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SURFACE WATER CHANGES IN THE NORWEGIAN SEA DURING LAST DEGLACIAL AND HOLOCENE TIMES

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Abstract — Stable carbon and oxygen isotopes of the polar planktic foraminifera *Neogloboquadrina pachyderma* sinistral from sediment cores of the Norwegian Sea reveal several anomalous ^{13}C and $\delta^{18}O$ depletions in the surface water during the last glacial to interglacial transition and during the later Holocene. The depletions that are observed between the Last Glacial Maximum (LGM) and the end of the main deglacial phase were caused by massive releases of freshwater from thawing icebergs, which consequently resulted in a stratification of the uppermost surface water layer and a non-equilibrium between the water below and the atmosphere. At ~8.5 ka (^{14}C BP) this strong iceberg melting activity ceased as defined by the cessation of the deposition of ice-rafted detritus. After this time, the dominant polar and subpolar planktic foraminiferal species rapidly increased in numbers. However, this post-deglacial evolution towards a modern-type oceanographic environment was interupted by a hitherto undescribed isotopic event (~7–8 ka) which, on a regional scale, is only identified in eastern Norwegian Sea surface water. This event may be associated with the final pulse of glacier meltwater release from Fennoscandia, which affected the onset of intensified coastal surface water circulation off Norway during a time of regional sea-level rise. All these data indicate that surface water changes are an integral part of deglacial processes in general. Yet, the youngest observed change noted around 3 ka gives evidence that such events with similar effects occur even during the later Holocene when from a climatic point of view relatively stable conditions prevailed.

INTRODUCTION

The physiography of the Norwegian Sea is subdivided by the Vøring Plateau and the Jan Mayen Fracture Zone into the Norway Basin to the south and the Lofoten Basin to the north (Fig. 1). On the western flank this area is bounded by the Jan Mayen Ridge and the Mohns Ridge, whereas the eastern side is defined by the Norwegian continental margin. The modern surface circulation pattern of the Norwegian Sea is dominated by the northward extension of the North Atlantic Drift, the Norwegian Current (NC), which has a thickness of 500–700 m and which carries water with salinities ranging between 35.1–35.3‰ (Swift, 1986). Towards its eastern boundary, the NC is partly superimposed by the Norwegian Coastal Current (NCC), in particular during summer (Swift and Aagaard, 1981). This current flows mainly along the Norwegian shelf. It is fed by water from the North Sea and the Norwegian coast, which lead to salinities <34.7‰ (Johannessen, 1986). Along its western side, the Atlantic water is separated from Arctic water by the distintive Arctic Front.

On its northward directed path into the Arctic Ocean, the Atlantic water cools whereby it gains density and sinks. This transporting mechanism is instrumental for the process of vertical overturn and formation of deep water in the Greenland and Iceland seas, water which is an important link of the global ocean circulation (Veum *et al.*, 1992). Subtle variations in this system triggered by, e.g. salinity fluctuations are postulated to account for some major northern hemisphere climatic changes in the glacial and deglacial past (e.g. Broecker and Denton, 1989). On the other hand, the Holocene is generally regarded a period of relative oceanic and climatic stability.

During recent years, intensive studies have been carried out on deep-sea sediments at higher latitudes in order to link past major environmental changes and short-termed climatic instabilities to oceanic processes (e.g. Broecker *et al.*, 1988). Much emphasis has been placed on the time period since the Last Glacial Maximum (LGM) by applying oxygen and carbon stable isotopes combined with AMS radiocarbon dating, and in particular on those processes involved during the last deglaciation (Termination I) that led to changes in circulation (Jones and Keigwin, 1988; Vogelsang, 1990; Lehman *et al.*, 1991; Weinelt *et al.*, 1991; Lehman and Keigwin, 1992; Sarnthein *et al.*, 1992; Veum *et al.*, 1992). Planktic foraminifera, their oxygen and carbon stable isotope composition as well as their faunal variability, have been

