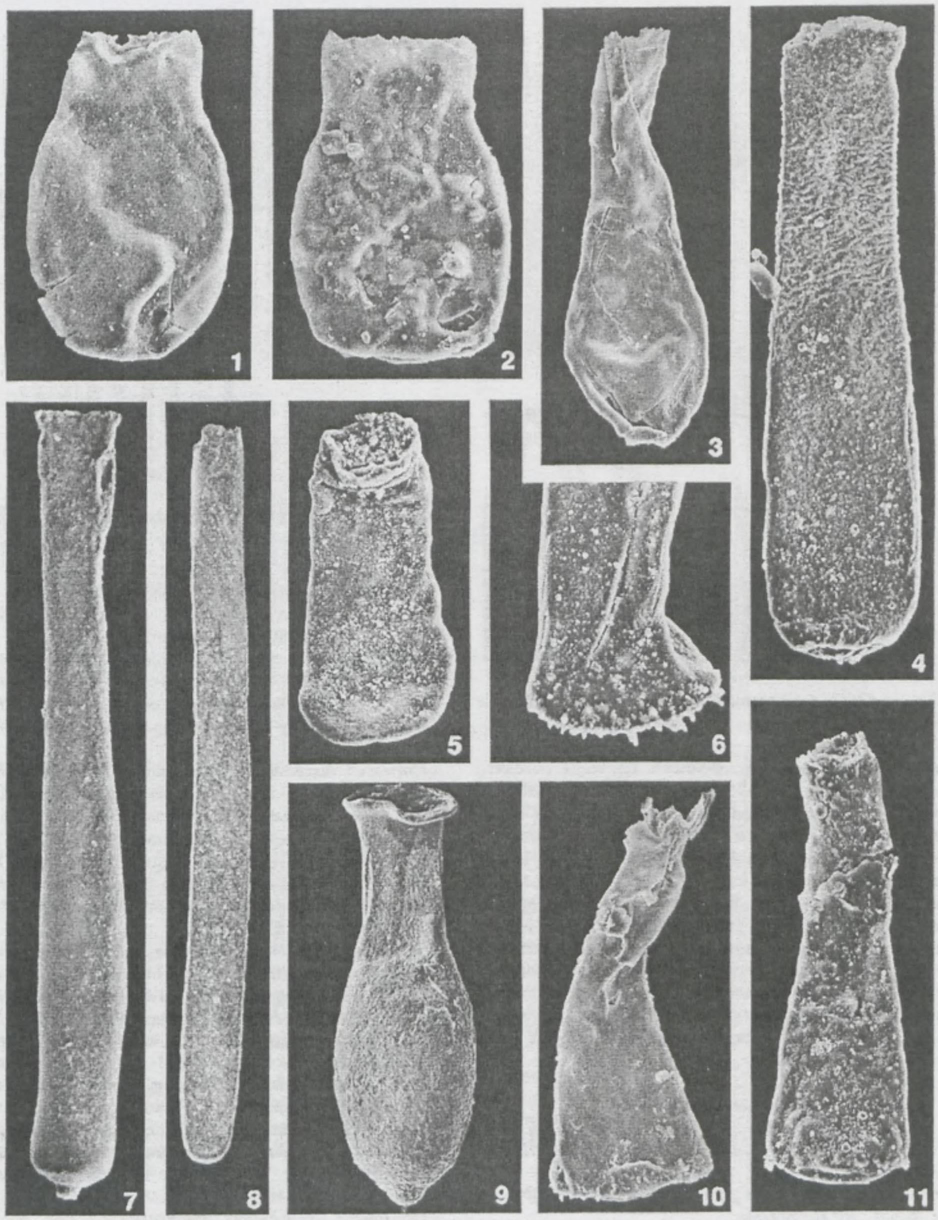

Plate 3

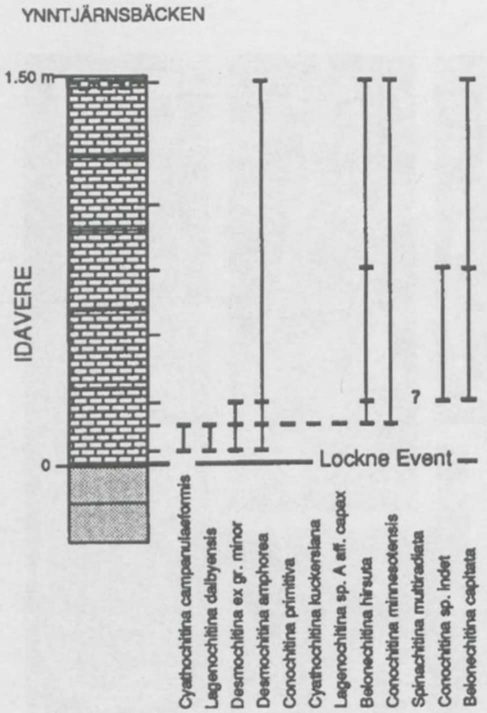

Fig. 12. Faunal log of the Ynntjärnsbäcken outcrop section, Lockne crater.

Corrected dimensions. (7 specimens, flattening corrected by a coefficient of 0.7.) Total length 247–568 μm (holotype >257 μm, broken neck), max width 73–117 μm (holotype 76 μm), aperture 41–78 μm (holotype, width of neck 43 μm), spines about 7 μm (holotype 7 μm).

Occurrence. *Spinachitina tvaerenensis* sp. nov. has a restricted range in the uppermost Kukruse Stage. It probably corresponds to *Coronochitina* sp. by Männil (1986), who indicated a range from uppermost Kukruse to lowermost Idavere Stage, where it disappears before the first occurrence of the index species for the chitinozoan Zone of *Armoricochitina granulifera* Nõlvak & Grahn 1993 (*Cyathochitina* cf. *reticulifera* by Männil, 1986) in the lowermost Idavere Stage.

Tanuchitina granbyensis sp. nov.
(Pl. 1, figs. 2–5)

V. ? 1981 *Tanuchitina* sp. aff. *achabae* Paris: 216–217, pl. 40, figs 14, 15.

Derivation of name. Latin, *granbyensis*, from Granby, the type locality for the species.

Diagnosis. A long subcylindrical species of *Tanuchitina* with its carina erected on an ovoid base.

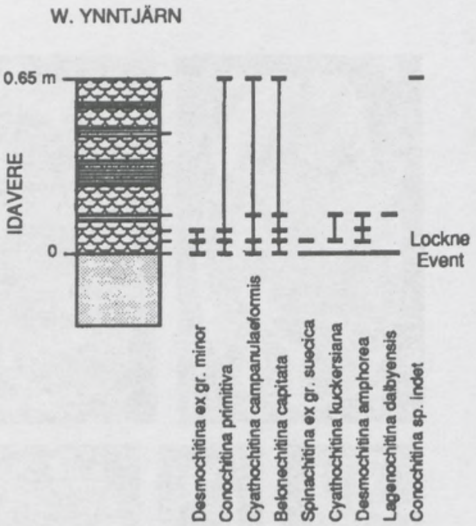

Fig. 13. Faunal log of the outcrop section west of Ynntjärn, Lockne crater.

Holotype. Pl. 1, figs 2, 3. IGR 58601 (D 43).

Type locality. Fylla 9 borehole, core sample at 272.20 m, Granby crater, Östergötland, south Sweden.

Description. A very long slender, almost cylindrical *Tanuchitina* with a smooth vesicle. Aperture straight. Greatest width about one quarter aperturewards from the ovoid base. Fairly long membranaceous carina surrounding the apex.

Corrected dimensions. (27 specimens, flattening corrected by a coefficient of 0.7.) Total length >672–1533 μm (holotype 1466 μm, mean value 1262 μm), max. width 82–138 μm (holotype 117 μm, mean value 97 μm), aperture 60–118 μm (holotype 85 μm, mean value 71 μm), and carina 16–34 μm (holotype 30 μm). The ratio vesicle length/chamber diameter for specimens from the type level is shown in Fig. 17.

Discussion. The length of this species makes it easily distinguishable from any other early Ordovician *Tanuchitina* species described to date. *Tanuchitina achabae* from the middle Arenig of western France (Paris, 1981) is half the size in terms of vesicle length. *Tanuchitina* sp. aff. *achabae* from the late Arenig of western France is probably a synonym to *T. granbyensis* sp. nov. They are of a similar length and the vesicle is frequently curved along its longitudinal axis. *T. granbyensis* sp. nov. may be confused with *Rhabdochitina gracilis* Eisenack when the carina is strongly eroded. The latter is also commonly curved along the long axis of the vesicle.

Occurrence. *Tanuchitina granbyensis* sp. nov. has so far only been found in late Arenig strata in the Granby crater, Östergötland, south Sweden. It is probably present

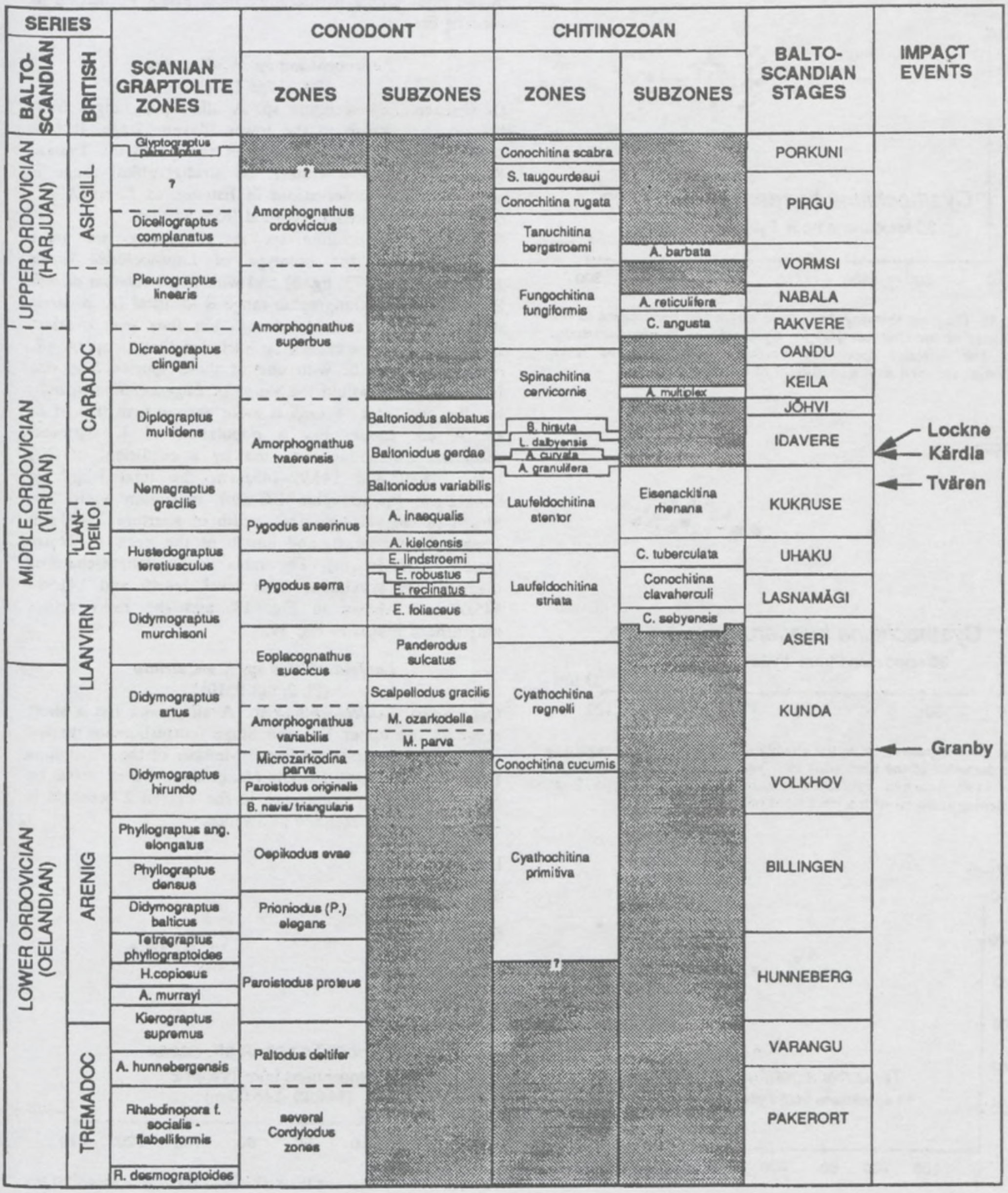

Fig. 14. Correlation table for the impact craters in the Ordovician of Baltoscandia.

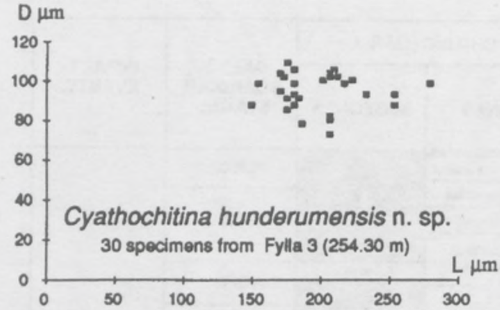

Fig. 15. Diagram showing the vesicle length (L) with regard to the diameter of the chamber (D) for *Cyathochitina hunderumensis* sp. nov. (30 flattened specimens measured from the type level, flattening restored with a coefficient of 0.7.)

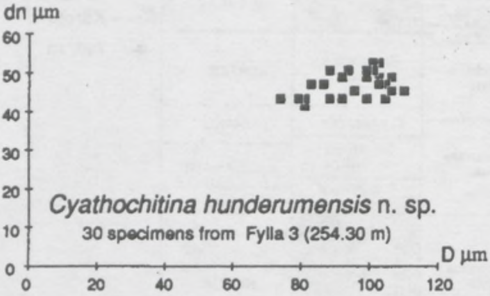

Fig. 16. Diagram showing the chamber diameter (D) with regard to the diameter of the neck (dn) for *Cyathochitina hunderumensis* sp. nov. (30 flattened specimens measured from the type level, flattening restored with a coefficient of 0.7.)

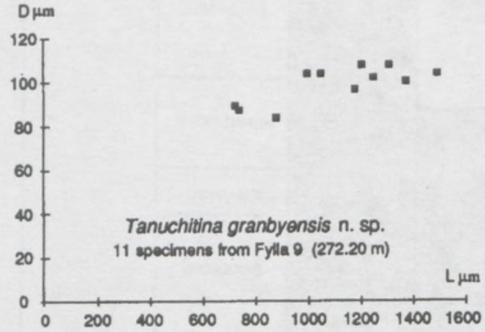

Fig. 17. Diagram showing the vesicle length (L) with regard to the diameter of the chamber (D) for *Tanuchitina granbyensis* sp. nov. (11 flattened specimens measured from the type level, flattening restored with a coefficient of 0.7.)

in the late Arenig in the lowermost Pissot Formation in western France.

Lagenochitina sp. A aff. *capax*

(Pl. 3, figs 1, 2)

Occurrence. *Lagenochitina* sp. A aff. *capax* ranges from the upper Kukruse to the lower Idavere Stage. It is a common species in the sequences related to the Tvären, Kärdla and Lockne craters. Its stratigraphical range is concluded from observations in Estonia, as *L.* sp. A aff. *capax* is not previously recorded from Sweden.

