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Paleohydrography of the Great Belt, Denmark, during the Littorina Transgression: the isotope signal

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WINN, KYAW; ERLENKEUSER, HELMUT; NORDBERG, KJELL; GUSTAFSSON, MIKAEL, 1998: Paleohydrography of the Great Belt, Denmark, during the Littorina Transgression: the isotope signal. (Paläohydrographie des Großen Belts während der Littorina-Transgression: das isotopische Signal). – Meyniana, **50**, 237–251, 6 Fig., Kiel.

The marine transgression into the Baltic Sea through the Great Belt took place around 9,370 calibrated ^{14}C-years B.P. The sedimentary sequence from the early brackish phase and the change to marine conditions has been investigated in detail through ^{14}C-datings, and oxygen and carbon isotope measurements, and is interpreted by comparison with modern analogs.

The oldest brackish sediments are the strongly laminated clays and silts rich in organic carbon followed by non-laminated heavily bioturbated silts. The bedding and textural characteristics and stable isotope analyses on *Ammonia beccarii* (dextral) and *A. beccarii* (sinistral) show that the depositional conditions respond to a change at about 9,100 cal. a B.P. from an unstratified brackish water environment in the initial stage of the Littorina Transgression to a thermohaline layered milieu in the upper unit. The oxygen isotope results indicate that the bottom waters of this latter period had salinities and temperatures comparable to the present day Kiel Bay waters. The isotopic composition of the total organic carbon and the δ^{13}C-values of *A. beccarii* reveal a gradual change from an initially lacustrine/terrestrial provenance toward a brackish/marine dominated depositional environment. A stagnation of the sea level at around 9,100 to 9,400 a. B.P. is indicated.

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Kurzfassung

Die marine Transgression in die Ostsee durch den Großen Belt fand ca. 9,370 Jahre B.P. (kalibrierte ^{14}C-Jahre) statt. Die Sediment-Sequenz von den frühen Meerwasser-Ingressionen bis zu den endgültigen marinen Bedingungen wurde detailliert mit Hilfe von ^{14}C-Datierungen

und der Messung stabiler Isotope an benthischen Foraminiferen untersucht, wobei die Interpretation durch Messungen an rezenten Analogverhältnissen gestützt wurde.

Die ältesten brackischen Sedimente sind laminierte Tone und Silte mit hohen C_{org}-Gehalten. Sowohl die Charakteristik der Schichtung wie die Korngrößeneigenschaften und die an *Ammonia beccarii* (sinistral) und *A. beccarii* (dextral) durchgeführten Analysen stabiler Isotope weisen darauf hin, daß diese Serie zunächst in einem Milieu mit ungeschichtetem Wasserkörper entstand, in dem sich dann im Zuge der Transgression eine ausgeprägte Sprungschicht entwickelt hat. Die ^{14}C-Altersbestimmungen ergaben 9.100 cal. a B.P. Die Ergebnisse der Sauerstoffisotopenanalysen deuten darauf hin, daß Salzgehalte und Temperaturen des Bodenwassers unter der thermohalinen Sprungschicht denen der heutigen Kieler Bucht vergleichbar waren. Das kältere, sauerstoffreiche Bodenwasser führte zu verstärkten biologischen Aktivitäten, wie sich an der vollständigen Zerstörung der Feinschichtung bzw. Durchmischung des Sedimentes zeigt. Die Isotopenzusammensetzung der organischen Gesamt-Kohlenstoff-Fraktion sowie die $\delta^{13}C$-Werte von *A. beccarii* zeigen einen Wechsel von einem terrestrisch/lakustrin zu einem brackisch/marin geprägten Ablagerungsraum.

Introduction

The Holocene marine invasion of the Baltic, commonly termed the Littorina Transgression after the Gastropod *Littorina littorea* LINNÉ, is recorded in the sedimentary succession through a sequence ranging from lake marls to peats, clay gyttjas and brackish to marine clays. The transgression first had an impact on the then existing freshwater lakes and made brackish environments develop before marine conditions were reached with foraminifera species as found today in the Skagerrak and North seas (WINN 1974). The transgression commenced around 9,370 dendrochronologically calibrated years B.P. (8,340 a B.P. conventional ^{14}C-age, ERLENKEUSER *et al.* 1975; calibration after STUIVER & REIMER 1993), when the sea reached the level of the thresholds in the Great Belt Channel at about 27 m below m.s.l. The marine incursions were first held back from the main Baltic Sea by the Darss threshold and remained confined to the Kiel and Mecklenburg Bays (SAURAMO 1958; KOLP 1965). Studies on the Late Weichselian to Recent hydrographic development of the Kattegat and Skagerrak indicated a stratified structure of the water column possibly related to inflow of saline water around 10,000 cal. a B.P. (CHRISTENSEN *et al.* 1993) and showed full attainment of the modern circulation patterns about 8,400 cal. a B.P. (CONRADSEN & HEIER-NIELSEN 1995). WINN *et al.* (1988) found a low salinity event in the Kiel Bay at around 6,800 cal. a B.P. NORDBERG (1991) gave evidence of a hydrographic shift in the Kattegat at around 4,450 cal. a B.P. and two minor shifts at 1,960 cal. a B.P. and 490 cal. a B.P.

