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The Middle Caradoc Facies and Faunal Turnover in the Late Ordovician Baltoscandian palaeobasin

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Abstract

A Late Ordovician episode of remarkable biotic, climatic, sea level and facies changes, named here as the Middle Caradoc Facies and Faunal Turnover, took place in the Baltoscandian area. This paper presents an integrated overview of these changes in the critical middle Caradoc interval. Data are given on carbonate rock composition, distribution and grain-size composition of the siliciclastic material and the carbon isotopic composition of whole-rock carbonates in cores of Estonia and Sweden.

The Middle Caradoc Facies and Faunal Turnover can be described as a succession of related environmental changes. The turnover began with a positive excursion in carbonate $\delta^{13}C$ and continued with sea level changes that led to a sedimentary hiatus on the shelf and a change from carbonate-dominated to siliciclastic-dominated sedimentation in the basin. The turnover ended with an extinction event and associated microfaunal crisis.

The middle Caradoc turnover in Baltoscandia is comparable to a similar succession of changes in North America. The turnover affected two palaeocontinents, and reflects a widespread, possibly global environmental change. Onset of glaciation on Gondwana and/or increased orogenic activity might have initiated the changes in ocean circulation and led to the initial carbon isotope excursion. The following sea level rise and faunal changes affected several different continents.

Keywords: Late Ordovician; Stable carbon isotopes; Faunal change; Sedimentology; Baltica; Estonia; Sweden

1. Introduction

The Baltica palaeocontinent suffered substantial climatic changes during the Ordovician. It drifted from southern high latitudes to the tropical realm (Torsvik et al., 1996), and the Middle Ordovician, temperate-climate carbonate sedimentation changed to Late Ordovician warm-water carbonate sedimentation (Nestor and Einasto, 1997). In the Baltoscandian palaeobasin, the most significant changes in sedimentation (Jaanusson, 1973; Nestor and Einasto, 1997; Ainsaar and Meidla, 2001) and faunal composition (Rõõmusoks, 1972; Hints et al., 1989; Kaljo et al., 1996) took place in the middle Caradoc, in Keila–Rakvere time. The faunal, facies and sea level changes introduced a distinct style of carbonate sedimentation (Männil, 1966; Hints et al., 1989), referred to as the transition from the "unification stage" to the "differentiation stage" of basin evolution (Nestor, 1990; Nestor and Einasto, 1997). Also, a positive carbon

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isotope excursion is reported from middle Caradoc sections of Estonia (Ainsaar et al., 1999). These changes are named here as the *Middle Caradoc Facies and Faunal Turnover*.

In the present study, we report additional evidence for related changes in carbonate-fine siliciclastic sedimentation. We also demonstrate an associated middle Caradoc isotope excursion from southern Estonia and Sweden, the deeper part of Baltoscandian palaeobasin. The main goal of the study is to give an integrated overview of the sedimentological, faunal and isotope events related to the Middle Caradoc Facies and Faunal Turnover. Observed changes in Baltoscandia are compared with available data from North American basins to propose scenarios for a global middle Caradoc change. The turnover scenario is discussed by comparing the middle Caradoc changes with well-studied Late Cambrian and end-Ordovician global oceanic turnover events.

2. Geological setting and stratigraphy

The Baltoscandian epicontinental palaeobasin developed on the Baltica plate during the early and middle Palaeozoic. Ordovician sedimentation rates were slow but constant due to the tectonically stable Baltic Shield. The post-Tremadoc Ordovician of the Baltoscandian area consists of a nearly continuous Arenig to Ashgill carbonate succession (Nestor and Einasto, 1997). The Palaeozoic rocks on the southern margin of the Baltic Shield are exceptionally well preserved. During the Phanerozoic, the burial depth of the studied rocks in Estonia did not exceed 2 km, possibly even 1 km (e.g., Hagenfeldt, 1996; Kirsimäe and Jørgensen, 2000).

The general distribution of fauna and lithofacies delineates large-scale facies zones within the basin (Männil, 1966; Jaanusson, 1976, 1995; Fig. 1). The Estonian Shelf (North Estonian Confacies Belt of Jaanusson, 1976) is characterised by limestones formed in inner to middle ramp settings, mainly below the fair-weather wave base. The Scandinavian Basin (Scanian and Central Baltoscandian Confacies belts of Jaanusson, 1976), including its broad embayment, the Livonian Basin (Livonian Tongue of Jaanusson, 1976) in southern Estonia and western Latvia, is characterised by argillaceous carbonate sedimentation in distal ramp settings. Põlma (1982) distinguished a transitional zone between the two main facies belts in the East Baltic.

The stratigraphy of the Caradoc sedimentary succession of the Baltoscandian palaeobasin is summarized in Fig. 2. Biostratigraphic correlation of the Estonian core sections is based on microfossils, mainly on ostracode data (Meidla, 1996; Ainsaar et al.,

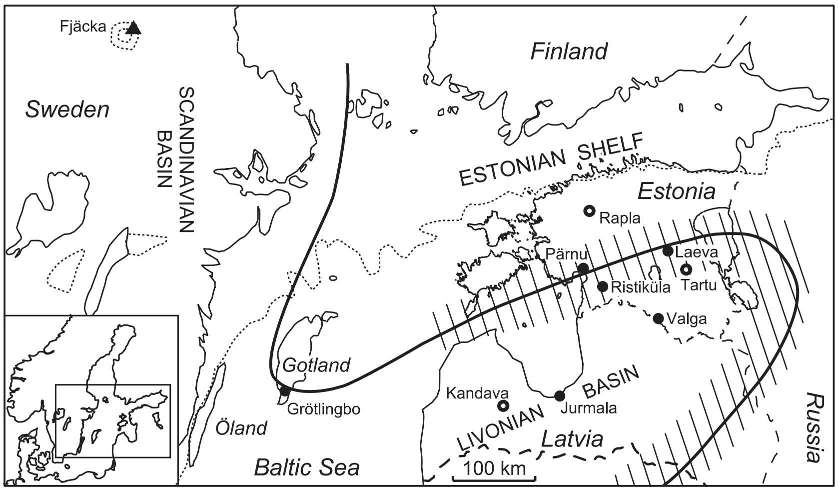


Fig. 1. Location of the studied drillcores (filled circles) and outcrop (triangle), and approximate boundary between the Estonian Shelf and Scandinavian Basin (solid line, modified from Jaanusson, 1976; striped area—transitional facies after Põlma, 1982). Dotted line—limits of the distribution of Ordovician sedimentary rocks; empty circles—other drillcores mentioned in text.