Remarks. *Lagenochitina* sp. A aff. *capax* has great similarities with the holotype of *Lagenochitina capax* Jenkins 1967 (pl. 73, fig. 3) and with *Lagenochitina deunffi* Paris 1974. The stratigraphic range is identical (*L. capax* is slightly younger than *L. deunffi* but they may overlap; Paris 1981), and it cannot be excluded that *L.* sp. A aff. *capax* is conspecific with one of these species. The size falls completely within the range of *Lagenochitina deunffi*, but the neck of *L. deunffi* is more narrow than that of *L.* sp. A aff. *capax*. For a population of 40 flattened specimens (flattening corrected by a coefficient of 0.7) from 144.40 and 144.99–145.00 m the total length is 88–122 µm (mean value 103 µm), maximum width 38–51 µm (mean value 44 µm), width of aperture 23–33 µm (mean value 30 µm), and length of the neck 13–22 µm (mean value 16 µm). The ratio vesicle length/chamber diameter for specimens from level 144.40 and 144.99–145.00 m is shown in Fig. 18, and the ratio vesicle length/neck length in Fig. 19.

Laufeldochitina sp. A aff. *striata*

(Pl. 2, figs 9, 10)

Occurrence. *Laufeldochitina* sp. A aff. *striata* has a short range in the upper Kukruse Stage (corresponding to the middle part of the lower Peetri Member of the Viivikonna Formation) in North Estonia (*Laufeldochitina* cf. *striata* by Männil, 1986). Its occurrence in the Tvären 2 borehole is the first safely established in Sweden.

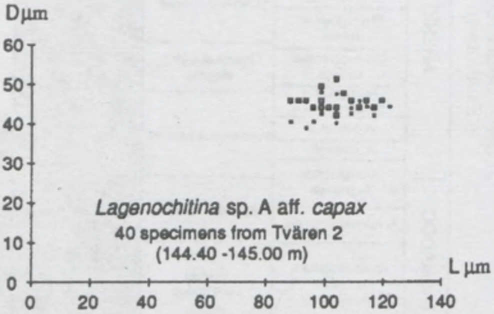

Fig. 18. Diagram showing the vesicle length (L) with regard to the diameter of the chamber (D) for *Lagenochitina* sp. A aff. *capax*. (40 flattened specimens measured from levels 144.40 m (black dot) and 144.99–145.00 m (squares) in the Tvären 2 borehole, flattening restored with a coefficient of 0.7.)

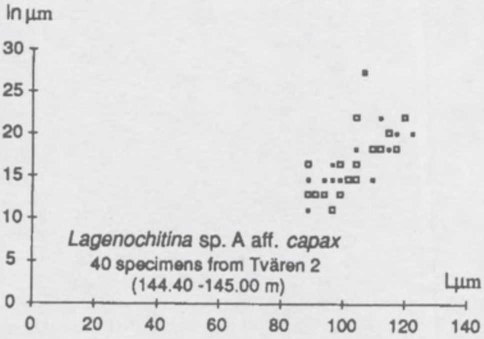

Fig. 19. Diagram showing the vesicle length (L) with regard to the length of the neck (ln) for *Lagenochitina* sp. A aff. *capax*. (40 flattened specimens measured from levels 144.40 m (black dot) and 144.99–145.00 m (squares) in the Tvären 2 borehole, flattening restored with a coefficient of 0.7.)

Remarks. Characteristic *Laufeldochitina* sp. A aff. *striata* differ from *Laufeldochitina striata* (Eisenack, 1937) in possessing a predominantly smooth wall. A striate ornamentation restricted to the basalmost part may occur on some specimens (Pl. 2, fig. 9). The dimensions fall within the range of *Laufeldochitina striata*.

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Paper X

Paris, F., Verniers, J., Achab, A., Albani, R., Ancilletta, A., Asselin, E., Chen, X., Fatka, O., Grahn, Y., Molyneux, S., Nõlvak, J., Samuelsson, J., Sennikov, N., Soufiane, A., Wang, X. & Wincester-Seeto, T. 1999.
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Correlation of Ordovician regional chitinozoan biozonations

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INTRODUCTION

A working group of the Subcommission on Chitinozoa of the "Commission Internationale de Microflore du Paléozoïque" (C.I.M.P.), has been convened to ascertain the correlation of the different regional chitinozoan biozonations for the Ordovician System. This work has been done under the auspices of the Chitinozoa Clade Team of IGCP n° 410, and in connection with the activities of the Subcommission on the Ordovician Stratigraphy (SOS). It is not our aim to propose chitinozoans as a substitute for the other fossil groups, such as the graptolites or conodonts, which are currently used to define the GSSPs of the new Ordovician series and stages. We are merely seeking possible ties between the different regional chitinozoan biozonations in a first attempt to establish a global chitinozoan biozonation for the Ordovician, as has already been achieved for the Silurian (Verniers *et al.* 1995) and the Devonian (Paris *et al.* in print).

The difficulties encountered in the Ordovician are considerable as demonstrated by the debate that has been raging for 20 years on the choice of index species, and the definition of the GSSPs for the Ordovician series and stages. The reasons for these difficulties are well known: the Ordovician is the Palaeozoic system where the palaeocontinents were the most dispersed, and this specific palaeogeographic configuration had important consequences for the provincialism of the Ordovician faunas. On one hand, the distribution of benthic forms is influenced by the presence of large oceans acting as barriers resulting in a high degree of provincialism. On the other hand, the distribution of pelagic faunas is mostly controlled by temperature and indirectly by palaeolatitude and hence by climate. This pattern of distribution is subsequently modified by palaeocurrents. The chitinozoans, or more precisely their parent organisms (see Paris and Nõlvak in print), are generally agreed to have a pelagic mode of distribution (see references in Miller 1996; Paris 1996) and are therefore also partly controlled by the palaeolatitudes, via the climate and palaeocurrents (Achab 1991; Paris 1993).

REGIONAL ORDOVICIAN CHITINOZOAN BIOZONATIONS

At present, three main regional chitinozoan biozonations are used for the Ordovician (see references in Paris 1996).

In eastern Canada twenty two biozones were defined by Achab (1989), with additional information in Achab and Asselin (1995). This biozonation can be applied to Laurentia (North America, Greenland, Spitzbergen). This huge palaeoplate was situated at low latitude, or straddling the equator, during the Ordovician (Scotese and McKerrow 1990, and references in Achab 1991).

In Scandinavia and the Baltic area, i.e., part of Baltica palaeoplate, fifteen biozones and eight subzones were defined by Nõlvak and Grahn (1993) based on numerous data from Baltoscandia (references in Grahn 1990) and the Russian Platform (references in Umnova 1981; Sennikov_and Obut 1997) (Fig. 1). Baltica and East Avalonia at that time were situated at rather intermediate latitudes (Scotese and McKerrow 1990). There are, however, a number of conflicting models on the mode and timing of the collision of Avalonia with Baltica (Cocks and Fortey 1982; Vannier *et al.* 1989; Prigmore *et al.* 1997). This is not a simple matter of precision in palaeogeographical reconstruction. It may have an influence on the correlations we want to document for the different regional biozonations, due to the intermediate and therefore strategic position of Avalonia between north Gondwana, Baltica and Laurentia.

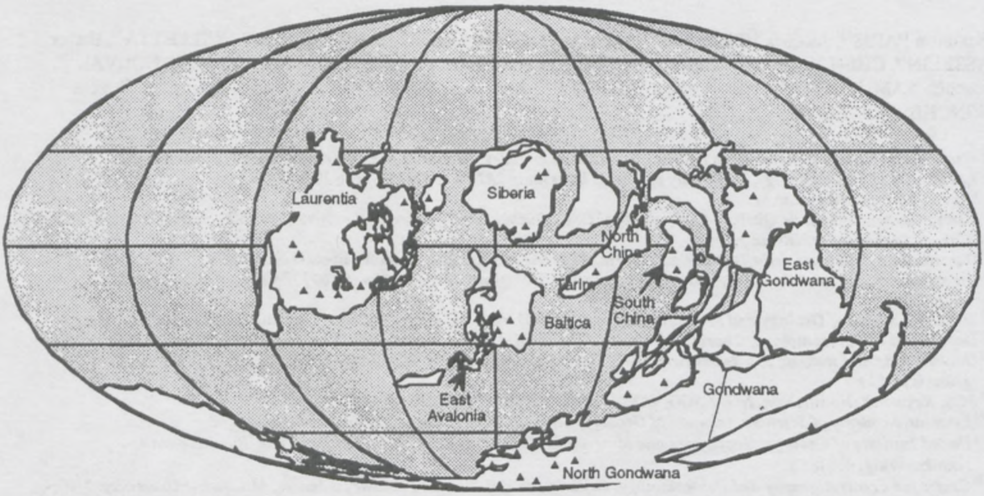

Fig. 1. Ordovician chitinozoan bearing localities on a Late Ordovician palaeogeographic reconstruction (based on Scotese and McKerrow 1990). Several localities may be represented with a single symbol in regions with numerous occurrences.

A third regional biozonation was described for northern Gondwana, i.e., from West Africa to the Middle East and from south of the Sahara to southern Europe. Initially twenty two biozones were defined (Paris 1990, and references), and now refined with twenty eight biozones for the Ordovician based on recent data (Fatka 1993; Al-Hajri 1995; Oulebsir and Paris 1995). Northern Gondwana regions are usually regarded as having been located at high latitudes during the Ordovician (Scotese and McKerrow 1990; Paris and Robardet 1990). This is well supported by end-Ordovician glacial evidence, e.g., in North Africa (see references in Robardet and Doré 1988).

Additional data are available now from other palaeoplates (Fig. 1) where new regional biozones are being erected; these will be progressively incorporated into a biozonation for the whole Ordovician system. This is the case for North China and Tarim, and for South China (Chen *et al.* 1996; Chen and Wang 1996, and references) which have been referred to low latitudes or to an equatorial position (Scotese and McKerrow 1990; Wang 1989).

Information is also available from the Australian part of eastern Gondwana (see references in Foster *et al.* 1999) which was located close to the equator during the Ordovician (Scotese and McKerrow 1990; references in Webby *et al.* 1999). At present, limited data on Ordovician chitinozoans are available from the South American part of western Gondwana (Grahn 1992; Heuse *et al.* in print). For the Siberia palaeoplate, situated at fairly low latitude during the Ordovician (Scotese and McKerrow 1990), published data are more and more numerous (Zaslavskaya 1984; Zaslavskaya *et al.* 1978). However, no detailed biozonation has been proposed yet for the Ordovician chitinozoans of this palaeoplate.

CHITINOZOAN TIES BETWEEN REGIONAL BIOZONES

When using fossils to document long distance correlation, one must accept that the geographical spread of a given species must be synchronous at the scale of geological time. This assumption seems acceptable for the Ordovician period as a slight difference in the first occurrence of a taxon in different remote sites will not be discernible because the available dating tools are not accurate enough to discriminate a duration of less than several hundred thousands to about one million years.

In order to prevent a circular reasoning, we have decided to document the correlation between the available chitinozoan biozones calling exclusively upon the distribution of chitinozoan taxa shared by two or several palaeogeographical provinces. This means that in the

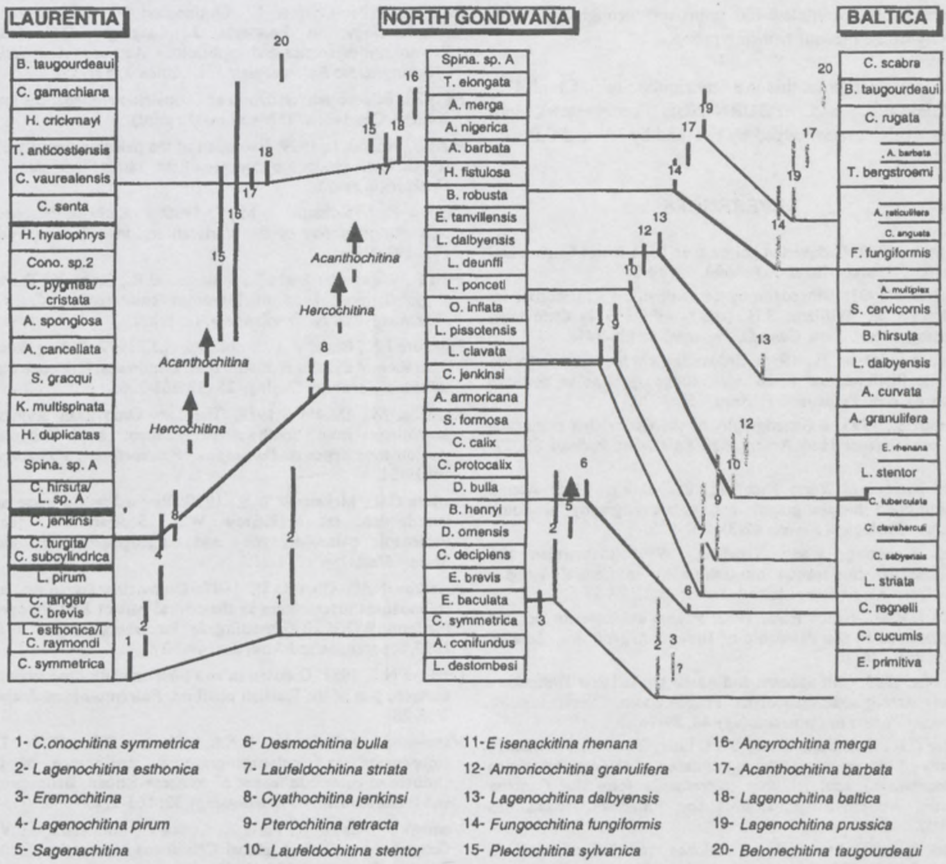

Fig.2. The three regional Ordovician chitinozoan biozones (based on Achab 1989, Paris 1990, and in print, Nõlvak and Grahn 1993) and tentative correlation based exclusively on chitinozoan data.

Among the Ordovician chitinozoans some geographically widespread taxa are long-ranging species (e.g., *Desmochitina minor, Belonechitina micracantha, Cyathochitina campanulaeformis*). Consequently, they are useless for establishing correlation between the different regional biozonations. Other species, also recorded on several palaeoplates where they coexist with endemic species, are relatively short ranging taxa, e.g., from the oldest to the youngest (Fig. 2): *Conochitina symmetrica, Lagenochitina esthonica, Lagenochitina pirum, Desmochitina bulla, Laufeldochitina striata, Cyathochitina jenkinsi, Pterochitina retracta, Laufeldochitina stentor, Eisenackitina rhenana, Armoricochitina granulifera, Lagenochitina dalbyensis, Fungochitina fungiformis, Plectochitina sylvanica, Acanthochitina barbata, Lagenochitina prussica, Lagenochitina baltica, Plectochitina sylvanica, Belonechitina taugourdeaui.*

These, and other similar taxa, will be used herein to document ties between two or more biozonations. The first appearance of peculiar morphological innovations which are discriminant at generic level should be also be taken into account (e.g. the perforated carina of *Sagenachitina*). The most classic pit-fall when making intercontinental correlation is the use of miss-identified taxa. To prevent this problem, we have initiated the exchange of material between participants for direct comparison. This procedure seems necessary because some widely used species were poorly described and did not have the benefit of SEM illustration, nor of biometric analysis when first described.