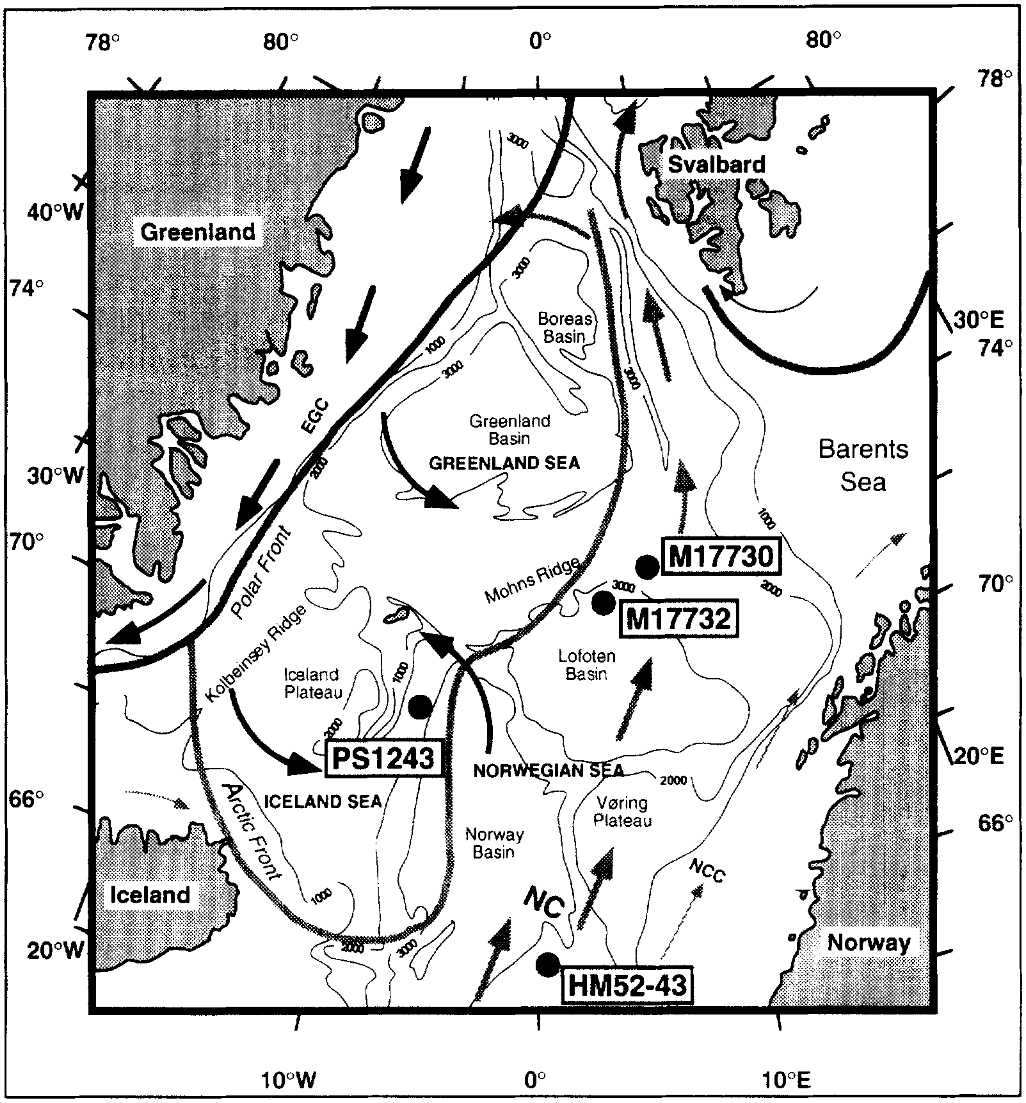

FIG. 1. Generalized overview of the surface water circulation and the distribution of the main water mass fronts in the Nordic Seas. EGC, East Greenland Current; NC, Norwegian Current; NCC, Norwegian Coastal Current; grey and black arrows denote warm and cold surface water, respectively.

serviceable tools to record past variations in surface water of the Nordic Seas, where oceanographic and climatic conditions are known to have changed rapidly during deglaciations. Oxygen isotope records from this area not only contain the glacial to interglacial shift in $\delta^{18}O$ caused by a varying global ice-volume, they are also affected by a strong regional overprint which is due to changes in temperatures as well as in salinities. Last glacial to Holocene sea surface temperature estimates indicate an increase of about 8–10°C for the Norwegian Sea (CLIMAP, 1981; Koç-Karpuz and Jansen, 1992). Variations in salinities in this region are mainly caused by the input of freshwater from melting icebergs during times of deglaciations such as Termination I (Sarnthein *et al.*, 1995).

The major aim of this study is to examine a series of stable isotope changes in the surface water during the last deglacial and the Holocene time period. This will be combined with faunal data and some important sedimentological observations.

MATERIAL AND METHODS

We investigated three cores from the slope of the Norwegian continental margin, all of which today underly the strong influence of inflowing Atlantic water (Fig. 1, Table 1). The southernmost core HM 43–52 is taken from the literature (Veum *et al.*, 1992). This core is ideally situated at today's main entrance area of inflowing Atlantic surface water into the Nordic Seas, whereas cores M17730 and M17732 from the Lofoten Basin are located directly along its northern flow path. For comparative purposes with these eastern cores, we also selected a core from the NW Norwegian Basin. This box core, PS1243, is located along the NE edge of the Iceland Sea close to the Arctic Front.