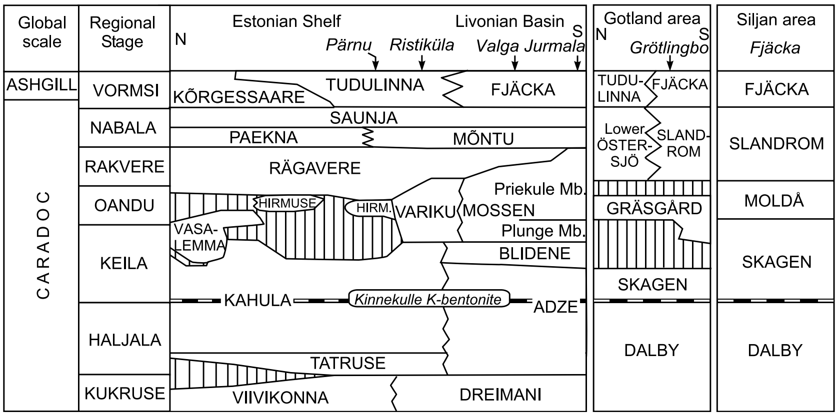


Fig. 2. Stratigraphy of the Caradoc deposits in Estonia (modified from Männil and Meidla, 1994), Gotland and Siljan areas (Sweden; from Nõlvak and Grahn, 1993). Vertical ruling—hiatus; arrows point at the approximate positions of the studied sections.

1999). Although details of the stratigraphic correlation between the Estonian Shelf and Scandinavian Basin (including Livonian Basin) sections is a matter of discussion at several intervals, the following summary can be made.

The lower boundary of the Keila Stage in Baltoscandia can serve as a key horizon. It is tied to the prominent Kinnekulle K-bentonite bed (Bergström et al., 1995). This marker horizon occurs within the Kahula and Adze formations of the East Baltic and at the base of the Skagen Limestone in Sweden (Jaanusson, 1995). Correlation of these three units is suggested by chitinozoa and ostracode evidence (Nõlvak and Grahn, 1993, and this paper), and it is verified by the bentonite occurrences (Fig. 2). Recent ostracode data also suggests Keila age for the calcareous claystone of the Blidene Formation (Meidla, 1996; Ainsaar and Meidla, 2001; Fig. 2).

The available biostratigraphic information suggests a late Keila–early Rakvere age for the Variku Formation (calcareous claystone and siltstone) in the transitional zone, and for the Mossen Formation (organic-rich claystone and calcareous claystone) in the Livonian Basin (Meidla, 1996, 2001; Fig. 2). According to the ostracode data and comparison of stable isotope curves, the Variku and Mossen formations must be considered time equivalents (Ainsaar et al., 1999; Ainsaar and Meidla, 2001). Both units, together with the Blidene Formation, are extraordinary within the Middle Ordovician–Silurian carbonate succession of Baltoscandia because of the prevailing siliciclastic component. In the Siljan area (central Sweden), the Moldå Limestone (marl) occurs in a similar stratigraphic position (Nõlvak and Grahn, 1993). Biostratigraphic correlations and detailed comparison of stable isotopic curves suggest their absence of the upper part of the Keila Stage in Estonian Shelf sections (Ainsaar et al., 1999; Kaljo et al., 1999; Meidla et al., 1999). In the Gotland area, along the westernmost margin of the Estonian Shelf, the most prominent hiatus in the succession is recognised in a similar position, at the top of the Skagen Limestone (Keila Stage; Fig 2; Nõlvak and Grahn, 1993) indicating the overall similarity of basin development in both areas. The bioclastic grainstone with mud mounds of the Vasalemma Formation occur in a limited area in northwestern Estonia, and represents shoal accumulations of different age (late Keila to early Oandu, partly pre-hiatus; Hints and Meidla, 1997; Fig. 2).

Sediments of different age and composition overlie the late Keila stratigraphic gap in shelf areas: the Gräsgård Siltstone (calcareous siltstone; uppermost Keila and Oandu stages) in the Gotland area, and the Hirmuse Formation (marl, Oandu Stage) and the Rägavere Formation (pure micritic limestone; upper Oandu and Rakvere stages) in northern Estonia. Siliciclastic-poor limestones of the Rakvere and Nabala stages (Rägavere, Paekna, Mõntu and Saunja formations in Estonia, Lower Östersjö Limestone in northern Gotland, Slandrom Limestone in central Sweden) have a notably wider distribution across different facies zones (Fig. 2; Nõlvak, 1997; Nõlvak and Grahn, 1993).

3. Material and methods

Data from the middle to late Caradoc (Haljala, Keila, Oandu, Rakvere, and Nabala stages) interval through six core sections from Estonia, Latvia and Sweden (Gotland), and the Fjäcka outcrop in central Sweden are presented (Fig. 1). For biostratigraphic correlation, ostracodes were separated by the disintegration method described by Meidla (1996). After description, 20–50 g rock samples were dissolved in dilute (3.5%) hydrochloric acid to remove the carbonate component. Insoluble residues were fractionated by gravity sedimentation and sieving into size fractions of <2, 2–8, 8–16, 16–63 and >63 μm. Semiquantitative data on the mineralogical composition of the carbonate component in the whole-rock samples was obtained by X-ray diffraction (XRD) using the DRON-3M diffraction system (Mn-filtered Fe-Kα radiation). Calcite 3.04 Å and dolomite 2.89 Å XRD peak intensities were corrected with the proportional factors 0.7 and 0.6, respectively, for calculating the calcite/dolomite ratio. Electron images and energy dispersive spectrometry were used for analyses of selected areas of thin sections from a few samples.

Whole-rock samples were crushed and material for isotopic analysis was selected avoiding obvious veins or burrows. A total of 157 samples for carbonate stable isotopes was analysed using the Finnigan-MAT "Delta E" mass spectrometer at the Institute of Geology at Tallinn Technical University. The samples were powdered to <10 μm and treated with 100% phosphoric acid at 100 °C for 15 min (see Kaljo et al., 1997). Carbon isotope results are reported, using the usual δ notation, as per mil deviation from the PDB standard. Reproducibility of replicate analyses was generally better than 0.1‰. Data on oxygen isotopes from whole-rock samples are not discussed here, as those are not considered sufficiently reliable.

4. Facies distribution

Mixed carbonate-fine siliciclastic sediments are common in the Haljala, Keila and Oandu stages of the study area. Some lithologies are predominantly fine siliciclastics with a minor carbonate content (Blidene, Mossen, Variku formations; Fig. 2) while carbonate material prevails in other sediments (Adze, Kahula formations). The Rakvere and Nabala stages are characterised by relatively pure limestones (Rägavere, Mõntu formations). Changes in the regional facies pattern during the middle Caradoc turnover are also reflected in the distribution and grain size composition of siliciclastic material (Figs. 3 and 4).

In Haljala–early Keila strata (Fig. 4A), the basinal argillaceous carbonates (Adze Formation, Valga drillcore; Fig. 3) differ from the coeval shelf carbonates (Kahula Formation, Pärnu, Ristiküla drillcores; Fig. 3) mainly in the increased abundance of insoluble residues and the composition of skeletal debris (Põlma, 1982). The composition of the insoluble residue of limestone (wackestone/packstone) of these two formations is similar, comprising about 50% clay fraction (<2 μm; mainly illite), 20–30% fine silt (2–16 μm) and 20–30% medium and coarse silt (16–63 μm; Fig. 3; Ainsaar and Meidla, 2001). The Skagen Limestone in the Gotland area (Grötlingbo section; Fig. 3) is lithologically similar to the Kahula Formation in western Estonia.