CONCLUSIONS

Our attempt to correlate chitinozoan regional biozonations based exclusively on Ordovician chitinozoan ties is summarised herein on Fig. 2. One should note the existence of fairly good ties between the Ordovician chitinozoan biozonations of Baltica and northern Gondwana. Correlation with Laurentia is weaker, especially for Middle Ordovician time. This preliminary scheme will be progressively completed and improved through the future activity of the present working group.

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Paper XI

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Carbon isotope excursions and coeval environmental and biotic changes in the late Caradoc and Ashgill of Estonia

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INTRODUCTION

The Ordovician was a dynamic period with rapid changes in biota and environment. Our study embraces the time interval beginning with the formation of the well-known Kinnekulle K-bentonite Bed in the middle Caradoc (bottom of the Keila Stage) until the end of the period. So far two event levels have been described from this interval – a late Caradoc *(sensu lato)* biotic (Sepkoski 1995; Kaljo *et al.* 1996) and isotopic event (Ainsaar *et al.* 1999; Patzkowsky *et al.* 1997; Pancost *et al.* 1998, etc.) and the well-known late Ashgill series of severe extinctions, together with environmental processes at least partly caused by a Hirnantian short-term glaciation and marked by significant carbon and oxygen isotope excursions (Brenchley *et al.* 1994; Kump *et al.* 1995; Marshall *et al.* 1997; Kaljo *et al.* 1998b; Ripperdan *et al.* 1998). A nearly 10 Ma interval between these event levels remains uncharacterized in respect of isotope events. Yet it is clearly essential for understanding the environmental history of the late Ordovician time.

Our data are preliminary since they come from only a few borehole cores of Estonia and additional sections should be studied for control. Most important for the study is the Rapla core situated in the North Estonian Confacies Belt (Fig.1) and penetrating the whole interval discussed. In addition to the Rapla the upper part of the interval was studied in the Kaugatuma core, for the lower part we used the data from the Tartu core by Ainsaar *et al.* (1999). The Hirnantian part of the section has been studied in a number of cores, but these data are available elsewhere (Kaljo *et al.* 1998b) and are not treated in detail here.

The stratigraphical terminology used is shown in Fig. 2. The lithology and subdivisions of the Rapla core are given according to Põlma (1972) with some later corrections (Kaljo *et al.* 1996).

In very general terms the North Estonian Confacies Belt represents the shallow to mid-shelf or ramp area of the basin, while the Livonian Tongue of the Central Baltoscandian Confacies Belt (Fig.1) forms a deeper shelf depression or lower ramp. Correspondingly, in the north mostly micritic and argillaceous limestones with different skeletal debris occur. Beginning with the Rakvere Stage, up to the top of the section algae dominate among the skeletal component of the rocks. The clay content is rhythmically changing, allowing us to distinguish a series of cycles, which in general are transgressive or regressive (Fig.2). In the upper part of the Ashgill (upper Pirgu and Porkuni stages), packstones with abundant echinoderm debris, and biohermal rocks indicate a general shallowing of the sea.

In South Estonia more argillaceous rocks prevail – usually marlstones with argillaceous limestones, at some levels dark argillites or even black shales. In the Livonian Tongue area (especially on the territory of Latvia) the late Ordovician sequence seems to be more complete than in the Rapla core, where several discontinuity surfaces indicate the presence of more or less extensive gaps.

Fig. 1. Location of boreholes and facies belts.

CARBON ISOTOPE EXCURSIONS

Carbon isotope data were obtained from whole rock samples analysed according to the methods explained in Kaljo *et al.* (1998a). The results of the analyses are presented in Fig. 2. Below we shall give several comments on the figure, where chitinozoan distribution in the Rapla core is used as a basis for correlation.

(1) The samples from the Keila and Oandu stages in the Rapla core show low $\delta^{13}C$ values, mostly in the limits of 0...1‰. In the Tartu core (Ainsaar *et al.* 1999) just above the Kinnekulle K-bentonite Bed the carbon isotope curve (Fig.2) is very similar to the Rapla curve, but in the upper part of the Keila Stage a positive shift until 2.2‰ is observed whereas in the Oandu Stage it remains on the

level of 1.2...1.6‰. This mid-Caradoc positive $\delta^{13}C$ excursion might be correlated into a gap in the Rapla core marked by a strong discontinuity surface in the top of the Keila Stage. Signs of erosion are recorded also in the middle of the Oandu Stage which together with highly reduced thickness are evidence of hiatuses in this part of the section in North Estonia. Our carbon isotope observations together with chitinozoan data support the conclusion about the gap drawn by Ainsaar *et al.* (1999).

(2) The Rakvere Stage in the Rapla core shows a distinct positive excursion ($\delta^{13}C$ max 1.9‰) in its lower part and a following relatively abrupt drop of the curve until –0.5‰ at a multiple discontinuity surface (at a depth of about 115 m). This Rakvere (or late Caradoc) excursion corresponds to the nearly entire range of the *Cyathochitina angusta* chitinozoan Subzone, represented in the Tartu core only by its uppermost part. Lower beds are missing due to a gap in boring.

(3) The carbon isotope curve in the Nabala Stage of the Rapla core seems to be very changeable between 0...2‰ with a small peak in the lower part of the Nabala Stage (Fig.2), but since we have no additional control data, we refrain from further comments.

(4) In the Vormsi Stage, the curve is, unlike that in the underlying stage, more steadily oriented – in the lower half it shows a negative shift in $\delta^{13}C$ until –0.8‰ and then returns to positive values (1.8‰) at the very beginning of the Pirgu Stage. The low stand in the Vormsi Stage could be named as an early Ashgill carbon isotope negative excursion.

(5) The $\delta^{13}C$ curve in the Pirgu Stage begins with a low positive excursion (note a 7 m gap in the sampling record), then follows steady lowering of the $\delta^{13}C$ values until –1‰ in the upper part of the stage. The lower Pirgu positive shift (max 2.5‰) is better represented in the Kaugatuma core (Fig.2); higher the course of the curve is very similar to that in the Rapla core. The lows of the curve in the upper part of the stage in both cores are in the *Conochitina rugata* chitinozoan Zone (Nõlvak and Grahn 1993).

(6) A high positive Hirnatian $\delta^{13}C$ peak reaches 5.6‰ in the Rapla core, 4.4‰ at Kaugatuma. The rise of the values more or less coincides with the *Spinachitina taugourdeaui* chitinozoan Zone in both sections. See Kaljo *et al.* (1998b) for further detail.

CORRELATION OF DIFFERENT EVENTS AND DISCUSSION

Microfossil diversity has proved a sensitive tool for identification of environmental changes. In the Rapla core the corresponding data were summarized for acritarchs and chitinozoans (Kaljo *et al.* 1996) and for scolecodonts (Hints 1998). These data show:

(1) The Keila Stage in the Rapla core represents a maximum diversity episode of acritarchs and a long-term decline of chitinozoans, but at the end of the period both groups experience serious extinction. The scolecodonts show only slight increase in disappearance rate in the Keila Stage of the Rapla core. In the summarized distribution data set from all localities this tendency is much better expressed being one of the most distinct changes in Ordovician jawed polychaete diversity dynamics. As a result, the Oandu microfossil assemblage is very scarce and the corresponding time is called the Oandu crisis (Kaljo *et al.* 1996). This mid-Caradoc diversity drop seems to coincide with an extinction episode noted by Sepkoski (1995).

(2) In the middle part of the section embracing the Rakvere, Nabala and in some aspects (disappearance rate of acritarchs, appearance rate of chitinozoans) also Vormsi Stage the diversity of the named microfossils is variable without any clear tendency. Some correlation with certain facies changes can be noted, but it is not enough consistent.

(3) Beginning mostly with the Vormsi Stage, i.e. during the whole Ashgill, the extinction rate of acritarchs and chitinozoans increases step by step, reaching the highest level in late Pirgu time. Only the total rate of disappearances of acritarchs is highest in early Pirgu time. The scolecodonts, on the other hand, display clearly increasing taxonomic diversity, which achieves its maximum in the Pirgu Stage. This contrast is likely a result of a relatively high number of long-ranging species and an evolutionary radiation period of jawed polychaetes. The latter begins already in Oandu time.

(4) Distribution of chitinozoans in the Tartu core (Bauert and Bauert 1998) shows a remarkable extinction event close to the upper boundary of the Keila Stage, where 55% of the taxa disappeared (in Fig. 2 at the end of the range of *Spinachitina cervicornis*) and a diversity low in the following Oandu interval. From these data we may conclude that the mid-Caradoc $\delta^{13}C$ positive excursion correlates in the Rapla core with a gap above the Keila rocks (Fig.2) and seems to coincide with the end of the diversity maximum and extinction event.

In general pattern of the Hirnantian (Porkuni) major carbon isotope excursion is different, because the main episode of microfossil extinction occurred earlier in late Pirgu time, or at the very beginning of the Porkuni time.

Between the above-mentioned two event levels both microfossil diversity and the carbon isotope curve are very variable. This similarity in characters of diversity and curve might not be incidental, but not clearly understood yet.

Tab. 1. Ratio of disappearing and appearing macrofossil taxa in regional stages (stratigraphical indexes see Fig.2)

Groups	D_{II}	D_{III}	E	$F_{I}a$	$F_{I}b$	$F_{I}c$	F_{II}
All groups	2.15	1.52	1.43	0.53	1.10	2.54	2.75
Brachiopods	1.17	0.94	2.00	0.30	0.70	1.88	0.70
Trilobites	2.40	1.00	0.60	0.75	1.00	3.00	0.80

Macrofossil diversity dynamics cannot be studied in a core section, but data from all localities are very much in line with those discussed above. Summarizing published data (Hints and Rõõmusoks 1997; Rõõmusoks 1997) and comparing ratios of disappearing and appearing taxa (Tab. 1) we can see the same pattern: the disappearance prevails clearly over the appearance of brachiopod and trilobite taxa in the Keila and Pirgu stages, i.e. before the $\delta^{13}C$ positive shifts. When considering all groups, different corals in particular, we note that the disappearance rate was high

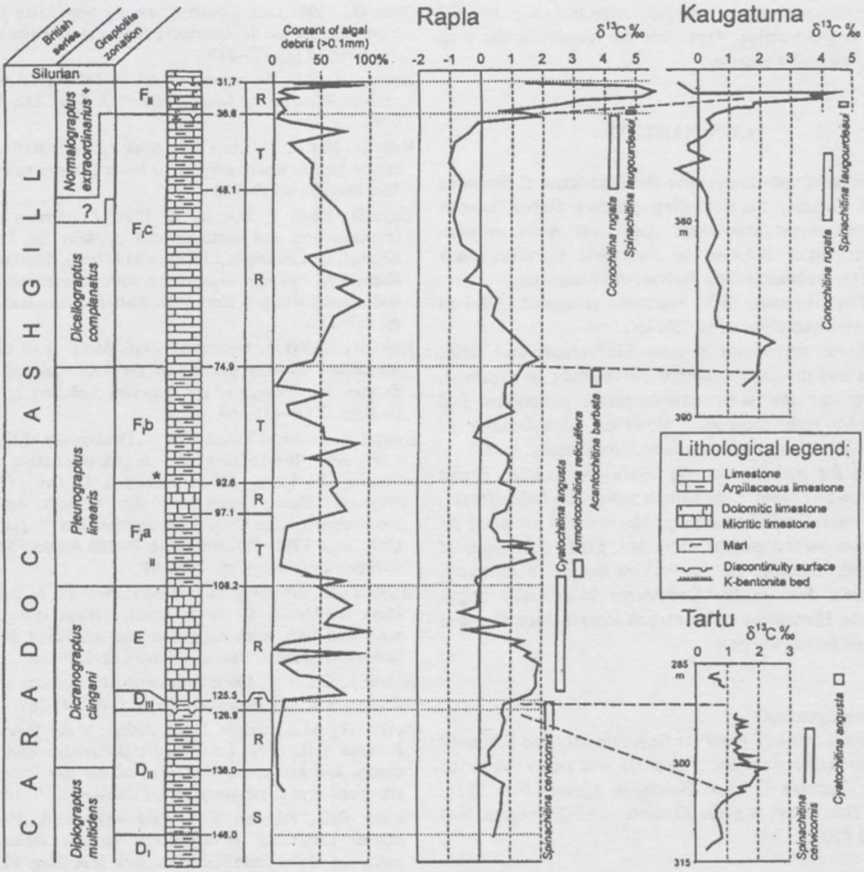

Fig. 2. Correlation of the carbon isotope curves and stratigraphy of the Rapla core. stratigraphical indexes of the Baltic stages: D$_I$ – Haljala, D$_{II}$ – Keila, D$_{III}$ – Oandu, E – Rakvere, F$_I$a – Nabala, F$_I$b – Vormsi, F$_I$c – Pirgu, F$_{II}$ – Porkuni. Letters R, S, T, intervals with prevailing tendency in environmental evolution: R, conventionally regression; T, conventionally transgression; S, without clear general tendency. Content of algal debris in the whole number of sceletal particles according to Põlma (1972). White bars mark occurrences of chitinozoan taxa. * – FAD of *Amorphognathus ordovicicus*.

during nearly the whole study interval, but decreased toward the end of the Caradoc and increased from the beginning of the Ashgill until mass extinction in Pirgu and Porkuni times.

Calcareous algae become an important component of the skeletal material of limestones beginning with the late Caradoc. In the Rapla core there are three intervals where the share of algal particles is above 50% (Fig.2). These are the Rakvere, upper Nabala – lower Vormsi and lower Pirgu stages. All these intervals but lower Vormsi are conventionally regressive cycles (R in Fig.2) represented by pure carbonates with less argillaceous material. According to Jeppsson`s (1990) terminology these intervals belong to the warm and dry secundo episodes. Intervening more clayey rocks (T in Fig.2) should be included into cooler humid primo episodes. The algae curve is much more complicated than the above simplified R–T cyclicity indicating the role of different agents in the basin evolution. The correlation of the algal content and carbon isotope curve shows some coincidences, e.g. three of five δ^{13}C positive shifts (in lower Rakvere, lower Nabala and Porkuni stages) are more or less clearly coeval with lowerings of the algal curve. It means that δ^{13}C positive excursions occur mostly during the cooler primo episodes. Unfortunately, the correlation is not unambiguous, as usual.

In the Central Baltoscandian Confacies Belt (Fig.1), at two levels black shales and dark marls containing abundantly light organic carbon are widely distributed. The lower level, the Mossen Formation, has been correlated with the Estonian sequence from the uppermost Keila to Rakvere stages (Nõlvak 1997). The correlation here is not sufficiently reliable and therefore, any conclusions would be premature.