Stable isotope measurements were performed on the polar species *N. pachyderma* sin. using a fully automated MAT 251 mass spectrometer (^{14}C Laboratory at Kiel University). To reduce the effects of stable isotope differences among morphotypes and different size classes

TABLE 1. List of investigated cores

Site	Geographical position		Water depth(m)
	Lat. (N)	Long.	
M17730	72°06.7′	07°23.3′E	2749
M17732	71°36.8′	04°12.8′E	3103
PS1243	69°22.3′	06°32.1′W	2710
HM52-43	64°31.0′	00°44.0′E	2781

of this common species (Healy-Williams, 1992; Donner and Wefer, 1994), 25–30 four-chambered, quadrate specimens of similar size from the 125–250 μm fraction were picked.

For faunal quantification, the >125 μm fraction was used. This is a crucial issue, because the most abundant subpolar planktic foraminiferal species in the Nordic Seas, the relatively small-sized *Turborotalita quinqueloba*, bears important paleoceanographic implications for inflowing Atlantic water (Bauch, 1994; Hebbeln *et al.*, 1994). Since this species contains symbionts, its life habitat is restricted to the photic zone of the upper surface water, i.e. approximately 50 m. In contrast, *N. pachyderma* sin. is known to show large differences in depth migration during its life cycle, ranging from below 100 m to several hundred of meters of depth for open water conditions (Hemleben *et al.*, 1989). It therefore reflects deeper water below the thermocline rather than the actual surface. But depending on the specific conditions, the upper depth limit in particular may be lower. In ice-covered regions of the southern Arctic Ocean, *N. pachyderma* sin. seems to be related to a shallower habitat (Carstens and Wefer, 1992).

Radiocarbon measurements (AMS ^{14}C) were carried out at the laboratories of Gif (France) and of Aarhus University (Denmark). All datings are reservoir corrected by 400 years. For further stratigraphical purposes and to gain a control on ice-sheet fluctuations and associated meltwater releases, we noted the point of the cessation of ice-rafted rock fragments (IRD) >125 μm in the cores, which in the case of the Norwegian Sea should be coeval with the termination of freshwater discharges from melting iceberg.

RESULTS

Stratigraphy and Sedimentation Rates

The stratigraphical framework of cores M17730 and M17732 was established by considering the oxygen and carbon stable isotope record guided by a frame of AMS radiocarbon datings. The oxygen and carbon stable isotope records of both cores show in many respects a very good visual agreement for the final stage of the last glacial, Termination I, and the Holocene (Fig. 2). Aided by nine AMS ^{14}C dates from core M17730 (Weinelt, 1993) this allowed us to establish a detailed age frame for core M17732 also, despite the fact that there are some differences, e.g. the relation between the end of IRD deposition and our final isotope/age correlations.

The average sedimentation rate of both cores is about 3 cm ka^{-1} for the past 140 ka, but is considerably higher for the postglacial to Holocene sections with >5 cm ka^{-1} (Weinelt, 1993). Particularly for the time covering the past 40 ka, stable isotope samples were taken at closer intervals (each 2.5 cm). The sites of cores M17730 and M17732 are located in the distal reaches of the Bear Island sedimentary fan (Fig. 1). The Quaternary history of this fan explains the relative high sedimentation rates (>10 cm ka^{-1}) found in cores from this region for the past 15 ka (Laberg and Vorren, 1995). These rates presumably result from an increased input of IRD during glacial and deglacial times and a continuous downslope transport of finer sediments (<63 μm) by winnowing from neighbouring topographic highs, e.g. Barents shelf and Vøring Plateau, during interglacials (Bauch, H.A. *unpublished data*).

The stratigraphy of core PS1243 is based on three AMS radiocarbon measurements (Fig. 3). Within the lower part of this core there is a prominent ash layer made up of transparent shards which is widely dispersed on the Iceland Plateau area and which has been interpreted elsewhere by geochemical means to represent the Vedde Ash (Sejrup *et al.*, 1989; Kvamme *et al.*, 1989; Lackschewitz, 1991). The tephra layer of this eruptive event has been dated at 10 600 BP (Mangerud *et al.*, 1984) and since this coincides with the Younger Dryas cold spell, it is a useful stratigraphical and climatic datum (Veum *et al.*, 1992; Koç and Jansen, 1994). From the position of the Vedde Ash it can be inferred that the base of box core PS1243 is not older than about 11.5 ka. This assumption gains support from previous investigations in the SE Norwegian Basin where a spike of lowered $\delta^{18}O$ at a similar stratigraphical level has an interpolated age of about 11 ka (Veum *et al.*, 1992).

Core PS1243 is situated on the western side of the Norwegian Basin, at the flank of the Jan Mayen Ridge. Due to its distal position from continental margins, this site is not so likely to receive as much terrigenous material as those cores which are closer to the Scandinavian shelf areas. Therefore, this core has a lower sedimentation rate than the other two studied cores from the Lofoten Basin. It averages a sedimentation rate of about 4.5 cm ka^{-1} for the past approximate 12 ka and was sampled in 2 cm intervals.