At present, a strong thermocline between 15–20 m separates the inflowing saline waters at depth from the outflowing brackish waters at the surface. This situation is particularly pronounced in the late spring and early summer, due to the seasonality of the meteorological cycle. The volume of outflow

(62%) far exceeds that of inflow (DIETRICH 1950), a feature typical of an inland sea in a humid climate (SEIBOLD 1971).

In the sediment succession, the basal brackish clays of the transgression phase are laminated (Unit D; see, e.g., Fig. 3) showing varve-like alternations on the radiographs. These deposits are followed by non-laminated clays (Unit E) with a significantly higher microfaunal diversity. The boundary between these two lithostratigraphic units is also marked by a noticeable change in the relative abundance of the sinistral to dextral *A. beccarii* (WINN 1974). The coiling ratio of these species varieties was applied successfully for paleohydrographic reconstructions in the Great Belt and the Kiel Bay (WINN 1974; WINN *et al.* 1988).

The objective of the present study is to investigate the history of hydrography in the early phase of the Littorina Trangression, to determine the change of the environmental factors related to the changes in the faunal assemblage and in the bedding characteristics. We have determined the stable isotope signature of the benthic foraminifer *A. beccarii* (sinistral and dextral) in order to analyse the temperature and salinity conditions and gain information as to the structure of the water column during this episode.

Cores 12519-2 (55°24.7'N, 10°58.4'E; water depth 25.5 m), 12522-2 (55°22.6'N, 10°56.8'E; 27.2 m) and 12523-1 (55°24.2'N, 10°59.5'E; 24 m) were raised north of the thresholds in the Great Belt, while core 12594-2 (55°5.0'N, 11°1.0'E; 39.6 m) was retrieved from the channel northeast of Langeland Island (Fig. 1). The specimens of the modern *A. beccarii* were collected from surface sediments of the Havstens Fjord, eastern Skagerrak (Sweden) between August 1993 and December 1994.

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Methods

A. beccarii occurred abundantly in the laminated clays of Unit D and non-laminated clays of Unit E in the Great Belt Channel. From the available samples, carbonate shells were concentrated through flotation in CCL_4 (WINN 1974). Sinistral and dextral forms of *A. beccarii* were picked and sorted into 250–400 μm, 400–500 μm and >500 μm size fractions under the microscope. In most cases, about 50 to 200 specimens were available. This large a number greatly smoothed out any natural isotopic variability among the specimens and ensured a high statistical significance of the results. The specimens were cleaned under methanol in a sonic bath for about 10 seconds. A longer sonification would have resulted in an intolerable heavy fragmentation as the shells were very thin and fragile. Most of the stable isotope measurements were made on a VG Micromass 602D mass spectro-