The absence of late Keila sediments (Fig. 4B) across the shelf in northern Estonia and the Gotland areas (Pärnu, Grötlingbo drillcores; Fig. 3) can be interpreted as the result of non-sedimentation. The hardground surface on the top of the Keila Stage in these areas shows no evidence for subaerial exposure or extensive erosion (e.g., karst or channel features). However, the occurrence of shoal sediments in northwestern Estonia (Vasalemma Formation) below the hiatus indicates that a sea level drop in late Keila time may be responsible for the hiatus (Nestor and Einasto, 1997; Ainsaar and Meidla, 2001). This regression terminated the slow seaward progradation of late Haljala–Keila highstand argillaceous carbonates (Ainsaar and Meidla, 2001). The following latest Keila–early Oandu transgression(s) probably inhibitated the carbonate sedimentation and contributed to the hiatus. Several similar Upper Ordovician unconformities in North American midcontinent platform have been described as drowning omission surfaces caused by sea level rise and submergence of the carbonate shelf (Kolata et al., 1998, 2001). The Keila–Oandu boundary in Estonian Shelf sections

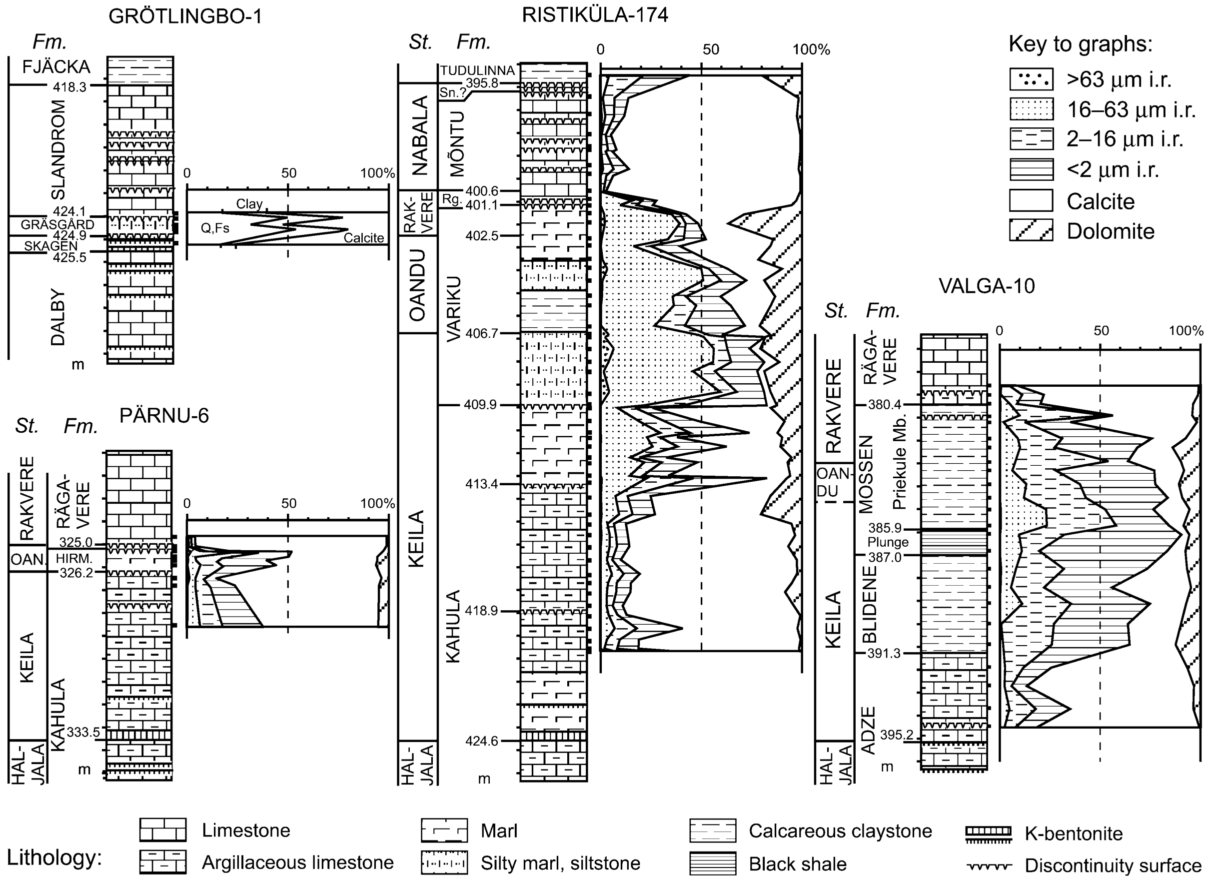


Fig. 3. Insoluble residue grain size composition and carbonate mineralogy of rocks in the Pärnu, Ristiküla and Valga drillcores, and semi-quantitative mineralogy (XRD data) of rocks of the Grötlingbo drillcore. Abbreviations: Sn., Saunja; Rg., Rägavere. Sample levels are shown by black boxes.

can be interpreted as a sequence boundary (in terms of Van Wagoner et al., 1988).

In the Livonian Basin, deposition of fine, and in part organic-rich, siliciclastics (Blidene and Mossen formations) replaced carbonate-dominated sedimentation (Adze Formation, Valga drillcore; Figs. 3 and 4B) in late Keila time. Silty quartz-rich sediments (Variku Formation, Ristiküla drillcore; Fig. 3) formed a 30–50 km wide belt in southern Estonia, bordering the clay-dominated facies within the Livonian Basin (Fig. 4B). Medium and coarse silt (16–63 μm), mainly quartz, comprises 20% (in clay beds) to 80% (in siltstone beds) of the insoluble residue of the Variku Formation, whereas fine sand (63–125 μm) reaches 10% in siltstone beds (Fig. 3). The Blidene and Mossen formations in southern Estonia are composed mainly of clay and fine silt with a low content of medium and coarse silt (0–20%; Fig. 3; Ainsaar and Meidla, 2001). The abundance of erratic boulders of siltstone or silty limestone (Gräsgård Siltstone; Fig. 3) on Öland, with a fauna indicative of both Keila and Oandu age (Martna, 1955; Schallreuter, 1977) suggests that this facies belt extended up to the western part of the present Baltic Sea in late Keila time (Fig. 4B; Ainsaar and Meidla, 2001). The base of silty sediments (Variku Formation) and black shale (Mossen Formation) can be interpreted as a transgressive surface and sequence boundary in Livonian Basin correlative to the hiatus in the Estonian Shelf (Ainsaar and Meidla, 2001). Thus, the upper Keila–Oandu fine siliciclastic accumulations in the basin formed mainly during sea level rise and drowning, which caused the