The Record from the Lofoten Basin

Oxygen isotope distributions of both cores in Fig. 2 exhibit a typical development after the LGM. This is marked by a low event of ≥2‰ right after the heaviest oxygen values of Isotope Stage 2. Particularly in core M17732, this early spike is synchronous with a major depletion in $\delta^{13}C$, a feature also seen in core M17730 but not quite as strong. The further evolution of Termination I is characterized by a general tendency towards decreasing ^{18}O. This general decrease is accompanied by typical step-like variations, which can be linked to variations in global temperatures during times of diminishing ice volume and sea-level rise. A first steady level of the

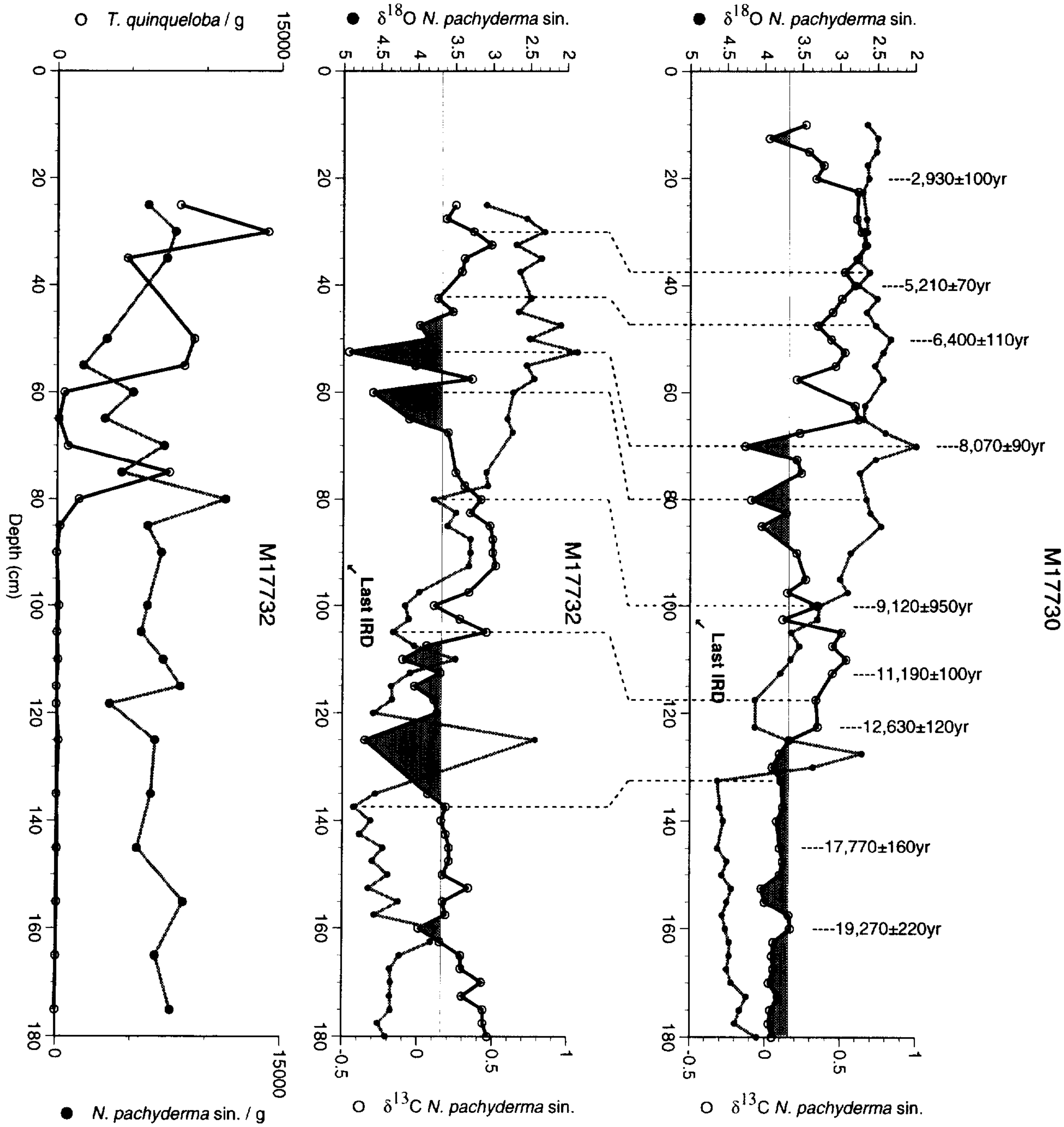

FIG. 2. Stable isotope records from the Lofoten Basin and correlation scheme of AMS-radiocarbon dated core M17730 with core M17732.

oxygen isotopes of about 3–2.5‰ is recorded between 9–8 ka (around 80 and 60 cm in both cores, respectively).