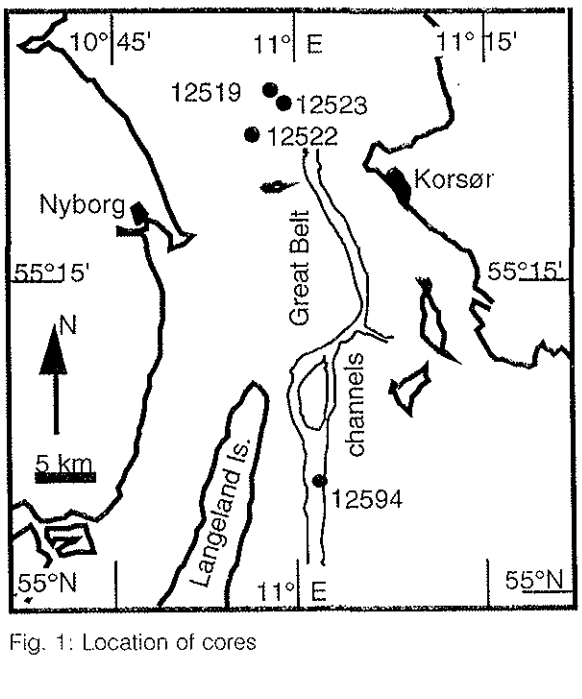

Fig. 1: Location of cores

meter. Non-automated CO_2 gas preparation was applied on individual samples reacting (>200 µg) with 100% orthophosphoric acid at about 50°C under vacuum. Later on, for sediments with limited occurrence *of A. beccarii* and for the modern series, a Finnigan Mat 251 mass spectrometer with the automated Kiel-1 carbonate preparation device was used, which allows analyses down to 10 µg of weight (6 µg in a modern set-up). Again, individual reaction with phosphoric acid is applied (in vacuo, 73°C). Isotopic stability was controlled on a daily basis by replicates of an internal standard (Solnhofen limestone). Calibration to the international PDB-scale was achieved via the NBS 20 international isotope standard. For both analytical procedures, the external reproducibility is better than ±0.10‰ for $\delta^{18}O$ and ±0.05‰ for $\delta^{13}C$.

The radiocarbon data were calibrated by means of the Bidecadal Data Set (Stuiver & Reimer 1993). A correction for a marine reservoir effect has not been applied, as the close atmospheric contact of the Baltic Sea waters will largely cancel any reservoir effect of the different source waters. Moreover, most samples show a significant contribution of terrigenic organic carbon as evidenced by the stable carbon isotopes.

In order to gain a modern environment of some relevance to the paleoisotopic aspect for reference, a series of living and dead specimens of *A. beccarii*, collected monthly from the Havstens Fjord, Sweden, was used for

isotope analysis. Regular salinity and temperature measurements were carried out along with the sediment sampling at the 12 m, 20 m, 30 m and 40 m stations. From these data, a hypothetical isotopic equilibrium composition of calcite was calculated applying the palaeotemperature equation of SHACKLETON (1974). The regression of the oxygen isotope composition of the water with salinity was taken into account as -0.25 ‰/psu (WINN *et al.* 1988), starting with an Atlantic Norwegian Sea water of 35.2 psu (psu = practical salinity units) and δ_w = 0.04‰ (ERLENKEUSER 1985).

Results

Core 12 522 (Fig. 2) penetrated the oldest sediments of this study. The radiocarbon dates show that the freshwater peat and lake marls were deposited between 9,870 and 9,370 cal. a B.P. (ERLENKEUSER *et al.* 1975). The clay unit contained more dextral coiling *A. beccarii* than sinistral specimens (WINN 1974). The oxygen isotope levels in Unit D remained fairly stable and ranged between –5.4‰ and –6.5‰. The carbon isotope ratios also reveal little variability but show a tendency towards the lighter ratios upwards. The youngest $\delta^{18}O$ and $\delta^{13}C$ levels, however, from the topmost centimeter of the core (Unit G) are significantly increased, at –1.95‰ and –1.32‰ respectively. The $\delta^{13}C$ values of the total sedimentary organic carbon measured earlier on this core (ERLENKEUSER *et al.* 1975), indicate a predominantly terrestrial to lacustrine source for the organic carbon of Unit D, but with an increasing autochonous input in the upper part of the sequence.

Cores 12 519 and 12 523 (Figs. 3, 4) represent the stratigraphical contiguity to core 12 522, and are mainly composed of Units D and E, unconformably overlain by thin covers of modern sediments (Unit G). ^{14}C-dating on core 12 519 (ERLENKEUSER *et al.* 1975) indicated that the major part of the section cored was deposited within a relatively short period of about 300 years giving very high sedimentation rates of 0.7 to 1.0 cm/yr and thus provides an excellent time resolution for this interval. The higher ages in the overlying laminated clays of Unit D are likely due to older reworked organic materials, a major part of which appears marine as evidenced by the lighter carbon isotopes of the total sediment ($\delta^{13}C_s$). Resuspension of older deposits, redeposition, and bioturbation are well known processes in the marine realm to provide too high an ^{14}C-age. For the modern Baltic, sediment surface ^{14}C-ages of about 1000 a have been found (ERLENKEUSER 1979).

The oxygen and carbon isotope ratios measured in Unit D in both cores fell within the same range as those in 12 522. Again, the stable isotope figures in the overlying Unit E differ drastically from those of Unit D, and closely group around –3‰ for oxygen and –1.0‰ for carbon. The transition occurs within 10 cm in the ^{14}C-dated core 12 519 (Fig. 3), indicating an abrupt event. $\delta^{13}C$ of total organic carbon showes a similar shift, which,

however, began earlier in the laminated clays of Unit D. The $\delta^{13}C$-values are much higher than in Unit D but do not attain typical marine levels.