The correlation of the Fjäcka black shale Formation with the Vormsi Stage seems to be well established (Nõlvak 1997). In such way the Fjäcka shales acting as a sink of organic carbon coincide with the drop in $\delta^{13}C$ values in the Vormsi Stage. But the reason for the drop needs to be studied further.

CONCLUSIONS

1. In the late Ordovician above the Kinnekulle K-bentonite Bed in Estonia, the following positive carbon isotope excursions were established: uppermost Keila or mid-Caradoc, lower Rakvere or early late Caradoc, early Ashgill (*complanatus*) and Porkuni or Hirnantian.

2. The Hirnantian $\delta^{13}C$ excursion reached 5...6‰, all others remained close to the 2‰ level.

3. Some correlation between biodiversity and facies changes and the carbon isotope curve could be observed, showing that the same environmental parameters had influenced both processes. However, identification of certain agents in question needs additional study.

4. In the mid-Caradoc the biotic and isotopic events seemed to be coeval. Late Ashgill bio-events (microfossils, certain macrofossils in the Pirgu Stage) partly preceded the Hirnantian carbon isotope excursion, partly both kinds of the events coincided (in the Porkuni Stage). In summary, the drastic drop in the Ordovician biodiversity began before the Hirnantian major carbon isotope excursion and continued during the peak.

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Paper XII

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New isotopic data solving an old biostratigraphic problem: the age of the upper Ordovician brachiopod *Holorhynchus giganteus*.
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New isotopic data solving an old biostratigraphic problem: the age of the upper Ordovician brachiopod *Holorhynchus giganteus*

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Abstract: Recent work has shown that there is a pronounced positive late Ordovician excursion in both $\delta^{13}C$ and $\delta^{18}O$ which appears to be global and identifies a clear chemostratigraphic interval that is confined to part of the Hirnantian Stage. Pre-Hirnantian (Rawtheyan) brachiopod carbonate values of $\delta^{13}C$ are typically in the range -1 to $+1$‰ PDB and lower Hirnantian values are typically in the range of $+4$ to $+7$‰ PDB. This contrast in isotopic values has been used to assess the age of the distinctive late Ordovician brachiopod *Holorhynchus giganteus*. The age of the *Holorhynchus* association is important in terms of its relationship to the major late Ordovician extinction that was initiated at the start of the Hirnantian. The species, which traditionally was regarded as uppermost Ordovician (Hirnantian) in age has been shown on stratigraphic evidence to be pre-Hirnantian at many locations. If this is so it predates the first phase of extinction and is not associated with the *Hirnantia* fauna, which characterizes the interval between the two main episodes of late Ordovician extinction. However, at its type locality in the Asker district of Norway the stratigraphic evidence has pointed to it being Hirnantian in age. To resolve this ambiguity, analyses have been made on *Holorhynchus* at three locations, (1) in the east Baltic region where *Holorhynchus* occurs below demonstrable Hirnantian rocks, (2) in the Boda Limestone of central Sweden where *Holorhynchus* occurs high on a carbonate mud mound where its age is unclear and (3) in the type area of the species in Norway. At all three locations the $\delta^{13}C$ values are low, indicating a pre-Hirnantian, pre-extinction, age for *Holorhynchus*. The re-assessment of the age of the Asker sequence in Norway implies the presence of a cryptic unconformity there and suggests uplift rather than subsidence during the latest Ordovician. This study emphasizes the value of chemostratigraphy in high-resolution stratigraphy.

Keywords: Ordovician, brachiopods, isotopes, chemostratigraphy.

Changes in the composition of the world oceans can be identified using the stable isotopes of oxygen and carbon in carbonates and it is progressively being recognised that distinctive excursions in an isotopic stratigraphy can identify chemostratigraphic horizons of chronostratigraphic value. Global shifts in $\delta^{18}O$ values commonly reflect world-wide changes in ocean water composition when they result from changes in the volume of polar ice-caps or a world-wide increase or decrease in ocean temperature. However, shifts may be only local when they resulted from regional or local changes in salinity or water temperature. Global shifts in $\delta^{13}C$ occur when there are changes in carbon cycling that affect the world's oceans, but the magnitude and even the direction (more positive or more negative) may differ according to the depth of ocean waters in which the carbonate formed (Berger & Vincent 1986).

Many of the isotopic excursions recorded in the geological record have a short duration and so define horizons with high chronostratigraphic resolution. Most notable of the high resolution records are the succession of $\delta^{18}O$ excursions recognised in deep-sea Quaternary cores reflecting climatic change on a Milankovitch cyclicity of about 20 000, 40 000 and 100 000 years (Hays *et al.* 1976). In pre-Quaternary rocks a Milankovitch cyclicity in the isotopic record can in favourable circumstances be recognised as far back as the Mesozoic (e.g. Ditchfield & Marshall 1989), but has only rarely been used in correlation. However, larger isotopic excursions have been recognised accompanying most of the major extinction events in the Phanerozoic (e.g. the Palaeocene, Kennett & Stott 1995; the K–T event, Perch-Nielsen *et al.* 1982; the Permian, Holser *et al.* 1986; the Devonian, Joachimski & Buggisch 1993; and the late Ordovician, Brenchley *et al.* 1994) and these excursions, which are generally of a duration of only tens to hundreds of thousands of years, form distinctive chemostratigraphic horizons.

Although chemostratigraphy is potentially a powerful tool in chronostratigraphy, there are problems in its application. Firstly there is the problem of identifying a specific isotope excursion in different sections. This problem is less severe in the late Quaternary where continuous, high quality deep sea cores have provided a $\delta^{18}O$ record that can be resolved to give a high-resolution chronostratigraphy spanning the last 300 000 years (Pisias *et al.* 1984; Martinson *et al.* 1987). Amongst older sequences the primary isotopic signal is distorted to some degree by the effects of diagenesis making a high resolution correlation of cyclicity on the Milankovitch timescale more difficult. However, amongst these older sequences some of the isotope excursions are sufficiently large to stand out clearly from background lower amplitude shifts and so may be identified with considerable confidence (Gale *et al.* 1993). A second problem that may arise in the application of chemostratigraphy is establishing that an excursion is a global response to a change in the isotopic composition of the oceans and not a local effect or a variety of local effects at different times and different places. A major first step in establishing global distribution is to demonstrate that an excursion occurred on separate plates or in different oceans at the same time, as near as that can be established. A global distribution is further supported if it can be shown that the excursion is present at different latitudes.

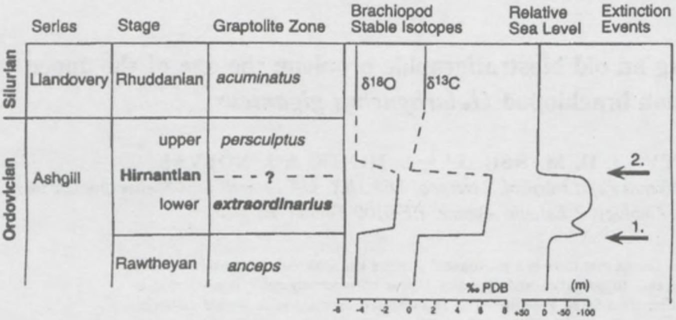

Fig. 1. Stratigraphy across the Ordovician–Silurian boundary, showing the late Ordovician sea-level changes, extinction events (marked 1 and 2) and global stable isotope curves for brachiopod compositions (Marshall *et al.* in press). Note that the events described in this paper are related to the lower Hirnantian.

In most chemostratigraphic studies of pre-Cenozoic rocks, isotopic data have been used to provide a high-resolution correlation of environmental and biotic events during mass extinctions. The isotopes have then been further used to interpret the environmental changes. In this study we concentrate on how chemostratigraphy may contribute to problems in biostratigraphy. Initially we establish the credentials of a late Ordovician isotope excursion as a global chronostratigraphic horizon and then show how it may be used to resolve ambiguities in the age of the large pentamerid brachiopod, *Holorhynchus giganteus*, one of the key elements amongst late Ordovician faunas.

Late Ordovician stratigraphy: the Hirnantian Stage

The isotopic excursion described in this paper is confined to the lower part of the Hirnantian, which is the topmost Ordovician Stage. As originally conceived the Hirnantian consisted of shallow marine deposits with a distinctive brachiopod fauna (Bancroft 1933). The brachiopod association, containing *Hirnantia sagittifera*, subsequently became known as the *Hirnantia* fauna (Temple 1965; Wright 1968). The lower boundary of the stage has never been formally defined, though Ingham & Wright (1970) characterized what they considered to be the limits to the stage. Most workers identify the stage by the presence of the *Hirnantia* fauna and the base of the stage by a change in lithology that generally marks a change to shallower marine facies. The top of the Hirnantian, as originally conceived, is generally marked by an abrupt change to deeper marine facies that in some places contains a *Glyptograptus persculptus* Zone fauna that traditionally was regarded as Silurian in age. When the Ordovician–Silurian boundary was formally defined, it was raised to the base of the *acuminatus* Zone (Cocks 1985) and the Hirnantian Stage by default was extended upwards to incorporate commonly two very different parts, a lower shallow marine part with the *Hirnantia* fauna and an upper deeper marine part with a graptolite fauna. The isotope excursion is confined to the lower part of the Hirnantian which will be referred to informally as the 'lower Hirnantian' throughout this paper (Fig. 1).

The late Ordovician isotope excursion

The Ordovician and Silurian are generally regarded as periods when there were greenhouse climates with high levels of atmospheric CO_2 and a high stand of sea-level (Fischer 1981, 1982; Berner 1990). However, in the late Ordovician, large ice caps developed on Gondwana, there were related glacio-eustatic changes in sea level and there were at least two substantial episodes of extinction (Brenchley 1989). These events were mainly confined to a relatively short period of time in the lower Hirnantian (Fig. 1). The start of a major positive excursion in both $\delta^{18}O$ and $\delta^{13}C$ coincided with the onset of growth of major ice-caps, sea-level fall and the first phase of extinction. The end of the positive excursion coincided with recession of the ice caps, rise in sea level and the second phase of extinction (Fig. 1) (Brenchley *et al.* 1994, 1995).

The isotope data that identifies this major excursion is based mainly on analyses of brachiopod shell material, screened whenever possible for the effects of diagenesis (Marshall 1992) using SEM, cathodoluminescence and trace element analysis on both shells and cements (Middleton *et al.* 1991). $\delta^{18}O$ determinations have come from sequences with low thermal maturity in Sweden, the Baltic Republics, Anticosti, Canada and mid-Continent, USA. In addition to these areas $\delta^{13}C$ determinations have been obtained from some more thermally mature sequences. The late Ordovician isotope excursion is clearly identifiable because values of both $\delta^{18}O$ and $\delta^{13}C$ are substantially more positive than the values before or after the lower Hirnantian. Values of $\delta^{18}O$ measured through Caradoc sequences in Sweden (Marshall & Middleton 1990) and mid-Continent USA (Brenchley *et al.* 1994) fall mainly within a band between -6 and -4‰. and values of $\delta^{13}C$ lie between -1 and $+1$‰. Evidence from the Baltic region shows that $\delta^{18}O$ values rise by 2–3‰ from the base of the Hirnantian to the peak of the excursion and decline to pre-Hirnantian values at the end of the lower Hirnantian. $\delta^{13}C$ values similarly rise by about 6‰ from pre-Hirnantian values to the peak of the excursion and then fall again to pre-Hirnantian values at the end of the excursion (Fig. 2). There are no excursions of nearly comparable magnitude throughout the entire Ashgill or lower Llandovery which could be confused with the Hirnantian excursion (Brenchley *et al.* 1994). The isotopic data from the Baltic region suggests the limits to the excursion coincide with the pronounced facies changes that mark the base and top of the lower Hirnantian. Data from the Oslo region of Norway (this paper) and from China (Wang *et al.* 1993) suggest the same situation. Elsewhere the data has so far been insufficient to establish a detailed isotope stratigraphy. However, the evidence at this stage in our investigations suggests that high late Ordovician isotope values are confined to the lower part of the Hirnantian Stage.

Evidence that the excursion is global comes from the wide geographic spread of the localities at which it has been

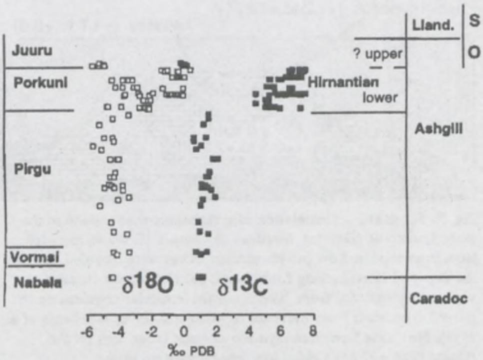

Fig. 2. Stable isotope stratigraphy through the upper Ordovician of the Baltic region. There might be a disconformity at the local stage boundary between the Porkuni and Juuru, in which case the lowermost Juuru could be lowermost Silurian, not upper Hirnantian. Note the positive $\delta^{18}O$ and $\delta^{13}C$ excursions in the Porkuni Stage (regressive interval = lower Hirnantian) with a sharp change in values at both the lower and upper boundaries to the stage (composite stratigraphic plot from Brenchley *et al.* 1994).

recorded and from the number of different plates on which it occurs. From Baltica it is recorded from Estonia, Lithuania, Sweden and Norway, from Laurentia it is known from Anticosti, from Gondwana it is recorded in Argentina and from China it is recorded from the Yangtze Platform (Wang *et al.* 1993). Laurentia and Baltica were located in the tropics in the late Ordovician but China was at mid latitudes and the data from Argentina is from a high latitude sequence containing tillites (Marshall *et al.*. in press). The cumulative evidence suggests that the isotopic excursion marks a chemostratigraphic interval that is globally synchronous and therefore has chronostratigraphic value.

The biostratigraphy and isotopic composition of the upper Ordovician brachiopod, *Holorhynchus giganteus*

Holorhynchus giganteus, an exceptionally large pentamerid brachiopod, was first described from upper Ordovician rocks of Stage 5b in the Oslo region by Kiaer (1902). The species came to be regarded as one of the distinctive elements of uppermost Ordovician faunas and was recognized at several other localities around the world. The species is important because it forms part of a distinctive ecological association, the *Holorhynchus* community, which was thought to be coeval with the *Hirnantia* fauna (Brenchley & Cocks 1982) and therefore post-dated the first phase of the late Ordovician extinction. However, in a review of latest Ordovician faunas, Rong & Harper (1988) noted that faunas containing *Holorhynchus giganteus* commonly lay below the *Hirnantia* fauna and were therefore probably pre-Hirnantian (Rawtheyan) in age. Nevertheless, the stratigraphic range of *Holorhynchus giganteus* remained uncertain because in the Oslo region rocks of Stage 5b with a *Hirnantia* fauna in the east of the region can apparently be traced laterally westwards into Stage 5b rocks with a *Holorhynchus* fauna (Kiaer 1902; Brenchley & Newall 1975). In an attempt to resolve whether *Holorhychus giganteus* is wholly of pre-Hirnantian age or whether it ranges up into the Hirnantian, stable isotope analyses have been made to ascertain whether the species consistently has pre-Hirnantian values. The investigation has used only measurements of $\delta^{13}C$ because previous work (Robertson 1992) had shown that the rocks in the Oslo region have relatively high thermal maturity that has affected $\delta^{18}O$ values (see the very negative $\delta^{18}O$ values in Table 3).