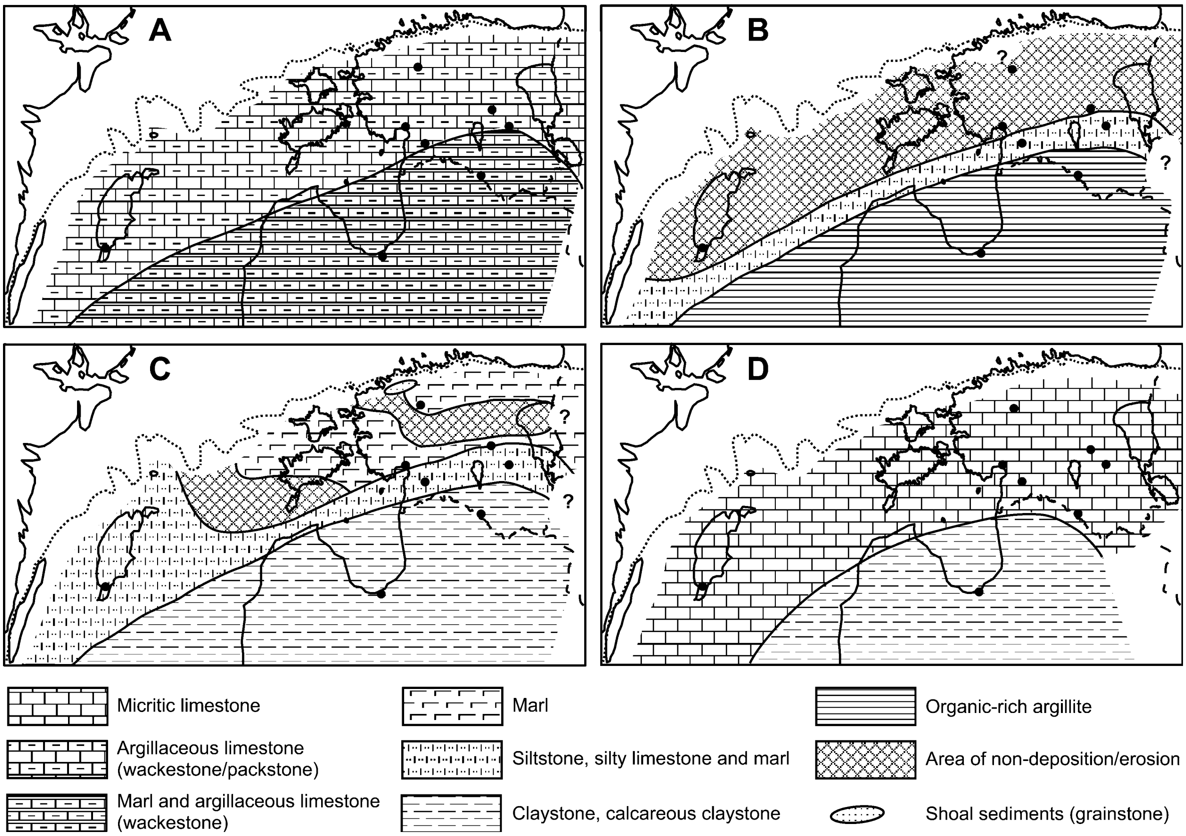


Fig. 4. Distribution of sediments and facies belts (Ainsaar and Meidla, 2001). (A) Early Keila time; (B) late Keila time; (C) Oandu time (the latest part excluded); (D) Rakvere time. Filled rings—drillcores.

carbonate sediment starvation and siliciclastic mud deposition.

Mixed carbonate–siliciclastic sedimentation was restored on the Estonian Shelf in early Oandu time (Fig. 4C). A discontinuous bed of marl with prevailing clay and fine silt fractions among the non-carbonate component (Hirmuse Formation), overlies the Keila–Oandu unconformity in two distinct areas (Fig. 4C). In western and central Estonia (Pärnu drillcore; Fig. 3), this marl grades basinward into silt-rich siliciclastic beds (Variku Formation, Ristiküla drillcore). The silty carbonate facies of the Gräsgård Siltstone is characteristic of the Gotland area in Oandu time (Grötlingbo drillcore; Figs. 3 and 4C). The non-carbonate component of the Gräsgård Siltstone is very similar to that of siltstone beds of the Variku Formation (Fig. 3), and the Gräsgård Siltstone is considered as an extension of the siltstone facies of the Estonian Shelf into the Gotland area (Ainsaar and Meidla, 2001).

In latest Oandu and Rakvere times (Fig. 4D), pure calcareous mud covered the older sediments on the Estonian Shelf. The micritic limestones of the Rägavere Formation contain a minor amount of non-carbonate material, mainly clay and fine silt (Pärnu, Ristiküla, Valga sections; Fig. 3). Similar sediments are distributed in the northern Gotland area (Lower Östersjö Limestone; Nõlvak and Grahn, 1993) and finely nodular or bedded micritic limestone (Slandrom Limestone; Grötlingbo drillcore; Fig. 3) is widespread in the neighbouring areas of the Scandinavian Basin. Deposition of clay-rich sediments (upper part of the Priekule Member) only continued within the Livonian Basin during Rakvere time. Deposition of pure lime mud across the entire shelf in Rakvere time can be interpreted as highstand sedimentation following late

Keila–Oandu transgression(s) (Ainsaar and Meidla, 2001).

5. Faunal changes

The main faunal change in the Caradoc was documented in the Estonian Shelf sections at the level of the Keila–Oandu unconformity in 1950s (Oraspõld and Rõõmusoks, 1956, Rõõmusoks, 1972; Hints et al., 1989; Kaljo et al., 1995, 1996; Meidla, 1996). This stratigraphic level was considered one of the most important faunal change horizons in the post-Tremadoc Ordovician. A following episode of rapid diversification in the Oandu Stage produced a remarkable change in the taxonomic composition of many groups. Several immigrant taxa appeared among brachiopods, trilobites, ostracodes and conodonts, although some of them were only short-lived (Männik, 1992; Meidla, 1996; Hints and Rõõmusoks, 1997). The appearance of tabulates and stromatoporoids, previously not recorded in the East Baltic area, is also related to this diversification event (Mõtus, 1997; Nestor, 1997). Of shelly fossil groups, only trilobites seem to be less affected; their distribution pattern displays a gradual decrease in diversity from the early Caradoc (Rõõmusoks, 1997).

Coeval changes in the taxonomic composition of organic-walled microfossils (chitinozoans and scolecodonts) are less distinct on average but are still remarkable in the dynamics of the particular groups (Kaljo et al., 1996; Meidla et al., 1999). A short-term (about 1 Myear) low-diversity episode of organic-walled microfossils following the faunal change at stage boundary is recognised in the Oandu Stage in Estonia (Oandu crisis: Kaljo et al., 1995, 1996).