Core 12 594, which was taken south of the thresholds from much deeper water (39.6 m), also showed similar shifts as the northern cores in the stable

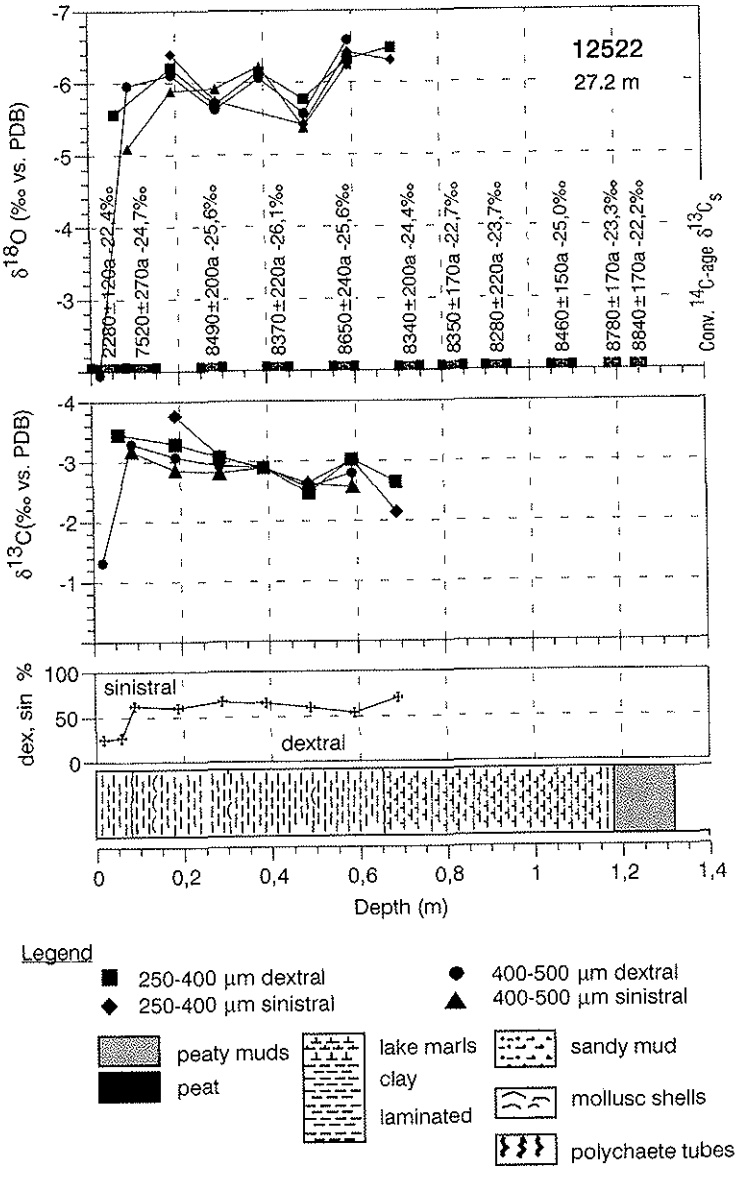

Fig. 2: Isotope results and sedimentary succession in core 12522, Great Belt.

isotope values (Fig. 5) to the heavier fractions across the Units D and E boundary. Despite of the different water depths, the similar isotope levels of the respective units in all the cores indicate a comparatively uniform and consistent bottom water mass in the region. In detail, however, core 12 594 reveals the isotopic transition from the lighter level of Unit D toward Unit E to start below the D/E-boundary.