As a preliminary investigation, analyses were made of *Holorhynchus* from the Taučionys Formation in two cores from Lithuania and one from the Pskov region of Russia (Fig. 3), where it had been shown that *Holorhynchus* occurred in rocks of the Pirgu Stage regarded as being pre-Hirnantian in age (Nõlvak *et al.* 1989; Hints 1993). The $\delta^{13}C$ values obtained ranged between 0 and 2‰ (Fig. 4, Table 1) typical of $\delta^{13}C$ values in the pre-Hirnantian (see Fig. 2).

A second investigation was made of *Holorhynchus* from the upper flanks of a carbonate mud mound in Osmundsberget quarry in the Boda Limestone of the Siljan district of central Sweden (Fig. 5). Previous analyses of the brachiopods from the core and flank facies of the mound showed pre-Hirnantian values, except for shells of *Hindella* in coquinas at the crest of the mound which yielded lower Hirnantian values for both $\delta^{18}O$ and $\delta^{13}C$ (Table 2; Marshall & Middleton 1990; Middleton *et al.* 1991). Jaanusson (1979, 1982) regarded *Holorhynchus* as being probably Hirnantian in age by comparison with Norway. However, $\delta^{13}C$ analyses of *Holorhynchus* have yielded values ranging between 1.3 and 1.6‰ suggesting a pre-Hirnantian age. This suggests that Hirnantian sediments are confined to the coquinas at the crest of the mound.

The third and crucial investigation was of the Oslo region where the evidence for *Holorhynchus giganteus* ranging into the Hirnantian was more compelling. There the uppermost Ordovician stage, 5b, is a regressive shale and sandstone sequence with conglomerate and sandstones filling channels incised to

Table 1. *Brachiopod stable isotopic data for the Baltic Republics and Russia*

Location	Formation	Depth (m)	Sample	$\delta^{13}C$‰ PDB	$\delta^{18}O$‰ PDB
Iljinskoje	Taučionys	480.6	*Holorhynchus*	0.2	−3.3
	Taučionys	481.0	*Holorhynchus*	1.2	−4.1
	Taučionys	480.2	*Holorhynchus*	1.1	−3.5
	Taučionys	482.3	*Holorhynchus*	1.9	−3.4
	Taučionys	482.3	*Holorhynchus*	1.9	−3.5
Taučionys	Taučionys	384.0	*Holorhynchus*	0.6	−4.9
	Taučionys	385.0	*Holorhynchus*	1.2	−4.5
Ukmerge	Taučionys	504.8	*Holorhynchus*	1.6	−3.0
	Taučionys	504.8	*Holorhynchus*	1.5	−4.4

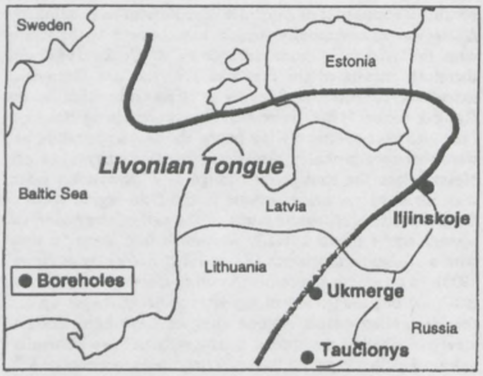

Fig. 3. Map of the Baltic region showing the location of the Iljinskoje, Ukmerge, and Taučionys, cores shown in Fig. 4. The thick black line separates the deeper ramp facies of the Livonian Tongue, from the shallower up-ramp facies.

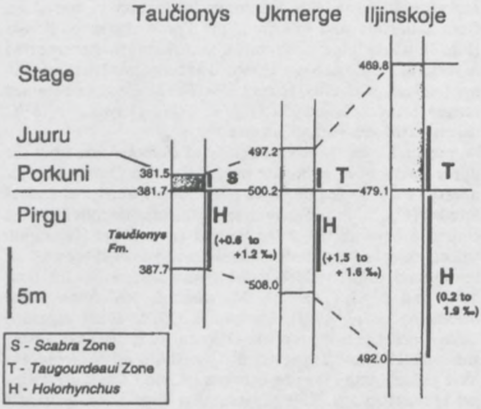

Fig. 4. Stratigraphy of the Taučionys, Iljinskoje and Ukmerge cores. The chitinozoan *taugourdeaui* Zone is lowermost Porkuni in age but might straddle the boundary, into the uppermost Pirgu (Nõlvak & Grahn 1993). The chitinozoan *scabra* Zone is Hirnantian in age. The recorded range of *Holorhynchus* in the Taučionys Formation (Nõlvak *et al.* 1989) is shown together with the low $\delta^{13}C$ values.

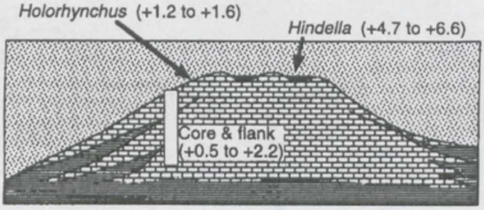

Fig. 5. Schematic representation of a carbonate mud mound in the Boda Limestone (Dalarna, Sweden), showing $\delta^{13}C$ values recorded from brachiopods. Low pre-Hirnantian values were recorded through the core and interfingering flank facies, but high, lower Hirnantian values were recorded from *Hindella* in the lenticular coquinas on the mound crest (data from Marshall & Middleton 1990; Middleton *et al.* 1991). New data from *Holorhynchus* in flank facies high on the mound (Table 1) has yielded low, pre-Hirnantian values.

depths of more than 14 m into the upper part of the sequence. The stage is bounded at its base, near Oslo itself, by a bioturbated sandstone with a distinctive cystoid fauna. Overlying the stage are dark grey shales (Stage 6) with sparse graptolites suggesting a lowermost Silurian age. In their investigation into the upper Ordovician stratigraphy of the Oslo region Brenchley & Newall (1975) showed that it was possible to trace the upper and lower boundaries of stage 5b southwestwards for about 16 km from near Oslo itself into the Asker district (Fig. 6) even though there were progressive changes in facies and the cystoid bed becomes more calcareous and less easily recognized (Fig. 7). A few kilometres further west and northwest in Asker, near Sandvika (Locality 29, Fig. 6) the sequence regarded as Stage 5b comprises mainly mudstones with variable proportions of nodular limestones and is cut by deep channels like those near Oslo. Here, however, the lower boundary of the stage is no longer identifiable, so that, although the overlying Stage 6 shales are present, the lower boundary of Stage 5b and even the presence of the stage itself, cannot be demonstrated with certainty. Nevertheless, Brenchley & Newall (1975) estimated where they thought the lower boundary of the stage might be (Fig. 7). *Holorhynchus* is found in the upper part of the sequence near Sandvika and ranges to within a few metres of the shales of the overlying Stage 6.

In order to establish whether $\delta^{13}C$ values are preserved in the thermally mature rocks of the Oslo region we analysed brachiopod shell material from the Oslo and eastern Asker districts where the Hirnantian Stage could be identified with confidence. The determinations showed high $\delta^{13}C$ values

Table 2. *Brachiopod stable isotopic data for Sweden*

Location	Stratigraphic unit	Position	Sample	$\delta^{13}C$‰ PDB	$\delta^{18}O$‰ PDB
Osmundsberget Quarry, Dalarna	Boda Limestone	Top of mound (see Fig. 5)	*Holorhynchus*	1.6	−4.9
Osmundsberget Quarry, Dalarna	Boda Limestone	Top of mound (see Fig. 5)	*Holorhynchus*	1.6	−4.7
Osmundsberget Quarry, Dalarna	Boda Limestone	Top of mound (see Fig. 5)	*Holorhynchus*	1.2	−4.7
Osmundsberget Quarry, Dalarna	Boda Limestone	Top of mound (see Fig. 5)	*Holorhynchus*	1.5	−4.2

Table 3. *New brachiopod stable isotopic data for the Oslo–Asker districts, Norway*

Section*	Stage†	Height‡ (m)	Sample	δ¹³C‰ PDB	δ¹⁸O‰ PDB
(Oslo/E. Asker)					
Konglungen E. (16)	6	++1.0	*Triplesia*	−0.2	−11.2
Hovedøya (10, *c.* 76)	5b	+40	*Thebesia*	4.43	−12.2
Hovedøya (10, *c.* 76)	5b	+40	*Thebesia*	4.64	−12.7
Hovedøya (10, *c.* 76)	5b	+40	*Thebesia*	4.4	−13.7
Hovedøya (10, *c.* 76)	5b	+40	*Thebesia*	3.1	−14.2
Hovedøya (10, *c.* 76)	5b	+40	*Thebesia*	2.8	−13.6
Hovedøya (10, *c.* 76)	5b	+40	*Thebesia*	2.9	−11.1
Hovedøya (10, *c.* 76)	5b	+40	*Thebesia*	3.2	−14.0
Hovedøya (10, *c.* 76)	5b	+40	*Thebesia*	1.6	−9.0
Hovedøya (10, *c.* 76)	5b	+40	*Thebesia*	1.5	−12.8
Hovedøya (10, *c.* 76)	5b	+40	*Thebesia*	3.8	−11.1
Hovedøya (10, *c.* 76)	5b	+40	*Thebesia*	4.7	−14.0
Hovedøya (10, *c.* 76)	5b	+40	*Thebesia*	3.9	−14.6
Hovedøya (10, *c.* 76)	5b	+40	*Thebesia*	2.5	−12.3
Hovedøya (10, *c.* 76)	5b	+40	*Thebesia*	4.1	−13.5
Hovedøya (10, *c.* 76)	5b	+40	*Thebesia*	4.3	−13.2
Hovedøya (10, *c.* 76)	5b	+40	*Thebesia*	4.9	−9.1
Hovedøya (10, *c.* 76)	5b	+40	*Thebesia*	3.5	−14.3
Hovedøya (10, *c.* 76)	5b	+40	*Thebesia*	2.2	−10.5
Hovedøya (10, *c.* 76)	5b	+40	*Thebesia*	3.0	−15.0
Høyerholmen (14 *c.* 53)	5b	+7.5	*Dalmanella*	6.0	−12.7
Langåra W. (17 *c.* 9)	5b	+3.0	*Cliftonia*	4.9	−18.5
Konglungen (15 *c.* 30)	5b	+2 to 3	*Cliftonia*	4.8	−5.8
Konglungen (15 *c.* 30)	5b	+2 to 3	*Cliftonia*	5.0	−8.5
Konglungen (15 *c.* 30)	5b	+2 to 3	*Hindella*	6.6	−5.7
Gressholmen (8 *c.* 1)	5b	+1 to 5	*Hirnantia*	6.2	−11.5
Langøyene SE (5 *c.* 18)	5b	+1.0	*Hirnantia*	6.4	−16.5
Rambergøya, W. (*c.* 107)	5b	−0.15	*Cliftonia*	5.8	−9.9
Konlungen E. (16 *c.* 30)	5b	−0.10	*Cliftonia*	5.3	−8.7
Langåra W (17 *c.* 33)	5b	−0.5	*Hindella*	4.2	−12.4
Langåra W (17 *c.* 33)	5b	−1.0	*Eospirigerina*	2.6	−14.2
Langåra W (17 *c.* 33)	5b	−1.0	porambonitacean	3.1	−15.3
Langåra W (17 *c.* 32)	5b	−6 to −11	*Eospirigerina*	0.4	−18.2
Kalvøya (28 *c.* 25)	5b	−12	*Dolerorthis*	1.2	−12.2
Bronnøya (*c.* 12)	5b	−15	*Triplesia*	0.3	−13.2
Langåra W. (17 *c.* 31)	5b	−21	*Triplesia*	0.1	−5.5
(W. Asker)					
Sandvika (29 *c.* 5)	?	9–11	*Holorhynchus*	1.4	−11.0
Sandvika (29 *c.* 5)	?	9–11	*Holorhynchus*	0.8	−6.4
Sandvika (29 *c.* 5)	?	9–11	*Holorhynchus*	1.2	−7.6
Sandvika (29 *c.* 5)	?	9–11	*Holorhynchus*	1.4	−13.0
Holmen (27)	?	10	*Holorhynchus*	0.9	−8.9
Holmen (27)	?	10	*Holorhynchus*	0.9	−8.7
Holmen (27)	?	10	*Holorhynchus*	0.2	−16.1
Holmen (27)	?	10	*Holorhynchus*	0.2	−15.1
Holmen (27)	?	10	*Holorhynchus*	1.2	−8.2
Holmen (27)	?	10	*Holorhynchus*	1.1	−8.9
Holmen (27)	?	10	*Holorhynchus*	0.5	−15.0
Holmen (27)	?	10	*Holorhynchus*	0.4	−15.0
Sandvika (29 *c.* 5)	?	10	*Holorhynchus*	5.7	−14.8

*Sample points from Brenchley & Cocks (1982). Section numbers relate to the localities shown on Fig. 1. Collection numbers are prefixed by 'c'. The brachiopod collections are in the British Museum of Natural History, London.

†Samples from Sandvika and Holmen all from Stage 5; previously assigned to stage 5b but now assigned to stage 5a on isotopic evidence.

‡Oslo/E. Asker: height above the base of stage 6 (++); above base stage 5b (+); below base stage 5b (−); W. Asker: height below the base of stage 6.

generally between 3 and 6‰ in the Hirnantian (Fig. 7, Table 3) consistent with Hirnantian values recorded in the Baltic region (Fig. 2) and elsewhere (Brenchley *et al.* 1994). Low $\delta^{13}C$ values were found in the underlying rocks (Fig. 7, Table 3), consistent with pre-Hirnantian values. However, the shift from low to high values does not occur precisely at the boundary between Stages 5a and 5b, as defined by Brenchley & Newall (1975), but within a metre interval below. This suggests that either the

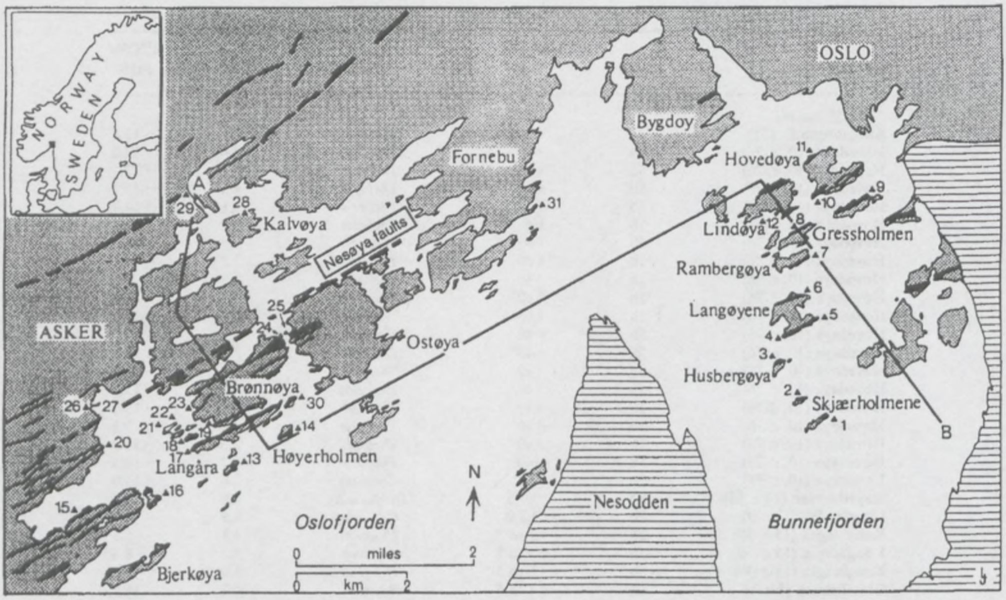

Fig. 6. Locality map of the Oslo-Asker Districts (after Brenchley & Cocks 1982). The upper Ordovician outcrop is shown with black shading. The section line links the localities shown in Fig. 7 and the thick dashed line marks the Nesøya fault zone.

base of Stage 5b does not coincide exactly with the base of the Hirnantian, but is slightly above it, or that the start of the positive carbon excursion slightly preceded the start of the Hirnantian. In either case the disparity in time between the two levels appears to be small. A single determination from Stage 6 showed a low $\delta^{13}C$ value (− 0.2‰) consistent with post-lower Hirnantian levels.