The decreasing oxygen values after the last glacial run parallel with an increase in $\delta^{13}C$. The latter reached a preliminiary peak towards the last occurrence of IRD. After that the $\delta^{13}C$ records of both cores show a return to steeply decreasing values. This depletion, which appears to be subdivided, culminates in a broad trough with the most negative values at its end. As a very significant feature, this strong depletion in ^{13}C is also coincident with the lowest oxygen values of the entire Holocene. A similar carbon to oxygen isotope relation is seen at the onset of Termination I. Above this major depletion, the carbon isotopes regain positive values of about 0.6–0.7‰.

The younger parts of the Holocene were not recovered in core M17732 but are present in core M17730. Here, oxygen isotopes show rather constant values between 2.5–2.7‰, whereas the $\delta^{13}C$ values decrease to almost zero.

The planktic foraminiferal results in core M17732 show that the polar species *N. pachyderma* sin. is present throughout the last glacial and the deglaciation, whereas *T. quinqueloba* indicates a significant peak in abundances at about 8.5 ka slightly lagging a maximum of *N. pachyderma* sin. (Figs 2 and 3). The highest test numbers of *T. quinqueloba* are noted at a depth of 30 cm (~5 ka).

Essentially, the two cores from the Lofoten Basin exhibit several events which are major depletions of $\delta^{13}C$

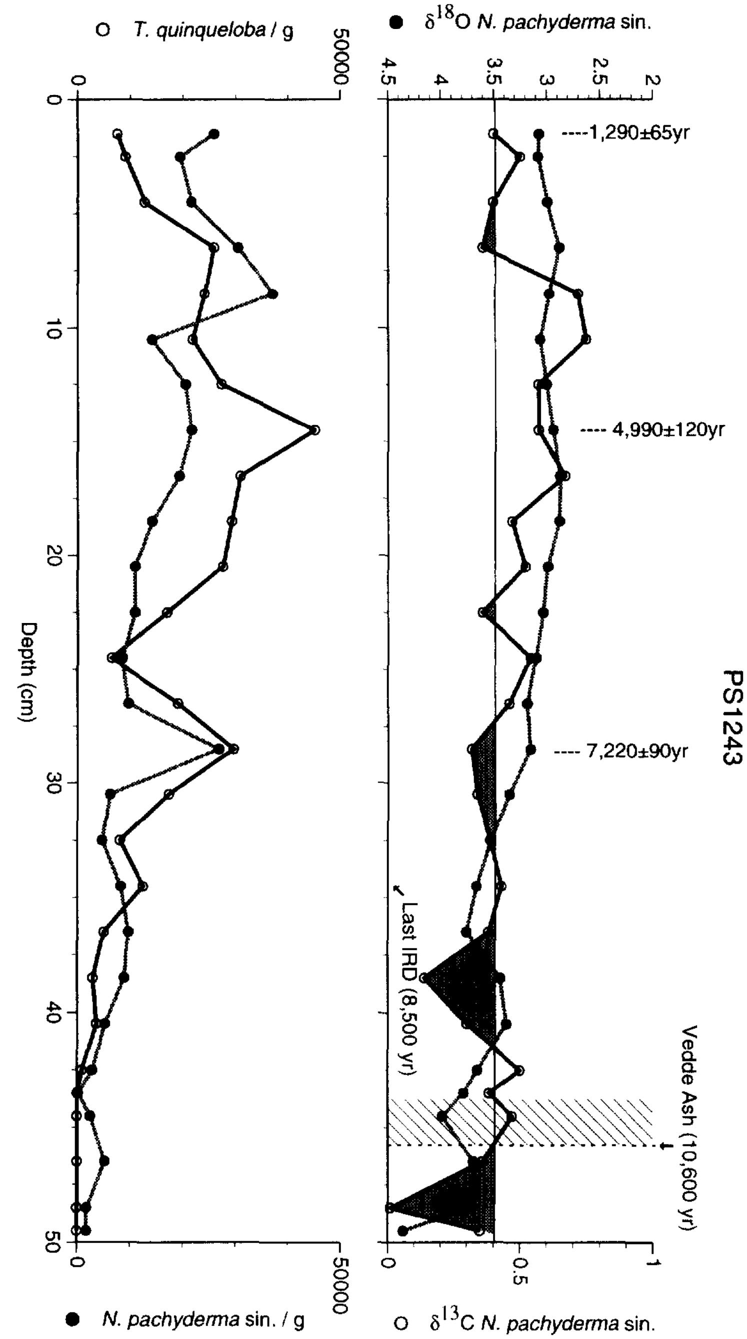

FIG. 3. Comparison of stable isotopes and faunal records from the western Norwegian Basin. The Vedde Ash represents a 2 cm homogenous layer of rhyolitic glass shards with a discrete base (stippled). Note the indicated higher level in $\delta^{13}C$ as compared with Fig. 2.

at the surface. It is noteworthy that the event around 8 ka occurred before the time when *T. quinqueloba* reached maximum abundances, but also well after the final deposition of IRD, which can be placed between 10–9 ka, i.e. after the Younger Dryas cold spell.