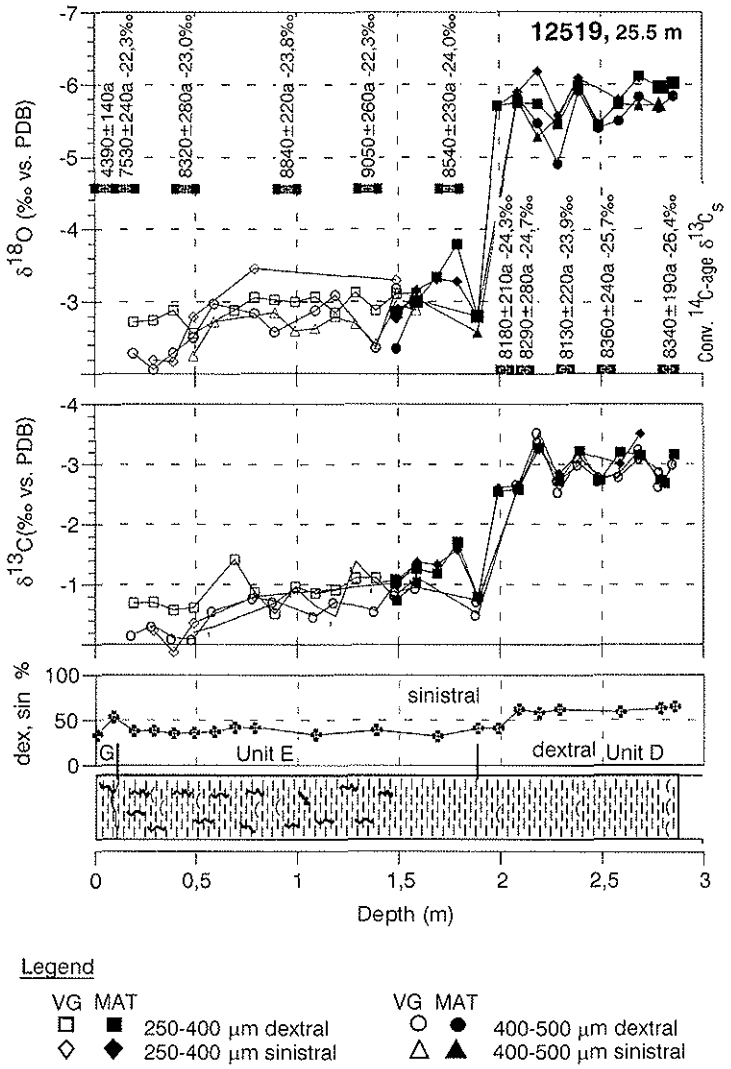

Fig. 3: Isotope results and sedimentary succession in core 12519, Great Belt.

$\delta^{18}O$ of the modern *A. beccarii* from 12 m water depth, Havstens Fjord, E. Skagerrak, show only greatly restrained seasonal variations. Comparing the measured $\delta^{18}O$ values with the hypothetical isotopic equilibrium composition of calcite indicates that the seasonal samples of *Ammonia* show little response to the expected variation (Fig. 5). Regardless of the possible uncer-

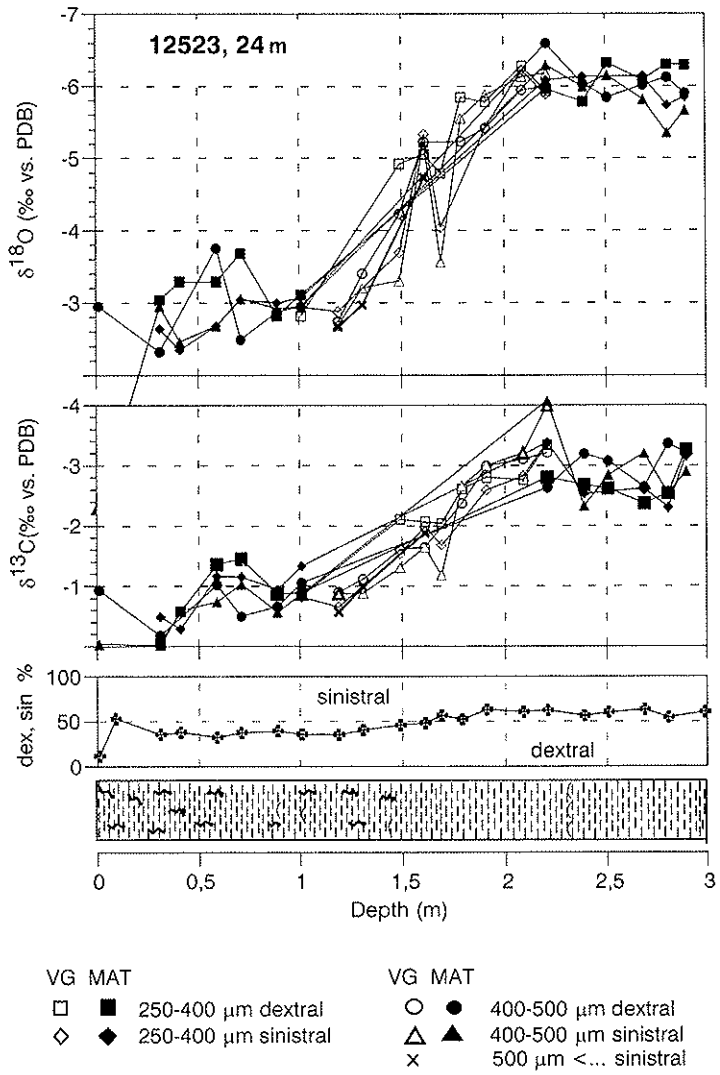

Fig. 4: Isotope results and sedimentary succession in core 12 523, Great Belt.

tainty of the vital effect - the 20 m station results (not shown) restrict this figure to 0.5 to 1‰ - *Ammonia* is very likely to experience its major shell growth and calcification in the late spring/early summer. This level is then retained and carried over into the following year.