Having established that the positive carbon excursion had been preserved in this region in spite of the high thermal maturity, $\delta^{13}C$ analyses were made on shells of *Holorhynchus giganteus* from high in the supposed 5b sequence in the Sandvika area of Asker where the presence of Stage 5b and therefore the Hirnantian had not been established with certainty. The shells yielded low $\delta^{13}C$ values between 0.5 and 1.5‰ (Fig. 7, Table 3), typical of pre-Hirnantian values and therefore supporting the contention of Rong & Harper (1988) that *Holorhynchus* is of pre-Hirnantian, probably Rawtheyan age. There is, though, one caveat. Analyses were made of a single sample of shells that came from an horizon about 45 m below the *Holorhynchus* beds and these gave values between 0.8 and 1.6‰, consistent with the entire sequence below Stage 6 being pre-Hirnantian. However, there is a locality about 30 m higher with shells of *Onniella kalvoyi* yielding high values between 5.5 and 5.7‰ (Fig. 7). The anomalous high values apparently sandwiched between low values are a source of some concern because they occur in sequences that appear to be complete. However, it is possible that the high values are the result of diagenesis because the *Onniella* shells are poorly preserved, associated with dolomite and stained and impregnated with limonite. Material more suitable for analysis has unfortunately not been recovered from the same part of the section.

Data from chitinozoa

A sample collected from the upper part of the sequence in Asker has yielded a chitinozoan assemblage of stratigraphic value. The sample came from the section that was exposed in Sandviksveien, at Sandvika, Asker (Loc.29, Fig. 6), but is now largely obscured by developments of the roadway. At this locality the uppermost Ordovician, which is in faulted contact with Stage 6 shales, consists of a breccia, less than 1 m thick, with blocks of oolite, overlying bioturbated mudstones that contain a *Holorhynchus* fauna through an interval 4–12 m below Stage 6 (Brenchley & Newall 1975; fig. 9). The sample collected for chitinozoa came from the siltstones in the metre interval below the breccia. The sample yielded a chitinozoan assemblage including *Belonechitina gamachiana*, *B. micracantha* (=*Conochitina*, in Nõlvak *et al.* 1989), *Rhabdochitina gracilis* and *Conochitina minnesotensis*. *Belonechitina gamachiana* occurs in upper Pirgu rocks in the Iljinskoje core (Nõlvak *et al.*. 1989) and appears to be pre-Hirnantian in its range. Its presence in Asker above rocks with *Holorhynchus giganteus* supports the pre-Hirnantian age of the brachiopod species.

Implications

The isotopic data described above (with the exception of the one sample) supports the conclusion of Rong & Harper (1988) that *Holorhynchus giganteus* is probably everywhere of pre-Hirnantian age. Communities with athyrides, atrypides and pentamerides were just beginning to establish themselves in the late Ordovician when they appear to have been mainly eradicated by the extinction at the start of the Hirnantian. The new evidence supporting the disappearance of *Holorhynchus* near the top of the Rawtheyan Stage strengthens the case for

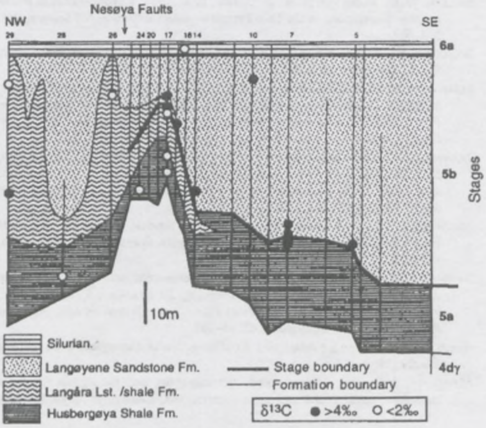

Fig. 7. Stratigraphic profile of upper Ordovician rocks in the Oslo–Asker Districts (see Fig. 6 for line of profile). The stages shown are the long-established Norwegian 'Stages'. Stage 5b is nearly or wholly equivalent to the lower Hirnantian and contains a *Hirnantia* fauna. The Stage 5a–5b boundary can be traced from locality 3 to locality locality 24, but cannot be identified at localities 26, 28, 29 (west Asker) Brenchley & Newall (1975). The stratigraphy in the West Asker district is therefore uncertain. $\delta^{13}C$ values show low pre-Hirnantian values in Stage 5a and a low post-lower Hirnantian value in Stage 6. Note *Holorhynchus* from the upper part of the sequence in the northwest of the area, previously thought to be stage 5b has also yielded low values, suggesting the horizon is in Stage 5a. High $\delta^{13}C$ values, typical of the Hirnantian are confined to Stage 5b or within a metre of the 5a–5b boundary. There is, however, one anomalous point in the section at locality 29.

pentameride extinctions at this time. It was not until the early Silurian that the pentamerides re-appeared in substantial numbers (Harper & Rong 1995).

The revision of the age of *Holorhynchus* has implications for the geology of the Oslo–Asker districts. The revision demonstrates that the sequence in west Asker, near Sandvika, belongs to Stage 5a, not Stage 5b, as was previously thought (Kiaer 1902; Brenchley & Newall 1975; Brenchley & Cocks 1982). The isotopic data described above (with the exception of the one sample) implies that in the Sandvika area, Stage 5b (the Hirnantian) is largely missing beneath the dark grey shales of Stage 6. The deeply incised channels with their conglomerate fill are probably an exception to this; channels are a characteristic element of the Hirnantian glacio-eustatic regression; and are characteristic of the Hirnantian Stage in the Oslo Region and elsewhere (Brenchley & Newall 1980) and are likely to have been preserved where they had cut down deeply into older rocks. The implication that derives from the re-assessment of the age of the *Holorhynchus* bearing beds is that, although the Hirnantian is continuously present from Oslo to Asker, there is a cryptic unconformity below Stage 6 to the northwest at which the Hirnantian is absent except where it is preserved in channels. Consequently the stratigraphy of this area as described by Brenchley & Newall (1975) is in need of revision and the palaeogeographic interpretations arising from this stratigraphy (Brenchley & Newall 1980) also need revising.

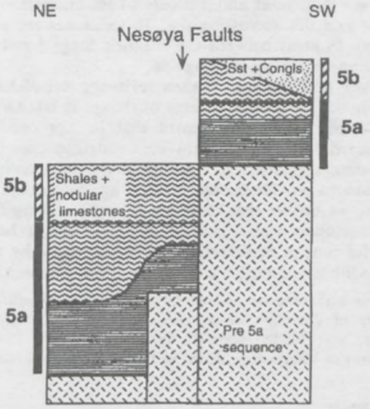

Fig. 8. Schematic reconstruction of the synsedimentary downthrow of the Nesoya Faults in Stages 5a and 5b on the assumption that the beds with *Holorhynchus* belong to Stage 5b (modified from Stanistreet 1983).

The increase in thickness of upper Ordovician strata between the sequences of east Asker and the Sandvika area (between localities 24 and 29 on Fig. 6) have been interpreted to be the result of syn-sedimentary movements on the Nesøya Fault (Figs 6 and 8) which significantly affected the facies and thickness of upper Ordovician sequences (Stanistreet 1983). On the basis that the thick nodular limestone and mudstone sequence with *Holorhynchus* was Stage 5b the Nesøya Fault was interpreted as having a synsedimentary downthrow to the northwest in both Stages 5a and 5b (Fig. 8). The redating of the supposed 5b sequence as 5a suggests that the thick 5a sequence was more probably uplifted during or at the end of

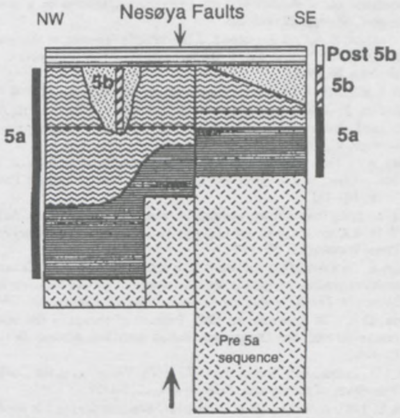

Fig. 9. Schematic reconstruction of the stratigraphy across the Nesoya Faults assuming the beds with *Holorhynchus* belong to Stage 5a. Downthrow to the NW occurred in Stage 5a, but the movement was reversed in 5b. As a consequence there is an unconformity between Stage 6 and Stage 5a, except where Stage 5b is preserved in deeply incised channels (modified from Stanistreet 1983).

the lower Hirnantian and it is only where channels are deeply incised into this succession that 5b sediments are preserved (Fig. 9). In areas between the channels Stage 6 rests with a cryptic unconformity on Stage 5a.

Harper & Rong (1995) when reviewing the *Holorhynchus* fauna in the context of patterns of change in late Ordovician brachiopod faunas, commented that its age can only be determined where independent age constraints exist. The isotopic data supported by the additional information from chitinozoans provide the independent age constraints. In conclusion, we believe that chemostratigraphy, arising from the high resolution stable isotope record described here, is a powerful tool in establishing a chronostratigraphy that will resolve lithostratigraphic and biostratigraphic ambiguities.

F. Steele is thanked for assistance with the stable isotopic analyses. Samples of *Holorhynchus giganteus* from Osmunsberget quarry, Sweden, were kindly provided by A. D. Wright. The isotope laboratory at Liverpool is funded by the University and NERC.

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Carbon isotope stratigraphy in the latest Ordovician of Estonia

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Abstract

On the basis of bulk carbonate carbon isotope data from six drill cores and the type section of the Porkuni Stage of Estonia, $\delta^{13}C$ curves were compiled, which altogether cover nearly the whole succession of the latest Ordovician (Hirnantian) —well known as a time of short-lived glaciation and mass extinction. In pre-Porkuni and Silurian rocks, $\delta^{13}C$ values were low (from −0.2‰ to 1.7‰). For the Porkuni Stage three intervals of the curve were established: (1) lower — a positive shift up to 6.7‰, (2) middle part with variable values from 3.3‰ to 5.4‰, and (3) upper phase with slightly decreased values from 3.2‰ to 3.6‰. The correlation with the Dob's Linn (Scotland) kerogen $\delta^{13}C$ curve shows that the first two intervals coincide with the *Normalograptus extraordinarius* graptolite Biozone, the third one with the *N. persculptus* Biozone. © 2001 Elsevier Science B.V. All rights reserved.

Keywords: Carbon isotopes; Isotope stratigraphy; Hirnantian; Late Ordovician; Baltic area

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1. Introduction

A distinct isotopic "event" associated with a glaciation and mass extinction in the latest Ordovician has been thoroughly investigated (Brenchley et al., 1994; Marshall et al., 1997, etc.) and different explanations of this phenomenon suggested. These were recently summarized and critically evaluated by Marshall and Brenchley (1998, p. 952) and therefore no additional comments are needed here. Their general conclusion that "the driving force(s) for the widespread changes are far from clear" seems a bit pessimistic, however, they stress that exact timing of environmental changes and their relationships with isotopic events are highly instructive for the understanding of the whole situation. This statement has also been our idea when compiling the paper.

The Estonian material (brachiopod shells, kerogen) was studied also by the above authors, but, due to the complicated local geology, only a generalized temporal trend could be established. Different lines of evidence indicate that the real environmental evolution during the Hirnantian has been more complex than seen in generalized curves (isotopes, sea level, biotic events, etc.). In order to identify better the actual succession of events in the Baltic during the latest Ordovician, a detailed, almost continuous composite section was compiled, with sedimentary rocks from boreholes and the stratotype of the Porkuni Stage (Fig. 1). The objective was to establish a carbon isotope curve that would be as continuous as possible for the Hirnantian time interval. This was achieved through analyses of bulk carbonate sam-

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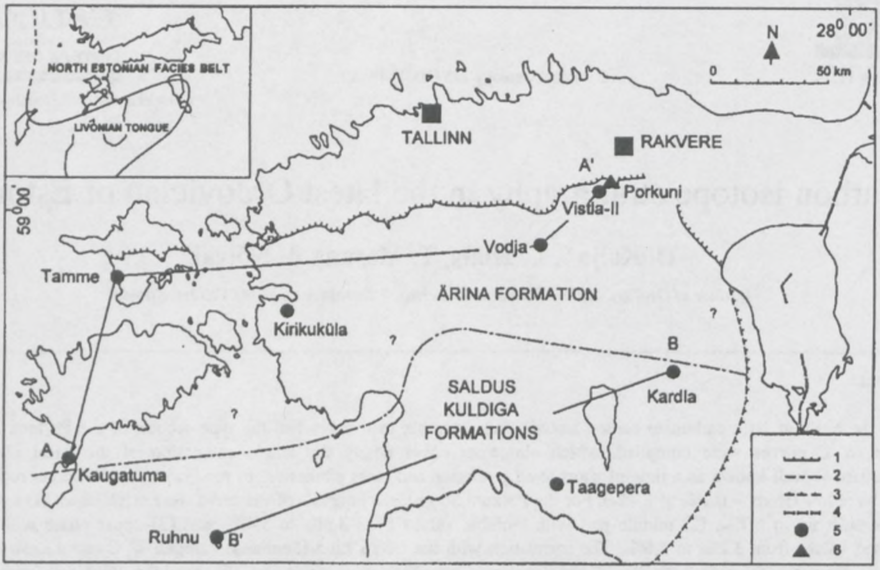

Fig. 1. Location of sampled boreholes and Porkuni stratotype outcrop. The inset shows the distribution of facies belts (Nõlvak, 1997). The Livonian tongue belongs to the Central Baltoscandian facies belt. Key: 1, limit of post-sedimentary erosion; teeth mark the area where rocks were eroded; 2, northern limit of the distribution of the Kuldiga and Saldus formations; 3, same as 1, unproved; 4, state boundary; 5, borehole; 6, outcrop.

ples, as brachiopod shells were not present in every bed. In addition, results from this investigation were correlated with the temporal trend of δ^{13}C (kerogen) at Dob's Linn, Scotland (Underwood et al., 1997), in order to link the Estonian sections to the graptolite biozonation and thus obtain a precise age determination.