The Record of the Norwegian Basin

The earliest $\delta^{18}O$ anomaly in core PS1243 is recorded at its base and is synchronous with slightly negative values in $\delta^{13}C$. A second, similar anomaly occurs above the Vedde Ash around 30 cm. After that, both stable isotope records show the characteristic trend towards a postglacial situation although with some variations. Whereas the upper Holocene oxygen values remain rather constant between 2.8–3.1‰, highest $\delta^{13}C$ values of about 0.8‰ appear at 10 cm (interpolated age between 4–3 ka), which markedly decrease after that time. This tendency was already noted in core M17730.

The subpolar and polar planktic record of *T. quinqueloba* and *N. pachyderma* sin., respectively, show a rather similar abundance pattern during lateglacial and Holocene times, suggesting that both species rely on similar ecological parameters. The subpolar faunal record contains virtually no specimens below and during the Younger Dryas. But above the Vedde Ash and clearly after the last observation of IRD at about 8.5 ka (interpolated age), a rapid increase in abundance is noted which reached highest test concentrations near 5 ka as in core M17732. Beyond that time and until now, the total abundance of *T. quinqueloba* decreased by about 85%.

In comparison to the cores from the Lofoten Basin, core PS1243 does not reveal $\delta^{13}C$ depletions on a similar scale after 9 ka although there are some minor indications visible between 9–6 ka. Also the faunal results of both regions agree reasonably well if one considers that in core PS1243 both shown species again have a preliminary spike well above the last observed IRD, and that *T. quinqueloba* gained highest abundances at about 5 ka.

DISCUSSION AND CONCLUSIONS

Planktic carbon isotope records and their interpretations are much debated. This is because of several factors which complicate the interpretation. These are caused by species-dependent 'vital effects' and depth-habitat, as well as global sea water $\delta^{13}C$ variations of total CO_2 due to gas exchange with the atmosphere in combination with local primary productivity in the surface water (e.g. Berger *et al.*, 1981; Labeyrie and Duplessy, 1985). Regardless of these obstacles, when comparing oxygen and carbon stable isotope time series from high latitude sediment records there appears to be a good agreement between both proxies, e.g. with regard to glacial-interglacial water mass evolutions (Jansen, 1989). Somehow, as indicated by our data, oxygen and carbon stable isotopes combined appear to be particularly valuable for detecting meltwater events. The $\delta^{18}O$ value of the water is directly linked to the freshwater component and temperature. Indirectly ^{13}C may be also affected in that a pronounced salinity lowering at the surface would lead to a well stratified halocline, little vertical mixing with the water below (Aagaard *et al.*, 1981) and thus, an enhanced non-equilibrium state between the latter and the atmosphere is then most likely. This situation is even reinforced when sea ice cover and cooler temperatures are involved in these processes (Johannessen *et al.*, 1994). Recent stable isotope studies on *N. pachyderma* sin. from surface samples across the Laptev Sea continental slope (Siberia), which today receives the highest amount of riverine outflow around the entire Arctic, reveal low $\delta^{13}C$ values towards the direction of increasing influence of freshwater, i.e. lowered salinity in the surface layer (Spielhagen and Erlenkeuser, 1994). Furthermore, bottom water samples from below the low salinity layer of the shallow Laptev Sea shelf (on average $\leq$50 m) indicate $\delta^{13}C$ values which are far from isotopic equilibrium with atmospheric CO_2 (Erlenkeuser, 1995). Freshwater from melting icebergs should reflect $\delta^{13}C$ values of the atmosphere (today between −7 to −8‰). Values measured on air bubbles trapped in ice cores indicate a slight increase in atmospheric ^{13}C by not more than 0.5‰ since the LGM (Leuenberger *et al.*, 1992). Therefore, freshwater itself may contribute significantly to a lowering of the $\delta^{13}C$ values in planktic foraminifera, once the particular depth habitat of the relevant species becomes affected. Major surface water anomalies of the past, which were clearly related to iceberg melting, reveal low $\delta^{18}O$ and $\delta^{13}C$ values in planktic foraminiferal tests (Veum *et al.*, 1992). They further indicate that ventilation processes within the upper hundreds of meters were reduced due to overlying low salinity surface water. It is believed that these meltwater lids can reduce the intensity of the so-called 'conveyor belt' and hence, are of important climatic significance (Broecker and Denton, 1989; Bond *et al.*, 1993; Maslin *et al.*, 1995).