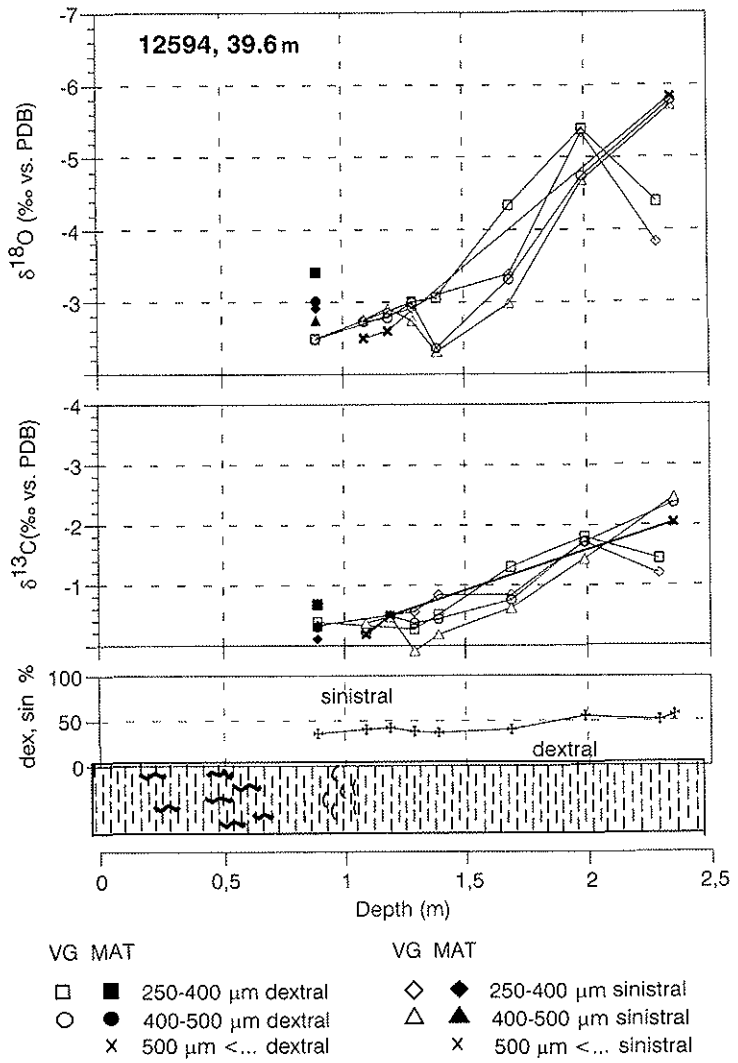

Fig. 5: Isotope results and sedimentary succession in core 12 594, N. Langeland Belt.

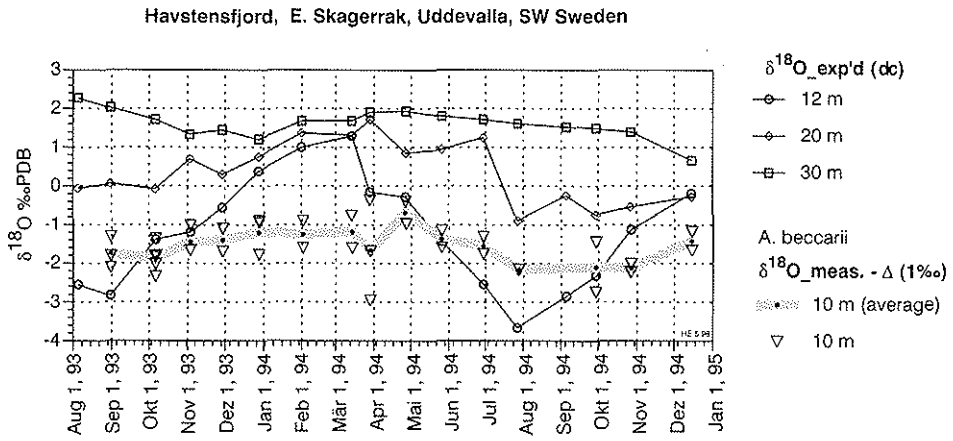

Fig. 6: $\delta^{18}O$ time series of *A. beccarii* in Havstensfjord, eastern Skagerrak, W. Sweden. Measured $\delta^{18}O$ has been corrected by –1‰ for vital offset.