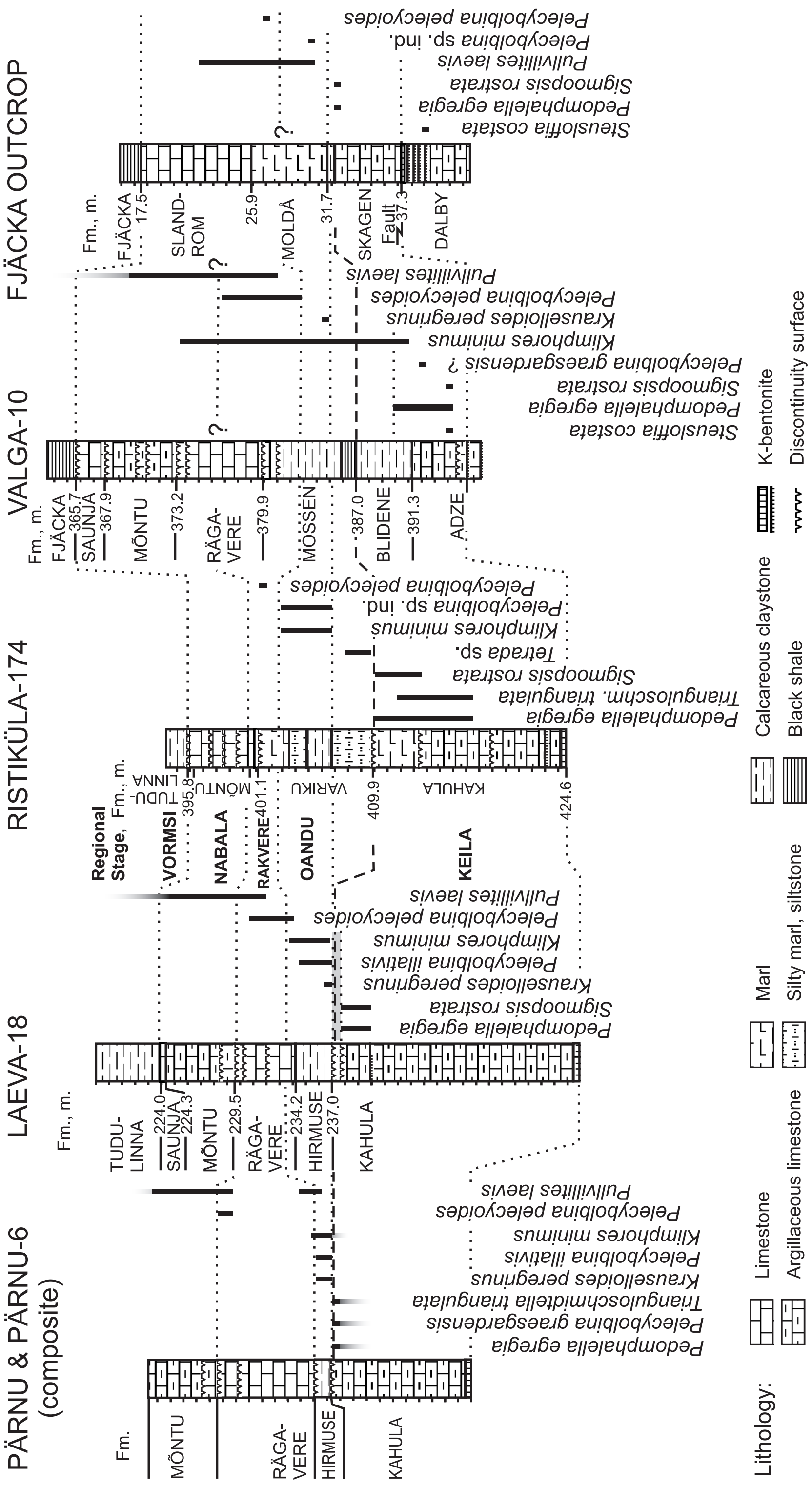
In basinal and transitional sections located in the subsurface of southern Estonia, the turnover is only partially documented, primarily due to the scarcity of the macrofossils. The macrofaunal and microfaunal record of shallow shelf sections reveals a drop in the diversity of brachiopods, ostracodes, echinoderms, gastropods and bryozoans at the Keila–Oandu stage boundary (see Meidla et al., 1999 for a summary). Changes in the ostracode record are distinct at this level due to a gap in the section (Pärnu composite section in Fig. 5). In deeper water settings, the sections are more complete and faunal change is less distinct. The transitional strata are characterised by impoverished assemblages, like the so-called *Tetrada* fauna in the lower part of the Variku Formation (Ainsaar et al., 1999; Ristiküla-174 section in Fig. 5). The appearance of this assemblage is ascribed to a facies change involving the appearance of silty siliciclastics in the basin margin area. Transitional strata with the impoverished faunal assemblage comprise the lower part of the Variku Formation and the lower part of the Mossen Formation (the Plunge Member and lower part of the Priekule Member; Figs. 2 and 5).

6. Stable isotopes

Carbon and oxygen isotope changes have been found to be associated with biotic crises (Holser et al., 1995) and with sea level fluctuation events in several Phanerozoic intervals, including the end-Ordovician and Silurian (Brenchley et al., 1994; Kaljo et al., 1997). Trends in the oceanic carbon stable isotopic composition ($\delta^{13}C$) are considered to be indicators of variations in organic carbon burial rates and/or primary bioproductivity.

A late Keila positive carbon isotope excursion has been reported from stratigraphically complete core sections in the Livonian Basin (Ainsaar et al., 1999). The carbon isotopic curves of the Keila Stage interval in the southern Estonian and Latvian drillcores (Ristiküla, Valga, and Jurmala; Fig. 6; Tartu; Ainsaar et al., 1999) show great similarities. The lower part of the Keila Stage has $\delta^{13}C$ values between +0.2‰ and +0.8‰, slightly decreasing upward. A significant increase in $\delta^{13}C$ values up to +2‰ is recorded in the upper part of the Keila Stage, below the base of the Variku and Mossen formations. A similar positive excursion in $\delta^{13}C$ values has been recorded 2–3 m below the Mossen Formation in western Latvia (Kandava drillcore; Brenchley et al., 1996). The base of the same positive shift occurs in the Pärnu drillcore 1 m below the Keila/Oandu unconformity, but the peak of the isotope excursion is absent due to a hiatus (Ainsaar et al., 1999). The $\delta^{13}C$ values gradually increase upwards in uppermost part of the Keila Stage, the Oandu Stage and the Rakvere Stage in the Ristiküla and Tartu drillcores (Fig. 6; Ainsaar et al., 1999).