2. Geological setting

Ordovician rocks in the East Baltic area are well exposed in outcrops located in northern Estonia and made accessible by many borehole sections in the subsurface of southern Estonia. The Baltic Ordovician basin is usually subdivided (Männil and Meidla, 1994) into two facies belts (Fig. 1): (1) North Estonia, mainly with shallow shelf sedimentary rocks and (2) the Livonian tongue of the Central Baltoscandian belt, occupying the southern part of the study area and represented by deeper shelf deposits, sometimes even by black graptolitic shales. At the end of the Ordovician (Porkuni time), due to general shallowing of the sea, the difference between both belts was still existing, but it was not as substantial as earlier in the Ashgill. In addition, regional tectonic movements caused several gaps in the series, which complicate correlation of rock sequences from these belts. No transitions between the belts have been recognized.

In Estonia, the latest Ordovician (approximately equivalent to the Hirnantian Stage) is represented by the Porkuni Regional Stage, which shows complex facies change both in succession and regionally. Accordingly, the stage consists of three formations and nine members distributed in different parts of Estonia (Fig. 2). This complexity causes some problems with correlation, but these are not the topic of this contribution and we address to Fig. 2, which is compiled using carbon isotope and geological data. The following paragraphs provide a brief description of all lithostratigraphical members with more or less uniform lithologies from bottom upward (thickness according to data from Estonia).

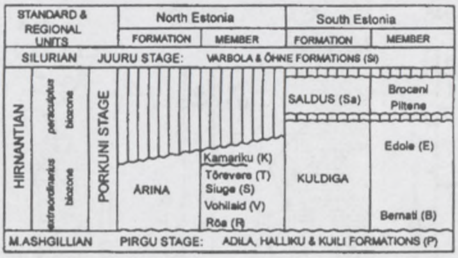

Fig. 2. Latest Ordovician stratigraphy of Estonia. Vertical ruling and wavy lines — gaps in the sections; letters in parentheses — stratigraphical indexes appearing in Figs. 3 and 4 and Table 1.

2.1. The Ärina Formation

The Röa Member — up to 5.5 m of mostly secondary dolomites with sparse organic remains. Most important are crinoidal columnals which form crinoid limestone interbeds at some levels, the *Elsaella* brachiopod Association (Hints, 1997) and *Spinachitina taugourdeaui* (Eisenack) as a micropalaeontological marker (Nõlvak and Grahn, 1993).

The Vohilaid Member — up to 3.7 m of varied skeletal limestones (grainstones in Porkuni quarry; Nestor, 1990) with numerous bryozoans and the *Streptis* brachiopod Association (distributed also in the overlying two members; Hints, 1997). A few interlayers of oolites have been recorded (Hints and Meidla, 1997).

The Siuge Member — up to 2.6 m of bituminous argillaceous limestones (packstones at Porkuni; Nestor, 1990) and marls with brachiopods, ostracodes, etc. Dendroid graptolites occur in several brown shaly interbeds.

The Tõrevere Member — up to 4.1 m of massive biohermal (skeletal wackestones at Porkuni; Nestor, 1990) and layered micritic limestones with abundant stromatoporoids, tabulate and rugose corals, and brachiopods. Locally admixture of quartzose silt and sand occurs (Oraspõld, 1975).

Three last members can be considered together as a "reef complex" because the first small organic buildups appear in the Vohilaid Member, whereas in Porkuni quarry a bioherm has grown through all these members (Nestor, 1990). Bituminous limestones can occur both below and above the reefs, and are interpreted as inter-reef facies (Oraspõld, 1975). Relations of these rock bodies within the complex are not clarified in detail yet, but thickness of the members is locally very variable.

The Kamariku Member (thickness up to 2.1 m), which forms the topmost part of the Ärina Formation, differs from the underlying members in a considerably increased content of sand and silt. Thereby, practically unfossiliferous arenaceous or silty limestones and quartzose sandstones (in the northeastern part) rest on different older rocks beginning with the Vohilaid Member (Männil and Meidla, 1994).

Usually, the top of the Ärina Formation bears clear signs of post-sedimentary erosion (palaeokarst, channelling). However, it is not clear how much of the sequence is missing. In South Estonia the Porkuni Stage appears to be more complete, consisting of the deeper shelf Kuldiga and shallower Saldus formations. Still small gaps are probable below and above the latter (Fig. 2).

2.2. The Kuldiga Formation

The Bernati Member — up to 5 m of argillaceous skeletal limestones (mostly wackestones) corresponding to the *S. taugourdeaui* Biozone contain rare macrofossils, e.g. in the Taagapera core, there occur some cephalopods, crinoid ossicles (often found also in the Röa Member in the north) and poorly preserved brachiopods.

The Edole Member — up to 15 m of interbedded marls and skeletal limestones. Some levels show admixture of sand and silt (Ulst, 1972). In the Ruhnu and Taagepera cores, the upper 3–5 m of this member are represented by argillaceous silty limestone and dolomite with characteristic laminar bedding. A rich shelly fauna occurs in this unit with brachiopods of the *Hirnantia* Association (Hints, 1997), trilobites of the *Dalmanitina* Beds (Männil et al., 1968) and numerous ostracodes (Gailite, 1970).

2.3. The Saldus Formation

The Piltene Member — up to 5 m of skeletal and oolitic limestones with considerable admixture of sandy-silty material, locally sandy limestones, displaying also laminar and cross-bedding. Fossils are very rare.

The Broceni Member — up to 2 m of argillaceous skeletal limestones and marls, sometimes more or less silty or siltstones with mud cracks and ripple marks. Amongst the fossils a few trilobites of the *D. mucronata* group and several ostracodes are most remarkable. The roof of the Saldus Formation, as well as its bottom, bears distinctive signs of post-sedimentary erosion (discontinuity surfaces).

Summarizing the paleontological and lithological characteristics of these sections, we can see that in North Estonia a pelmatozoan association, occurring in the Röa Member (observed also in the Bernati Member) is followed by a bryozoan association in the Vohilaid Member, which is in turn replaced by a stromatoporoid-coral assemblage in the Tõrevere Member. This succession can be interpreted as a regressive cycle. The last conclusion is in harmony with the comment by Brenchley and Harper (1998), that similar community changes are often interpreted as ecological replacements in response to falling sea level. Considering a discontinuity surface at the top of the Tõrevere sequence in a few cores and some difference in the areal distribution of the Kamariku Member compared to the underlying member, a gap between the two uppermost members of the Ärina Formation cannot be excluded. However, this is not sufficiently proved yet.

In South Estonia, two sedimentary cycles corresponding to the Kuldiga and Saldus formations have been established. Both formations were formed in shallow sea conditions and terminated with a clear erosional episode. However, the Kuldiga cycle is more variable, since interbedding of marls and various skeletal limestones including grainstones, suggests frequently changing conditions in the Edole sea. Judging from the distribution of the *Hirnantia* Association in the Edole Member, the last seems to reflect a somewhat deeper environment, most probably open or mid-shelf area. The same conclusion can be inferred from the distribution of a low-diversity assemblage including acid-resistant microfossils, rugose corals, *Rhabdo~~po~~por~~a~~*re algae and some bivalves in the same interval as brachiopods of the *Hirnantia* faunas (*H. saggitifera* (M'Coy), *D. testudinaria* (Dalman), *Eostropheodonta hirnantensis* (M'Coy); Hints, 1997). Scarcity of fossils is particularly true for South Estonian sections, which are located in a more carbonate-rich periphery of the Kuldiga facies area. In the roof of the Kuldiga open shelf rocks a discontinuity surface occurs in several sections. The overlying Saldus Formation is represented by much shallower Piltene oolites and sandy limestones, which transgressively overlap the Edole rocks. This indicates a gap between the formations. On the other hand, some features common to the Kuldiga and Saldus faunas (*D. mucronata* a.o.) suggest that the gap between these formations could be only short. The early Saldus deepening of the basin continued also in Broceni time, but was very soon followed by shallowing and sedimentation in rather changeable near-shore conditions evidenced by mud-cracks, ripple marks, sandy rocks and discontinuity surfaces.

In summary, in South Estonia the Porkuni Stage displays two cycles which are both transgressive in the beginning and later turn into regression and erosion. The Saldus transgression might be coeval with the commencement of the black shale distribution in the latest Ordovician of Britain.

3. Material and methods

Samples were derived from drill cores, and additional material from the stratotype of the Porkuni Stage. The latter was combined with samples from the Vistla-II well about 1 km SW of the old Porkuni quarry, specially drilled in order to complete the stratotype section (both boundaries of the stage were penetrated). Distinct lithological markers allow the firm correlation of both sections. However, only Vistla-II is shown in Fig. 3. The studied wells are relatively shallow — mostly less than 200 m from the surface, but may reach also 620 m (Ruhnu). Sampling intervals were 50–70 cm or less in most North Estonian sections, 25–30 cm in two South Estonian sections. The studied cores are housed in the Geological Survey of Estonia and, together with palaeontological and other material in the Institute of Geology at Tallinn Technical University. Results of 141 carbon and oxygen isotope analyses are given in Table 1, identifications of the most important fossils in Figs. 3 and 4.

The isotope analyses were performed at the Palaeoclimatological Laboratory of the Institute of

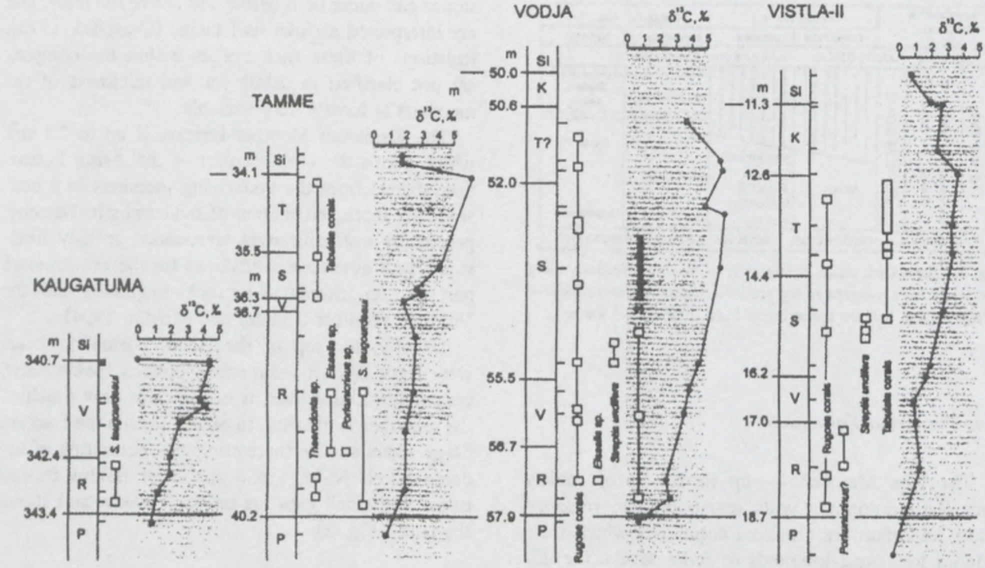

Fig. 3. Carbon isotope curves and occurrence of characteristic fossils in well sections of North Estonia. Black dots — values see Table 1; white squares — occurrences of identified fossils. Letters (for explanation, see Fig. 2) indicate lithostratigraphical units described in the text.

Geology. Whole-rock samples were crushed and material for isotopic analysis was selected avoiding obvious veins or burrows. The material was powdered and treated 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. Reproducibility of replicate analyses was generally better than 0.1‰.

Further aspects of the method used and diagenetic influence are discussed in Kaljo et al. (1997) and summarized in short below. Geochemical studies showed in most of the sections a high Sr/Ca ratio, close to marine calcites, which suggests good preservation of carbonate material. Low Mn content and low conodont colour alteration indices (1–2) evidence that the Baltic Ordovician and Silurian sedimentary rocks and corresponding $\delta^{13}C$ values are not seriously affected by diagenetic processes. The same conclusion was earlier stressed by Brenchley et al. (1994) for Ordovician rocks of Estonia and by Samtleben et al. (1966) and Wenzel and Joachimski (1996) for Silurian rocks of Gotland adjacent to our study area.

The oxygen isotope ratios are more sensitive to diagenetic alteration, as shown by Marshall (1992). In the studied cores the measured $\delta^{18}O$ values in bulk carbonates varied generally between −2.0‰ and −6.8‰ VPDB (Table 1). This is close to the mean oxygen isotopic ratio in the Ordovician carbonates (Qing and Veizer, 1994) and is in accordance with the above opinion about insignificant role of diagenesis in studied sections. However, since calcite and dolomite have clearly different oxygen isotope fractionation factors and the results of oxygen isotope analyses depend on the calcite/dolomite ratio (which is highly variable in our samples), we cannot consider our data on oxygen isotopes to be reliable for environmental reconstructions and we refrain ourselves from discussing them further.