Series of Events

The various records presented here from the Norwegian Sea give evidence of major surface water changes during the past 15 ka. Because of its proximity to the ice-covered Norwegian continent during glacial times, these changes must be primarily viewed under the circumstances of a variable Fennoscandian ice sheet. In Fig. 4 such fluctuations are documented in the isotope records of the last 30 ka. It becomes apparent that typically negatively correlated events of $\delta^{13}C$ and $\delta^{18}O$ are not solely restricted to the past 15 ka. In particular, a significantly older event occurred at about 26 ka, which may be attributed to Heinrich Layer 3 as one event of a series that were induced by oscillating instabilities of the northern ice sheets (Bond *et al.*, 1993; Grousset *et al.*, 1993; Andrews *et al.*, 1994; Fronval *et al.*, 1995). But most of these phenomena, which are clearly related to enhanced discharges of icebergs and consequent meltwater release into the open North Atlantic, happened during times of relatively intermediate insolation and glacial conditions. From a climatic point of view they therefore should be separated from those types of events described here, which are inherent to an irreversible deglaciation (i.e. Termination I) as well as the ensuing full Holocene interglacial.

A comparison of our data from Termination I and the Holocene reveals three major events (events 1–3) that can be traced along a N–S transect from the Lofoten Basin to the southern Norwegian Basin (Fig. 4). The first event correlates to early deglacial processes, most likely those of the Barent Sea ice sheet during the first 2 ka of Termination I (Jones and Keigwin, 1988; Sarnthein *et al.*, 1992; Forman *et al.*, 1995). All cores show some notable irregularities between 13 to 11 ka, which may be due to a prolonged time of meltwater discharges (MWP IA of Fairbanks, 1989), but eventually culminate in high $\delta^{13}C$ values between 11–10 ka, i.e. the Younger Dryas cold period (YD).

Event 2 appears to be tripartate: the first two parts (2a, 2b) occurred around 10–9 ka and 9–8 ka, respectively. They are associated with the renewed meltwater release after the YD and are equal to MWP IB (Fairbanks, 1989). The near-synchronous disappearance of IRD in our cores