Discussion

The results confirmed the large shift in $\delta^{18}O$ and $\delta^{13}C$ of *A. beccarii* observed at the base of cores 12 864 and 12 880/12 886 in the Kiel Bay (Winn *et al.* 1988). The oxygen and carbon isotope niveaus in Units D and E (Fig. 2–5) are comparable and indicate that these units which depict similar bedding characteristics, are also present in the Kiel Bay.

Investigations of the Holocene sea level rise in the Western Baltic (Kolp 1979; Winn *et al.* 1986) showed that the sea level stagnated or even regressed slightly around 27 m below m.s.l. for a brief period around 9,000 to 9,400 cal. a B.P. Since the investigated cores are situated in a tectonic hinge area stable as to uplift or subsidence (Gutenberg 1941), the present levels in the cores represent, more or less, the original depths of the sediment horizons. Therefore, after the sea-level curve (Winn *et al.* 1986), the water depths in the Great Belt must have been just a few meters around 9,400 a B.P. during the deposition of Unit D at the geographical positions of cores 12 519, 12 522 and 12 523.

The $\delta^{18}O$-levels of *A. beccarii* are lighter by about 3‰ than the present day Kiel Bay values (Winn *et al.* 1988) and by over 4‰ compared with the results from 12 m water depth in the Havstensfjord (Fig. 6). This offset is caused not only by salinity and but also by temperature differences to the modern environments due to the shallowness of the former depositional environment.

The low faunal diversity, the absence of high marine foraminifera species, and the almost exclusive, but abundant occurrence of *A. beccarii* in Unit D

(Winn 1974) indicate that the salinity of the water column was less than in the Kiel Bay today (with an average of 20.3 psu, Rumohr 1979).

By the time of Unit D, the Baltic Sea was a large freshwater lake (Ancylus Lake), and the present-day Kiel and Mecklenburg Bays showed interconnected elongated bodies of freshwater occupying the deep channels (Kolp 1975; Winn & Averdieck 1984). The only freshwater outlet of the then Western Baltic was through the Great Belt. The $\delta^{13}C$ measured on the total organic carbon (Erlenkeuser *et al.* 1975) also indicate the prevalence of a non-marine environment, i.e. organic detritus from lacustrine production or an allochthonous carbon input from terrestrial sources.

Under controlled conditions in the laboratory, *A. beccarii* could reproduce at salinities as low as 13‰ (Bradshaw 1957). Adopting this lower salinity tolerance level, 13‰, for the Great Belt waters during Unit D, the salinity difference (–7‰) from today's average accounts for about –1.8‰ of the total $\delta^{18}O$-effect (–3‰). The remaining amount, –1.2‰, could be attributed to a summer temperature about 5°C higher than at present (cf. discussion for further evaluation).

The stable isotopic ratios became much heavier in Unit E at around 9100 cal. a B.P. (Erlenkeuser *et al.* 1975). From the ^{14}C-datings and the resulting sedimentation rates in core 12519, these changes took place within a very short period of less than 15 years duration. This unit was earlier interpreted to have been deposited under full marine conditions due to the higher faunal diversity with foraminiferal species found today only in the North Sea and the Skagerrak. A drastic and abrupt onset of higher salinity conditions in a time span less than 15 years can be postulated when the mixing effects through bioturbation are taken into account.

In core 12523 (not ^{14}C-dated) this event is spread over 60 cm of the succession with indications of a regressive event during the transition (Fig. 4).

The sea level rose very rapidly during this time to about –10 m below present sea level. With the larger head of water now available, and taking into consideration that the outflowing waters were brackish, such high salinities in Unit E would only be possible when a thermohaline layering existed. The inflow of higher saline waters at depth must have occurred with some continuity, advecting oxygen rich waters at a sufficient rate to allow the benthic biotopes to enhance, as is evidenced by the abundance of bioturbate structures in this unit (Winn 1974).

The $\delta^{18}O$-levels of *A. beccarii* in early Unit E (ca. –3‰) are similar to those attained in the non-laminated sequence of the Kiel Bay cores 12865/12887, 12880/12886 and 12864, and also closely match a modern series of *Ammonia*, collected 1974 (Wefer 1976) at 13 m water depth at Boknis Eck, western Kiel Bay ($\delta^{18}O \approx -2.8$‰; Winn *et al.* 1988). Of the modern series 1993/94 from the Havstensfjord, 12 m station, the measured

values (ca. –0.5‰, i.e. uncorrected for vital offset) are heavier by about 2.5‰ than in Unit E.