Carbon isotopes for this contribution were measured in whole-rock samples collected from sections

Table 1
Whole-rock isotopic data from the sections investigated

Depth (m)	Stratigraphy	$\delta^{13}C$ (‰)	$\delta^{18}O$ (‰)	Depth (m)	Stratigraphy	$\delta^{13}C$ (‰)	$\delta^{18}O$ (‰)	Depth (m)	Stratigraphy	$\delta^{13}C$ (‰)	$\delta^{18}O$ (‰)
Kaugatuma				*Porkuni*				*Kardla*			
340.65	Si	−0.1	−5.1	0.01	T	4.1	−5.3	160.2	Si	0.7	−4.3
340.70	V	4.4	−3.2	0.9	T	1.3	−6.2	161.0	Si	2.1	−4.1
340.80	V	4.4	−4.2	1.5	T	1.0	−4.8	161.2	Sa	4.6	−4.7
341.35	V	3.7	−4.9	1.7	S	0.3	−5.6	161.5	Sa	4.9	−5.1
341.45	V	3.9	−5.1	2.4	S	2.2	−5.9	162.0	Sa	5.3	−5.3
341.80	V	2.5	−5.0	3.2	V	1.6	−5.8	162.5	Sa	5.4	−5.3
342.30	V	1.8	−3.3	3.7	V	1.2	−5.9	163.0	Sa	5.8	−4.9
342.70	R	1.8	−4.0	4.2	R	1.6	−3.4	163.5	Sa	5.7	−4.9
343.00	R	1.2	−3.9	4.9	R	1.3	−3.3	164.0	Sa	5.6	−5.0
343.55	P	0.8	−3.7	5.5	P	0.1	−5.1	164.5	Sa	5.6	−5.1
								165.0	Sa	5.6	−5.1
Tamme				*Ruhnu*				165.5	Sa	5.4	−4.8
33.7	Si	1.7	−6.5	600.3	Si	1.7	−3.4	166.0	Sa	5.4	−4.8
33.9	Si	1.8	−6.8	600.9	Si	2.1	−2.8	166.5	E	6.0	−4.8
34.2	T	6.0	−3.7	601.3	Sa	3.6	−4.7	167.0	E	6.0	−4.3
35.8	S	4.0	−4.8	601.8	Sa	3.3	−4.8	168.0	E	6.2	−4.0
36.1	S	2.6	−4.6	602.3	Sa	3.6	−5.0	169.0	E	6.4	−3.6
36.2	S	2.8	−5.1	602.8	Sa	3.5	−4.7	169.5	E	6.2	−3.8
36.3	S	1.8	−4.4	603.3	E	3.4	−4.2	170.0	E	6.6	−4.0
36.9	R	2.5	−4.6	603.8	E	3.2	−2.5	170.5	E	6.6	−3.4
38.0	R	2.4	−4.0	604.3	E	3.2	−2.4	171.0	E	6.7	−3.6
38.6	R	2.1	−4.5	604.8	E	4.1	−2.6	172.0	E	6.7	−2.9
39.6	R	1.9	−3.8	605.3	E	4.0	−4.2	173.0	E	6.4	−3.0
40.5	P	0.8	−4.1	605.8	E	3.8	−3.3	173.5	E	6.0	−3.8
				606.3	E	4.1	−3.0	174.5	E	6.3	−3.5

Vodja				606.8	E	4.0	−2.8	175.0	E	6.2	−3.4
50.9	T	3.8	−2.0	607.3	E	3.8	−2.8	175.0	E	6.1	−3.5
51.6	T	5.7	−4.8	607.8	E	5.0	−2.6	175.5	E	5.8	−3.8
51.8	T	5.9	−4.6	608.3	E	3.3	−2.8	175.5	E	5.9	−3.6
52.4	S	4.9	−4.3	608.8	E	4.0	−2.9	176.0	E	3.6	−3.2
52.5	S	5.9	−4.7	609.3	E	4.3	−2.5	176.5	E	3.7	−3.2
53.5	S	5.7	−4.9	609.8	E	5.0	−2.2	177.0	E	3.5	−3.0
55.1	S	4.5	−4.6	610.3	E	4.3	−2.6	177.5	E	5.4	−3.8
55.7	V	3.9	−5.3	610.8	E	4.3	−2.8	178.0	E	4.9	−4.0
56.1	V	3.4	−5.9	611.8	E	3.4	−3.0	179.0	E	5.1	−3.8
57.3	R	2.2	−4.8	612.3	E	5.2	−3.3	179.5	E	5.3	−3.5
57.6	R	2.4	−4.3	612.8	E	5.4	−2.8	180.0	E	4.6	−3.9
58.0	P	0.7	−4.4	613.3	E	3.6	−3.2	181.0	B	2.8	−3.2
				613.8	E	4.4	−3.6	181.5	B	3.1	−3.1
Vistla II				614.3	E	6.0	−3.3	182.0	B	2.3	−2.9
10.8	Si	0.6	−6.4	614.8	E	5.1	−3.2	184.0	B	1.9	−3.2
11.3	Si	1.7	−6.5	615.3	E	5.1	−3.6	185.0	B	1.5	−3.2
11.4	K	2.5	−4.0	615.8	E	4.4	−3.9	186.0	P	0.9	−3.4
12.1	K	2.4	−6.5	616.3	E	3.7	−3.4				
12.6	K	3.6	−5.8	616.8	E	3.1	−3.2				
13.5	T	3.0	−6.0	617.2	B	2.5	−2.4				
14.0	T	3.1	−6.1	617.5	B	2.4	−2.7				
15.1	S	2.7	−4.8	618.0	B	2.3	−3.3				
16.0	S	1.9	−6.3	618.3	B	1.9	−2.3				
16.2	S	1.7	−5.9	618.7	B	1.9	−3.3				
16.6	V	0.8	−6.2	619.0	B	1.4	−3.4				
17.0	R	0.9	−4.4	619.3	P	1.5	−2.7				
17.9	R	1.2	−3.0								
19.4	P	−0.5	−5.5								

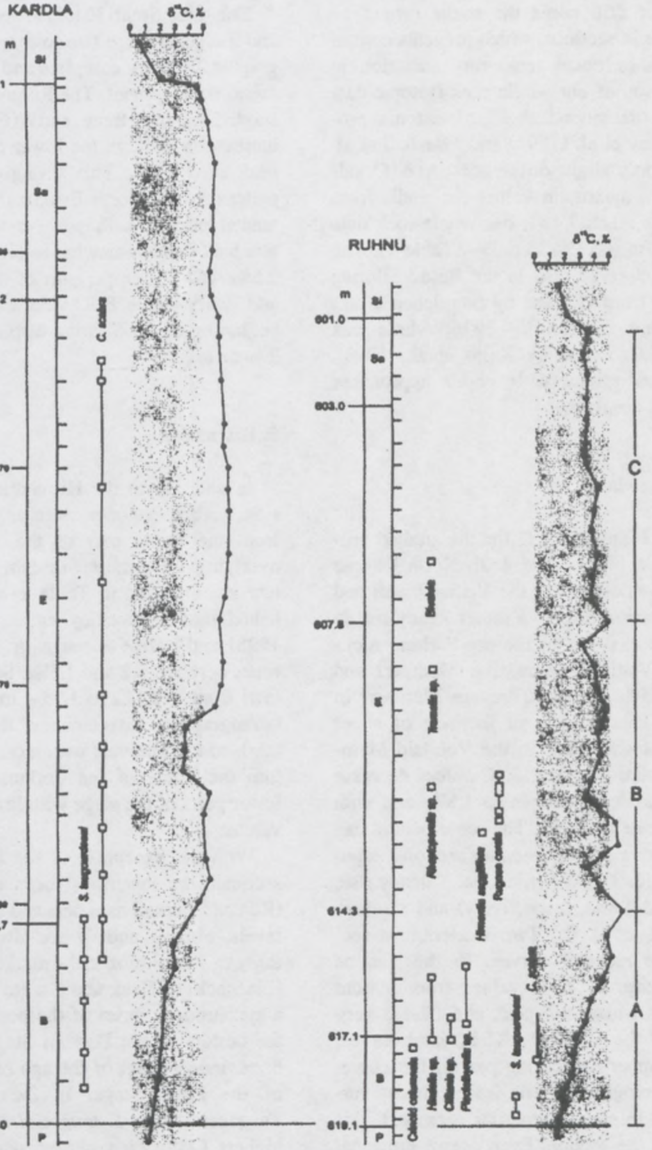

Fig. 4. Carbon isotope curves and occurrences of characteristic fossils in two well sections of South Estonia. Letters A, B, C mark the intervals of the curve discussed in the text. For additional explanations, see Fig. 3.

at more or less regular intervals and with the aim to characterize every significant bed of the succession. We accept that brachiopod shells are the best material for isotopic studies, but due to ecologic control and small size of drill cores the shells cannot be found everywhere in sections, which prevents compilation of an uninterrupted temporary variation in δ^{13}C. A comparison of our whole-rock isotopic data and those from brachiopod shells of Estonia provided by Brenchley et al. (1994) and Marshall et al. (1997) displays only slight differences in δ^{13}C values. For example, maximum values for shells from the Porkuni rocks reach 7.4‰; our whole-rock data show the values from 6.1‰ to 6.7‰ (Table 1). The same pattern is observed also in the Baltic Silurian (brachiopod data from Gotland by Samtleben et al., 1966, Wenzel and Joachimski, 1996; whole-rock ones from the East Baltic by Kaljo et al., 1998), where the curves produced by both approaches demonstrate great similarity.

4. Carbon isotope data

Temporal variations in δ^{13}C for the studied sections are shown in Figs. 3 and 4. In North Estonia two types of curves occur: (a) the Vistla-II well and Porkuni quarry section display a rather steady rise in δ^{13}C values from −0.4‰ in the pre-Porkuni rocks up to 3.6‰ at Vistla-II (Kamariku Member) and from 0.2‰ to 4.1‰ (in the Tõrevere Member) in Porkuni quarry. This reflects an increase of about 4‰, with a minor shift back in the Vohilaid Member. In the Kamariku Member, δ^{13}C values decrease with the last Ordovician sample to 1.8‰ and with the first Silurian one to 0.6‰. The same pattern can be observed in other North Estonian sections, especially at Vodja (Fig. 3) and Kirikuküla: a steady rise, a peak (5.9‰ and 4.8‰, respectively) and a subsequent decrease in δ^{13}C. (b) Two westernmost sections demonstrate reduced curves. In the Tamme section, the increase in δ^{13}C values from bottom upwards (with a set-back and peak at 6.1‰) is very similar to that of the Vistla-II and Porkuni curves, but without the upper decreasing part of the curve. Together with geological criteria (discontinuity surface in the roof, missing Kamariku Member), this indicates a gap in the section. Even deeper erosional cutting is seen in the Kaugatuma section, where a lower isotope maximum of 4.4‰ is reached in the top of the Vohilaid Member, followed by an abrupt negative shift to −0.2‰ in the lowest Silurian.

The three South Estonian sections (Kardla, Ruhnu and Taagepera) are two to three times thicker, stratigraphically more complete and quite variable in their facies development. The Ruhnu δ^{13}C curve might be subdivided into three parts (Fig. 4): (A) A steady increase in δ^{13}C in the lower interval up to the first peak at +6.0‰. This strongly resembles a similar pattern in the North Estonian sections. (B) An extended plateau with positive values between 5.0‰ and 5.4‰ and somewhat lower values from 3.3‰ to 3.5‰. (C) The upper part of the section with lower and less variable δ^{13}C values (just above 3‰). The beginning of the Silurian displays δ^{13}C values of ca 2‰ or less.

5. Discussion

In addition to the Hirnantian part of the succession, carbon isotopes were analysed both for rocks from the upper part of the Pirgu Stage and the overlying lowermost Silurian. Our results (only a few are included in Table 1) agree well with published data (Brenchley et al., 1997; Kaljo et al., 1998) and define a range in δ^{13}C for pre-Porkuni rocks between 0.2 and 1.5‰. Silurian values are also low: from −0.1‰ to 1.7‰, in some cases to 2.1‰ (stratigraphical assignment of these samples is uncertain), so the intervals with increased δ^{13}C values fall into the limits of the Porkuni Stage, as do some lower parts of the stage with low but increasing δ^{13}C values.

With the exception of the Porkuni and Vistla-II sections, the lowermost beds of the Porkuni Stage (Röa and Bernati members and sometimes also higher levels, Figs. 3 and 4) are always well constrained through the biozonal marker *S. taugourdeaui* (Eisenack) (Nõlvak and Grahn, 1993). This indicates a synchronous onset of the positive shift in δ^{13}C at the bottom of the Porkuni Stage in all studied sections, independent of the age of the underlying beds of the Pirgu Stage. In the Kardla, Ruhnu and Taagepera cores, *S. taugourdeaui* is followed by the highest Ordovician chitinozoan biozone of *Cono-*

chitina scabra (Nõlvak and Grahn, 1993). *C. scabra* is not found in North Estonian sections due to unfavourable conditions for the preservation of chitinozoans. However, similarities in the carbon isotope pattern suggest that the $\delta^{13}C$ maximum at the end of the increasing curve falls into the lower part of that biozone (Fig. 4), which corresponds to the uppermost beds of the Ärina Formation in North Estonia.

Another refinement results from the correlation of the corresponding sections in Estonia, Scotland and Anticosti Island, Canada. Underwood et al. (1997) correlate their kerogen $\delta^{13}C$ peak in the top of the *Normalograptus extraordinarius* graptolite Biozone at Dob's Linn, Scotland, with an interval of increased carbonate $\delta^{13}C$ values in the Laframboise Member and the basal Oncolite Platform Beds of Anticosti Island. This interval occurs, as in Estonia (B in Fig. 4) at some distance above the *S. taugourdeaui* Biozone.

The Dob's Linn section is a highly condensed section where the entire Hirnantian is less than 4 m thick. In contrast, sections on Anticosti Island and in South Estonia are several times thicker. As a consequence, there the positive $\delta^{13}C$ excursions are not sharp peaks as at Dob's Linn but broad intervals of positive $\delta^{13}C$ values. These intervals are very similar in both regions. Thus, analogous to Underwood et al. (1997), we correlate the interval B in the Ruhnu core with the peak level at Dob's Linn. On the other hand, our conclusion is partly supported by kerogen $\delta^{13}C$ data from several closely located boreholes on Hiiumaa Island (Marshall et al., 1997), where the corresponding mean values are $-25.2‰$ and $-28.2‰$ in the rocks of the Tõrevere Member and the underlying Siuge Member, respectively. This positive shift in kerogen $\delta^{13}C$ values is coinciding with a carbonate $\delta^{13}C$ excursion according to our data from the Tamme core ($K = -38$ by Marshall et al., 1997, see our Fig. 3, Table 1). The coincidence of positive kerogen and carbonate $\delta^{13}C$ excursions allows us to believe that the correlation of the carbon isotope maximum in the Porkuni Stage with the mid-Hirnantian peak at Dob's Linn is well grounded. This means that the whole Ärina Formation and the main part of the Kuldiga Formation can be correlated with the *N. extraordinarius* Biozone. In South Estonia, below the peak level and higher (Fig. 4), typical representatives of the Hirnantian brachiopod-dominated association occur (Hints, 1997). According to this correlation, the upper Hirnantian part of the carbon isotope curve corresponding mainly to the *N. persculptus* Biozone is represented in the uppermost Kuldiga and Saldus formations, but has no counterpart in North Estonia. Considering also sea level changes, discussed earlier, correlation with the Saldus Formation seems more appropriate. The decline in $\delta^{13}C$ in the *N. persculptus* Biozone is much less pronounced (2–3‰) in the Baltic area than at Dob's Linn (4–5‰), which may largely be due to differences in the sedimentary environments and/or sample material.

Geological data indicate the occurrence of some gaps in the sections (see above). The one in the top of the Ärina Formation is profound and is clearly reflected in the carbon isotope curve. Yet, this gap is not so important for our discussion, since it is filled by the Kuldiga and Saldus formations in South Estonia. The gaps below and above the Saldus Formation, which may change the temporal trend in $\delta^{13}C$, should be treated with more caution. The smooth shape of the $\delta^{13}C$ curve suggests that these hiatuses embrace only relatively short time intervals.

6. Conclusions

The carbon isotope curve measured in whole-rock samples from the Porkuni Stage of Estonia displays a strong positive excursion in the middle of the Hirnantian and a slow decrease in values towards the end of the Ordovician. Overall, similarity exists between the carbon isotope variations for sections from Estonia and at Dob's Linn. However, as a consequence of greater sediment thickness, the temporal trend is much smoother in Estonia. Carbon isotope data provide a means of correlating sections on the Baltic shelf with the sequence at Dob's Linn, Scotland, and thus, the shelly fauna with the deep water graptolite biozonation.

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