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

Dimitri KALJO, Linda HINTS, Olle HINTS, Tõnu MARTMA, Jaak NÕLVAK

Institute of Geology at Tallinn Technical University, Estonia Boulevard 7, 10143 Tallinn, Estonia; e-mail kaljo@gi.ee

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.

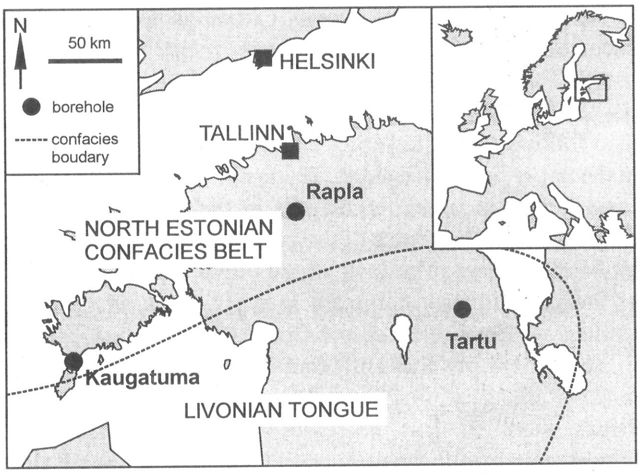

Fig. 1. Location of boreholes and facies belts.

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

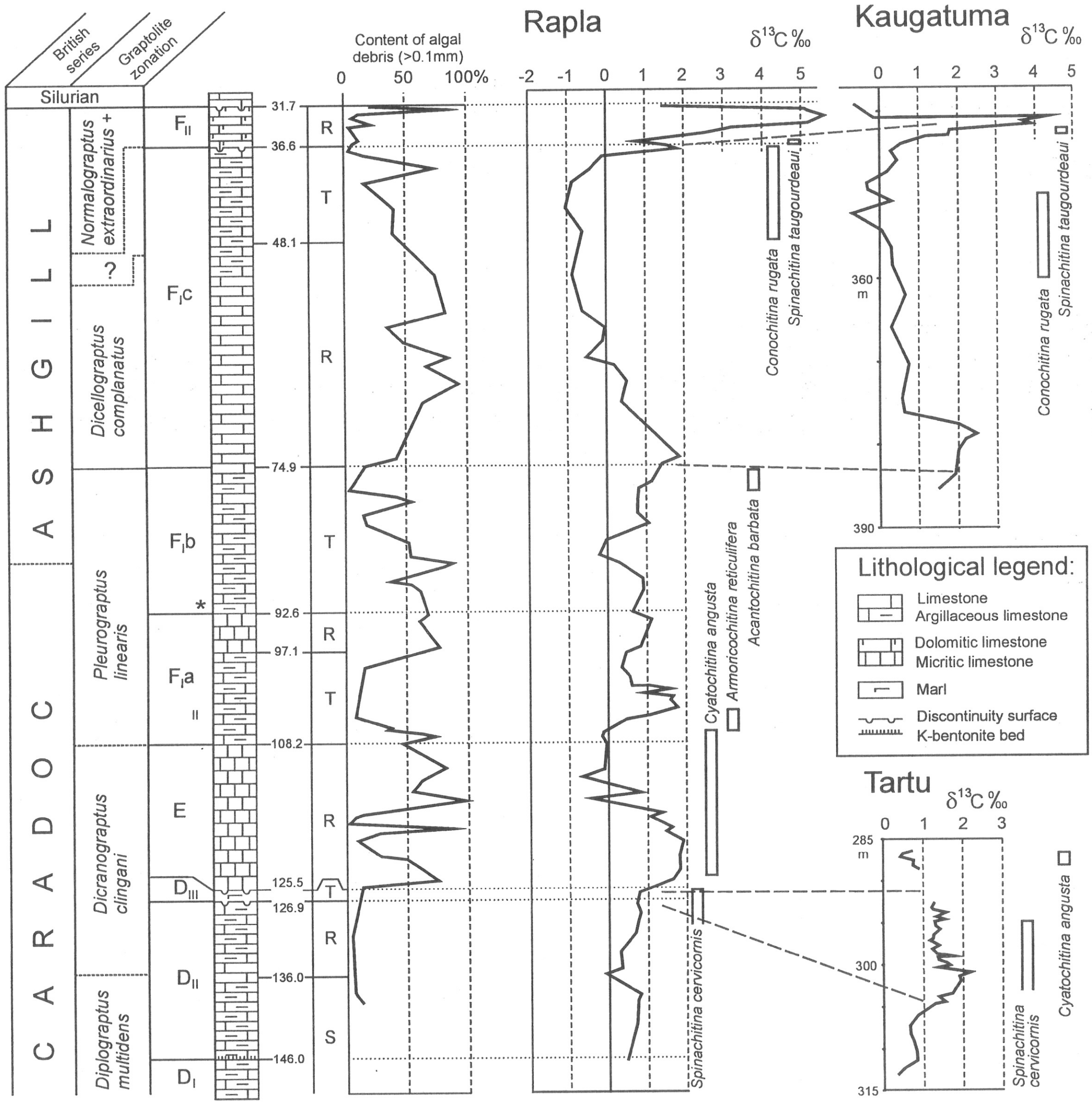

Fig. 2. Correlation of the carbon isotope curves and stratigraphy of the Rapla core. stratigraphical indexes of the Baltic stages: D_{I} – Haljala, D_{II} – Keila, D_{III} – Oandu, E – Rakvere, $F_{I}a$ – Nabala, $F_{I}b$ – Vormsi, $F_{I}c$ – Pirgu, F_{II} – Porkuni. Letters R, S, T, intervals with prevailing 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 $\delta^{13}C$ 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 $\delta^{13}C$ 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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