As unit E has been deposited during rapid sea level rise, water depth may have attained a level allowing to presume similar temperature regimes in the benthic realm for Unit D and the present time. Accordingly, from the $\delta^{18}O$-difference (–0.2‰) in *Ammonia*, salinity should have been lower by about 1‰ than presently at Boknis Eck (13 psu, 10°C as effective figures for a likely shell growth in May, June 1974; Winn *et al.* 1988). Compared to the 12m-station in the Havstensfjord (23 psu, 8°C for May, June 1993), the difference is –2.5‰ in $\delta^{18}O$ or –12 psu in salinity. Accordingly, both modern references suggest salinities slightly below 15 psu for Unit E. Even if some freedom is allowed for as to the numerical values applied above for the physical parameters, the estimated salinity for the time of Unit E will not change basically.

This salinity level for Unit E is close to the lower tolerance level for *Ammonia*. Accordingly, the lower $\delta^{18}O$ figures of Unit D cannot be taken to signal a still lower salinity. Instead, these values more likely are indicative of a higher temperature. The temperature equivalent of the isotope shift of about -3‰ between Unit E and D is +13°C, so that the calcification temperature of *Ammonia* is found as 23°C based on the modern T,S-pair of 10°C, 13 psu adopted for Unit E. The temperature of 23°C is not unlikely for late spring/early summer conditions in the shallow waters of the sill area in the Great Belt during the early Littorina transgression which fell into the warmth of the early Holocene.

Smaller corrections may become neccessary. One uncertainty is the regression coefficient of $\delta^{18}O$ vs. salinity. The contribution of isotopically light meltwaters from Scandinavia to the Ancylus Lake would still have played some role for Unit D and part of Unit E, thus increasing the steepness of the regression. So, even with temperature and $\delta^{18}O$ being the same for the modern environment and Unit E, the palaeosalinity should be higher, if the freshwater source is given a lighter oxygen isotope composition than it has today. Also the ice cap effect on $\delta^{18}O$ of the marine water would not have decayed completely and the marine source should be given a slightly different composition than today. These points will particularly bear on model salinities if other temperature scenarios are considered.

The carbon isotope ratios also exhibit a distinct shift towards the heavier fractions across the D/E- boundary. It is noteworthy that this change also commenced and ended synchronously with those of the oxygen isotopes. These changes occurred very rapidly in contrast to those documented in the Skagerrak core GIK 15530-4 (Erlenkeuser 1985). In the Great Belt area they are associated with the role of the fresh waters which becomes largely reduced by the marine ingression. Part of these fresh waters originate from

the glacial marls in the southern marginal zone of the last glaciation and show a comparatively high content of dissolved inorganic carbon and with it a light carbon isotope composition. This isotope signature shows up in the limnic biota and its detrital remnants (cf. WEFER *et al.* 1978). Additionally, slowly exchanged bottom waters could rapidly accumulate, within a couple of months, significant quantities of isotopically light, remineralized CO_2 released by degradation of organic matter. The intrusion of salty water at the bottom may have enhanced this carbon-isotopic signal. Probably the thermohalocline formed for the first time around this period.

The relative sea level curve for the Western Baltic (WINN *et al.* 1986) showed that the sea level rose very rapidly, resulting in fast erosion of the surrounding land and vegetation in the Kiel Bay area (cf. Healy & Werner 1987). The sediments also showed higher organic carbon contents. In the Great Belt, an outflow-inflow system similar to the present day is indicated to have been established during this period.

The stable isotope measurements carried out separately on sinistral and dextral *Ammonias* in all the investigated cores do not show any consistent or significant differences to support a lower growth temperature for the sinistral variety, as previously postulated from the coiling ratio (LONGINELLI & TONGIORGI 1960). However, the left coiled forms were far more abundant, and dominated the foraminiferal assemblages of Unit E (WINN 1974).

Conclusions

$\delta^{18}O$ and $\delta^{13}C$ of *A. beccarii* in the sediment cores clearly indicate the beginning of the Littorina Transgression with the onset of brackish water conditions (13‰) at around 9,400 cal. a B.P.

After a period of about 300 years without significant changes, a drastic increase in salinity with more marine conditions (~25‰+) occurred at 9,100 cal. a B.P.

Measurements on modern *A. beccarii* indicated shell growth in early summer. The summer temperatures at the bottom of the shallow waters of the early phase of the Littorina Transgression (Unit D) were estimated at 23°C, with salinity being about 13 psu.

Comparable isotope signals in sediments having similar bedding characteristics indicate that the laminated and non-laminated Units D and E, respectively, are contemporaneous and regionally correlable from the Great Belt into the Kiel Bay.

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