The $\delta^{13}C$ curve of the Fjäcka outcrop in central Sweden is very similar to curves from the southern


PÄRNU & PÄRNU-6 (composite)
LAEVA-18
RISTIKÜLA-174
VALGA-10
FJÄCKA OUTCROP
Regional Stage
VORMSI
NABALA
RAKVERE
OANDU
KEILA
Fm.
Fm., m.
MÕNTU
RÄGA-VERE
HIRMUSE
KAHULA
TUDU-LINNA
SAUNJA
VARIKU
FJÄCKA
MÖSSEN
BLIDENE
ADZE
SLAND-ROM
MOLDÅ
SKAGEN
Fault
DALBY
Pedomphalella egregia
Pelecybolbina graesgardensis
Trianguloschmidtella trianguiata
Krausellodes peregrinus
Pelecybolbina illativis
Klimphores minimus
Pelecybolbina pelecyoides
Pulviliites laevis
Sigmoopsis rostrata
Trianguloschm. trianguiata
Tetrada sp.
Pelecybolbina sp. ind.
Steusloffia costata
Pelecybolbina graesgardensis ?
Lithology:
Limestone
Argillaceous limestone
Marl
Silty marl, siltstone
Calcareous claystone
Black shale
K-bentonite
Discontinuity surface

Estonian Ristiküla and Tartu drillcores (Fig. 6). The positive δ^{13}C shift of Keila age begins in the Fjäcka outcrop 1 m above the Kinnekulle K-bentonite, whereas in Estonian drillcores that change, it occurs 6–7 m above this marker (Fig. 6). This suggests a gap at the lower part of the Skagen Limestone in the Fjäcka outcrop, perhaps due to local faulting documented on the level of K-bentonite (Martna, 1955). The positive late Keila δ^{13}C excursion is recorded throughout the Skagen Limestone, with a small decline in its upper part. A second rise of δ^{13}C values occurs at the base of the Moldå Limestone, followed by a gradual decrease through the Moldå Formation and the lower part of the Slandrom Limestone, similar to the δ^{13}C curve in the Oandu and Rakvere stages in the Ristiküla and Tartu drillcores (Fig. 6; Ainsaar et al., 1999). These isotope data support the biostratigraphical correlation of Estonia and central Sweden successions by Nõlvak and Grahn (1993).

A significant positive δ^{13}C excursion was found in the pure limestones of the Rakvere–Nabala boundary interval in the Valga drillcore (Fig. 6). No similar isotope trends have been found in Ristiküla and Jurmala drillcores and Fjäcka outcrop. Two positive carbon isotope shifts have been recorded from the Rakvere–Nabala interval of the Rapla drillcore in northern Estonia (Kaljo et al., 1999), but the correlation of these excursions with that found in the Valga drillcore remains unclear. The Rakvere–Nabala positive excursion has not been found in other sections where the succession is incomplete or difficult to sample due to sedimentary gaps or condensed sections. Another positive shift of δ^{13}C values at the top of the Nabala Stage can be traced in all studied drillcores (Fig. 6).

These data can be interpreted as a basin-wide positive isotope carbon excursion starting in the middle part of the Keila Stage. The peak of the excursion is correlated with the upper part of the Keila Stage, and the curve shifts to lower carbon isotopic values in the uppermost Keila and Oandu stages (Fig. 6; Ainsaar et al., 1999). Thus, the carbon isotopic excursion preceded the main faunal change at the Keila–Oandu stage boundary. The peak of the excursion is also below the upper Keila sequence boundary (base of the Mossen and Variku formations), marking the major sea level changes in stratigraphically complete sections.

7. Regional versus global turnover

In the Baltoscandian palaeobasin, the middle Caradoc turnover lasted for a few million years, during a first part of the *Dicranograptus clingani* graptolite zone (Fig. 7). It included a succession of events: (1) a change in the carbon isotopic composition of seawater in Keila time (Ainsaar et al., 1999); (2) rapid late Keila to early Rakvere sea level changes (sea level fall followed by one or more transgressions) that led to a hiatus on Estonian Shelf and to change from carbonate-dominated to siliciclastic-dominated sedimentation in Scandinavian Basin during late Keila and Oandu time (Ainsaar and Meidla, 2001); (3) the appearance of new fauna at the boundary between the Keila and Oandu stages (Meidla et al., 1999) that was followed by a microfaunal crisis in Oandu time (Kaljo et al., 1995, 1996).

The magnitude of the faunal change in the Keila–Oandu transition far exceeds the background level of the faunal rearrangement during the Middle–Late Ordovician. Such a sequence of related events suggests a global oceanographic background. Another argument for the global nature of the mid-Caradoc turnover is given by Kaljo et al. (1995, 1996) who suggest a relationship of the mid-Caradoc changes in the Estonian sections with a coeval global drop in the diversity of different invertebrate groups, as noted by Sepkoski (1995). Palaeontological data from the East Baltic suggest that the corresponding fluctuation of global diversity may be a faunal response to real changes on an interregional or global scale and not a sampling artefact (Meidla et al., 1999).

The succession of mid-Caradoc events occurred during gradual climatic change from temperate to tropical during Keila–Oandu time, resulting from

Fig. 5. Biostratigraphic correlation of drillcore sections in Estonia and the Fjäcka outcrop, Sweden. Only selected ostracod species are shown. Estonian sections costitute a profile along the deepening gradient, Estonian Shelf is represented by the Pärnu composite section and the Valga section is in the most basinal position. Grey shading (Laeva-18) marks heavily bioturbated zone with mixed fauna (see Meidla, 1996 for details). Dotted lines—stage boundaries correlations; dashed line—late Keila sequence boundary.

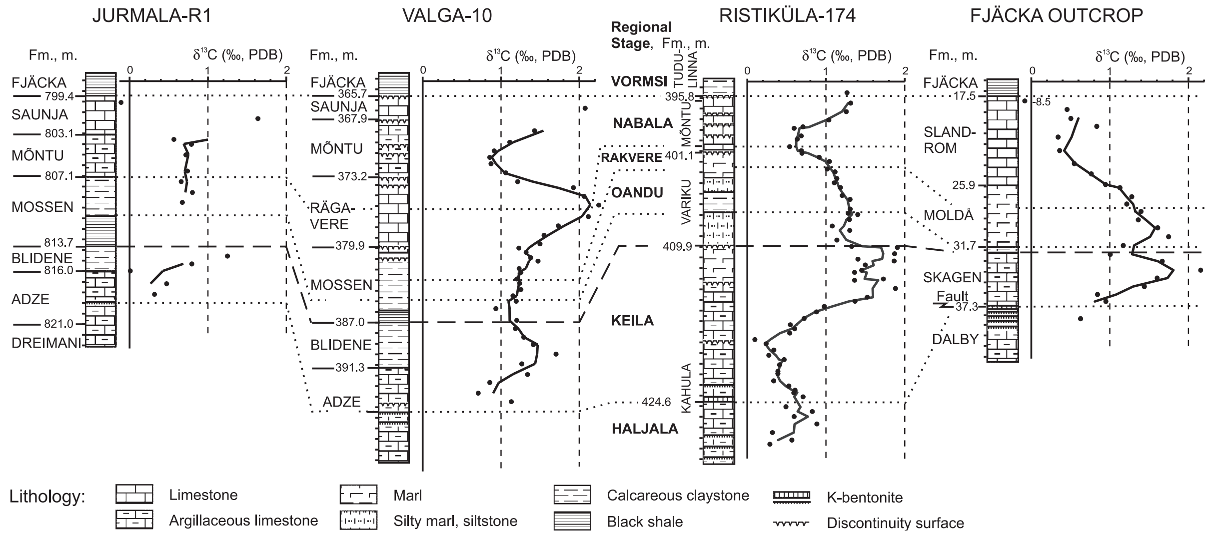


Fig. 6. $\delta^{13}C$ data from southern Estonian and Latvian drillcores, and the Fjäcka outcrop (Siljan area, central Sweden). Raw data (solid circles) have been filtered with a three-point moving average (solid line). Dotted lines—stage boundaries correlations; dashed line—position of transgressive surface coinciding with the late Keila sequence boundary in these basinal sections.

the northward continental drift of Baltica (Webby, 1984; Torsvik et al., 1996; Nestor and Einasto, 1997). Both the faunal and facies changes in the Keila–Oandu interval have been ascribed to a change to a warmer and apparently more arid climate (Jaanusson, 1973; Webby, 1984; Hints et al., 1989). At the same time, most of the Early Palaeozoic has been interpreted as a greenhouse period with high CO_2 levels in the atmosphere (e.g., Gibbs et al., 1997). Greenhouse oceans are thought to have been salinity stratified, with warm, saline deep waters of low oxygen content during most of the Ordovician (Railsback et al., 1990). The reason for the faunal and facies turnover recorded from epeiric sea may be global climate changes which led to changes in ocean circulation.

The global nature of the middle Caradoc turnover is supported by data from the time-equivalent carbonates of North America, where a positive $\delta^{13}C$ excursion, both in kerogen and in bulk carbonates, is described from sections in the mid-continent (Iowa and Kansas; Hatch et al., 1987; Ludvigson et al., 1996; Pancost et al., 1999) and eastern regions (Pennsylvania; Patzkowsky et al., 1997). The highest Caradoc $\delta^{13}C$ values occur slightly above the Millbrig K-bentonite bed, i.e., in similar stratigraphic position as in Baltoscandia. Based on geochemical data, the Millbrig K-bentonite is correlated with the Kinnekulle K-bentonite in Baltoscandia (Huff et al., 1992; Bergström et al., 1995). This suggestion is in accordance with transcontinental biozonal correlation (both beds are in *Diplograptus multidens* graptolite zone; Huff et al., 1992), although recent studies in radiometric dating showed 7 Myears age differences of these K-bentonites (Min et al., 2001). Although the $\delta^{13}C$ excursion in the American mid-continent region was interpreted primarily as a local phenomenon (Hatch et al., 1987; Ludvigson et al., 1996), the isotopic data from Pennsylvania (Patzkowsky et al., 1997) and southern Ontario, Canada (Brookfield, 1988), suggest a more widespread background for this North American excursion. The presence of similar and approximately synchronous isotopic changes in both cases in association with environmental and biotic changes within shelf carbonates of different palaeocontinents, Laurentia and Baltica, argues strongly for a widespread, possibly global oceanographic change.

This similarity is emphasised by parallel trends in sedimentation history in both regions. A change from carbonate-dominated to siliciclastic-rich sedimentation, similar and nearly contemporaneous with the middle Caradoc turnover in Baltoscandia, is described from the eastern part of North America (Holland and

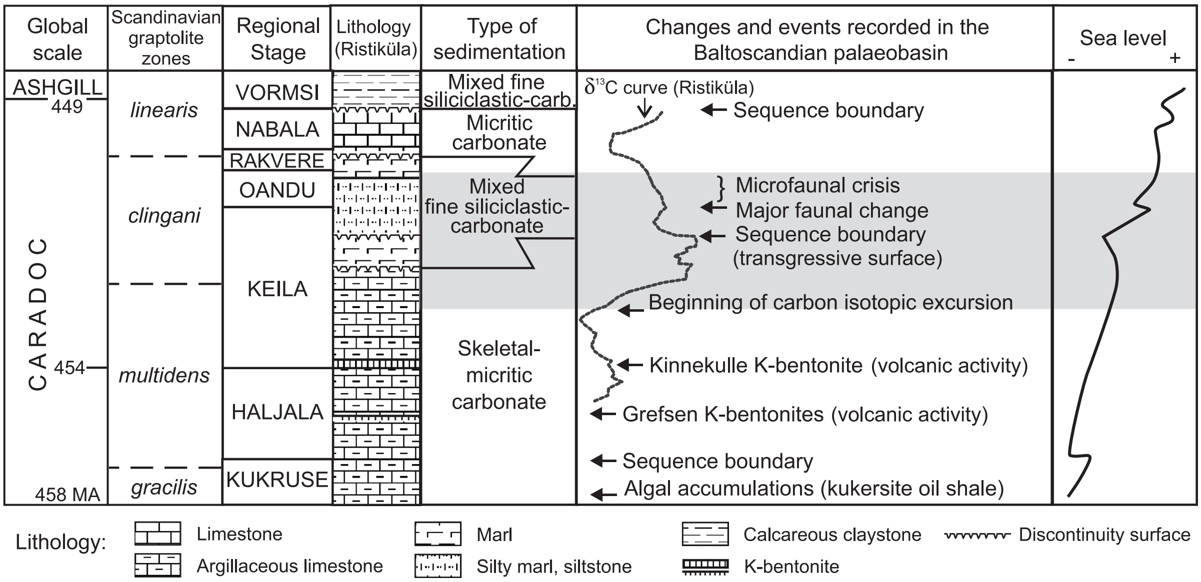


Fig. 7. Relative timing of the sedimentological, geochemical, faunal and sea level changes in the Baltoscandian palaeobasin during Caradoc time. Gray area marks the changes considered here as the Middle Caradoc Facies and Faunal Turnover. A gradual climatic change from temperate to warm-water basin took place during the same period.

Patzkowsky, 1996, 1997; Kolata et al., 1998). This change coincides with major faunal and sea level changes at the middle Caradoc (M4/M5) sequence boundary (Holland and Patzkowsky, 1997), which has been interpreted as a result of a local lowstand exposure followed by a regional transgressive drowning and sediment starvation (Kolata et al., 1998), or by a short-term eustatic fluctuation of the sea level (Holland and Patzkowsky, 1998). If the Millbrig and Kinnekulle K-bentonites are correlative, the M4/M5 sequence boundary in North America and the late Keila sequence boundary in Baltoscandia may have caused by common eustatic sea level change.

8. Other global events in Early Palaeozoic

Two major global Early Palaeozoic positive carbon isotope excursions are known: the Late Cambrian SPICE event (Saltzman et al., 1995, 1998, 2000) and the end-Ordovician Hirnantian event (Brenchley et al., 1994, 2003; Kump et al., 1999). The end-Ordovician excursion and the associated environmental change were clearly related to the glaciation in Gondwana. According to Brenchley et al. (1994, 1995), the change in ocean circulation, global cooling and the glacioeustatic sea level fall changed the carbon cycle and caused a positive excursion in both $\delta^{18}O$ and $\delta^{13}C$. The latter was due to intensive thermohaline circulation, which caused a sudden influx of nutrients to the surface waters, promoting high productivity and/or extensive deposition of the light carbon isotope. Rapid changes in ocean circulation and global temperatures caused a major extinction, which is documented as the primary evidence of the Hirnantian event (Sepkoski, 1995).

The Hirnantian isotopic excursion is a classic example of a positive carbon isotopic excursion explained by productivity-driven changes in the burial flux of organic carbon. Scenarios emphasise the increase in organic carbon burial as either due to an ocean anoxic event accompanied by a high sea level (e.g., Cretaceous Cenomanian–Turonian event; Arthur et al., 1988; Kump and Arthur, 1999), or as due to increased thermohaline circulation accompanied by a glacioeustatic sea level fall (e.g., end-Ordovician event; Brenchley et al., 1994, 1995).

However, a markedly different model to explain positive carbon isotope excursions was developed by Kump et al. (1999) and Kump and Arthur (1999). The model explains some of the positive carbon isotope excursions by increased carbonate platform weathering and ascribes related changes in the isotopic composition to changing riverine runoff during glacioeustatic sea-level lowstands. According to this model, the Hirnantian glaciation was initiated by the Late Ordovician Taconic orogeny, which caused a long-term decline in atmospheric pCO_2 through increased weathering of silicate rocks. The sea level fall exposed low-latitude carbonate platforms to weathering and led to an increase of δ^{13}C in the riverine input to the ocean (Kump et al., 1999). A similar model can be used to explain the Late Cambrian excursion and associated extinction, previously explained as an oceanic anoxic event (Saltzman et al., 1995, 2000).

Although these scenarios have been developed independently, they are not contradictory, and multiple reasons for changes in the carbon cycle are easy to construct.

9. Discussion

The middle Caradoc turnover in Baltoscandia shares some similarities with both the Late Cambrian and end-Ordovician environmental changes. Both these examples are related to eustatic sea level fall and erosion (Brenchley et al., 1994, 1995, 2003; Saltzman et al., 1998, 2000). Similarly, the middle Caradoc carbon isotopic excursion in Estonia and possibly coeval excursion in North America coincided with sedimentary hiatuses, partly described as results of sea level fall (Nestor and Einasto, 1997; Holland and Patzkowsky, 1998).

In Baltoscandia, the middle Caradoc excursion preceded the episode of the most widespread accumulation of organic-rich muds (Plunge Member), similar to the pattern described for the Late Ordovician event by Brenchley et al. (1995). Comparison of faunal changes signals an important difference between the middle Caradoc turnover in Baltoscandia and other Lower Palaeozoic events. In the other events, the first extinction peak is related to the beginning of a positive shift in δ^{13}C values (Late Cambrian event, end-Ordovician event, basal Wenlock (Ireviken) event in the Silurian; Brenchley et al., 1995, 2003; Kaljo et al., 1997, 2003; Saltzman et al., 1998, 2000). In Baltoscandia, the late Keila faunal changes occurred after the δ^{13}C values reached their maximum value. The middle Caradoc event in Pennsylvania has a similar pattern: the excursion in $\delta^{13}C_{carb}$ values reaches its maximum below the boundary between the *Phragmodus undatus* and *Plectodina tenuis* zones, claimed to correlate with the M4/M5 sequence boundary and extinction level (Patzkowsky et al., 1997). In Estonia, the excursion in late Keila time also preceded the shift from carbonate-dominated sedimentation into siliciclastic-dominated sedimentation, whereas in the Late Cambrian, the change in sedimentation type occurred at the beginning of the excursion or, in some areas, during the maximum excursion (Saltzman et al., 1998).

Some differences in the middle Caradoc climatic trends in Baltoscandia and in North America should also be noted. In the Baltoscandian area, the climate shifted to an arid subtropical–tropical condition (Jaanusson, 1973; Webby, 1984; Hints et al., 1989; Nestor and Einasto, 1997). In eastern North America, however, the middle Caradoc changes included a switch from tropical-type to temperate-type carbonates (Brookfield, 1988; Patzkowsky et al., 1997; Lavoie and Asselin, 1998), although the continent was still positioned in the tropical belt (Torsvik et al., 1996). Some authors have argued for global cooling in Caradoc, which changed the sedimentation regime in North America (Lavoie, 1995; Pope and Read, 1998), whereas others explain these changes as an effect of local upwelling transporting cool mid-column ocean water to the shelf (Holland and Patzkowsky, 1997). These seemingly contradictory interpretations can perhaps be resolved by considering the timing of these changes, as the cooler marine conditions, with or without upwelling, reached equatorial North American seas only in late Caradoc, i.e., a few million years after the main turnover (Lavoie and Asselin, 1998).

In spite of some unique features, the middle Caradoc turnover and related isotope excursion could be explained by a scenario similar to the Late Cambrian and end-Ordovician events. A possible controlling factor of the Caradoc turnover may be the fluctuation of the continental ice sheet in Gondwana. Although several recent papers advocate a short-lived end-Ordovician glaciation and lack of direct evidence for a continental glaciation in the Caradoc (Brenchley et al., 1994), several authors (Lavoie, 1995; Pope and Read, 1998) have argued that a glaciation might explain the rapid changes in eustatic sea level and in isotopic composition in the Caradoc. There is also some evidence of continental glaciation in areas of Gondwana (Hamoumi, 1999). If the end-Ordovician glaciation was triggered by increased orogeny around the Iapetus Ocean (as in the weathering model), the middle Caradoc turnover may mark an early shift to the weathering-induced climatic changes. Such timing is highly possible as Taconic Orogeny in North America was initiated in middle Caradoc time (Holland and Patzkowsky, 1997).

10. Conclusions

The Middle Caradoc Facies and Faunal Turnover in the Baltoscandian palaeobasin can be described as a succession of related changes in sea-water chemistry, sea level, fauna and sedimentation. The turnover began with a positive shift in carbonate $\delta^{13}C$ in Keila time. The isotope excursion reached its maximum before rapid sea level changes led to a sedimentary hiatus on the shelf (sequence boundary) and to change from carbonate-dominated to siliciclastic-dominated sedimentation in the basinal part of the Baltoscandian sea. The turnover ended with the extinction event and following microfaunal crisis (Oandu crisis).

A similar succession of middle Caradoc isotope, faunal and sea level changes has been recorded in North America. Described in two palaeocontinents, Baltica and Laurentia, the middle Caradoc turnover reflects a widespread, possibly global environmental change. Onset of glaciation on Gondwana and/or increased orogenic activity might have initiated the changes in ocean circulation and led to the widespread carbon isotope excursion. The initial events were followed by a sea level rise together with an invasion of anoxic bottom waters into the shelf area, and by global faunal changes.

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