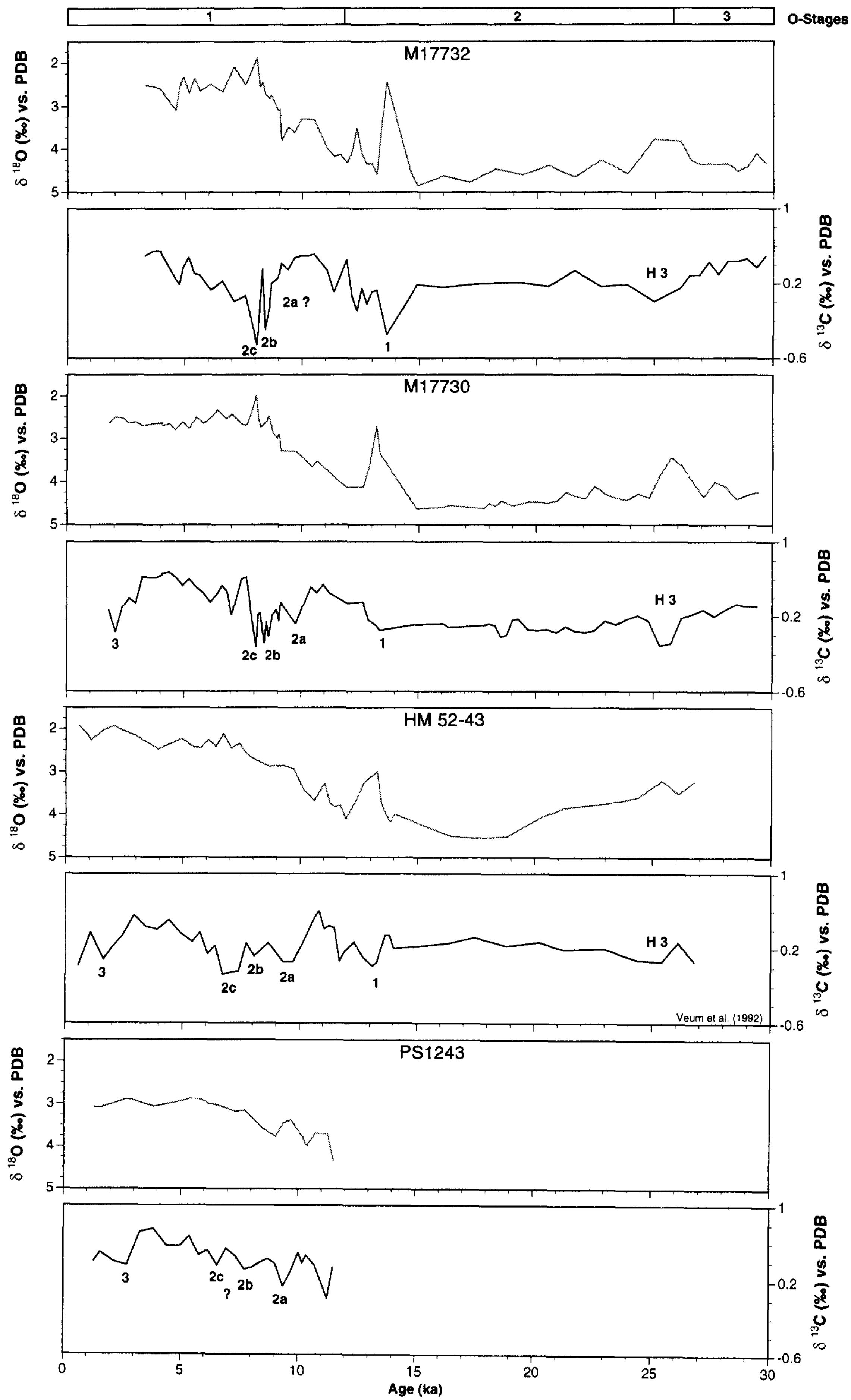

FIG. 4. A comparison of stable isotope records from the Norwegian Sea and position of discussed events.

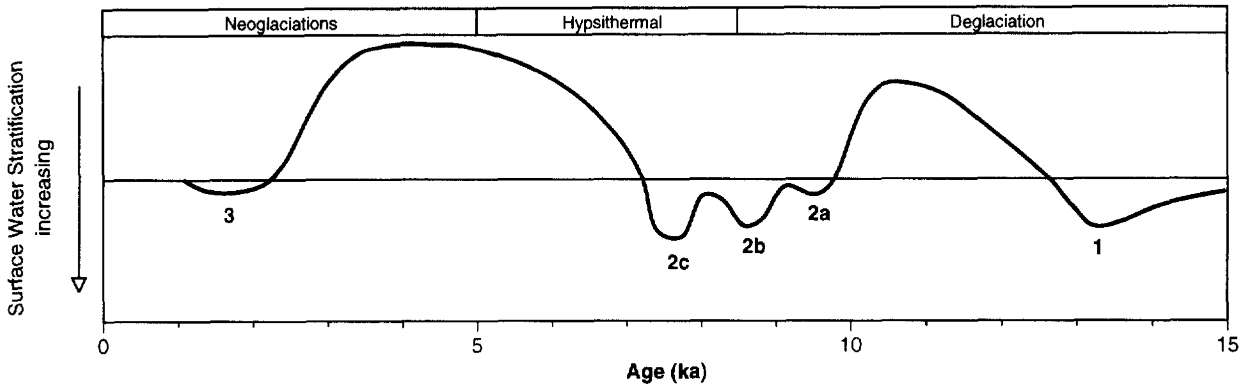

FIG. 5. Schematic drawing of observed surface water instabilities in the Norwegian Sea since the LGM.

is most likely related to the end of drifting icebergs in the Norwegian Sea. This is corroborated by other data from Norway which suggest at least a halt or even readvance of glaciers during the Younger Dryas and a full retreat of the glaciers into the inner fjord systems some time after this cold period (e.g. Andersen, 1980). Interestingly, the YD is not clearly recognizable within the $\delta^{18}O$ record of our most northern cores, but noted as a well defined trough of increasing $\delta^{18}O$ throughout the Vedde Ash layer of core PS1243. This is possibly due to a much smaller temperature gradient in the surface water of the Lofoten Basin before and during the YD.

Event 2c is observed between the end of iceberg melting and before full interglacial conditions are reached as indicated by *T. quinqueloba* peak abundances. The age of Event 2c ranges between 7 ka in core HM 52-43 and 8 ka in core M17730. The differences in age determination greatly depends upon the different depth levels of the datings in each core. But this event in *N. pachyderma* sin. $\delta^{13}C$ records can be traced also in other cores from the south-eastern Norwegian continental slope (~7.5 ka in Koç-Karpuz and Jansen, 1992). It falls within the onset of the Hypsithermal when highest temperatures were recorded for this region (Nesje and Dahl, 1993).

The youngest observed depletion in planktic $\delta^{13}C$ at about 3 ka (Event 3) shows up in all our studied cores and coincides with a time of steady decrease of *T. quinqueloba* abundance. This decrease started after 5 ka (onset of the Neoglaciation) indicating a significant change within the surface water of particularly the western Norwegian Sea and Iceland Sea, which may have to do with a drop in sea surface temperatures due to a general climate deterioration (Koç-Karpuz and Jansen, 1992) and/or an increase in surface water ventilation (Bauch *et al.*, in preparation). A similar cooling trend after 5 ka is noted in records from the East Greenland shelf for the same time (Williams *et al.*, 1995) and is also well recorded in coastal Greenland and Canadian Arctic ice cores (Paterson *et al.*, 1977; Johnsen and Dansgaard, 1992). But as Event 3 occurred well after the main deglaciation, it together with Event 2c may not be so easily linked to the series of our postulated deglacial events. Both most recent depletions in $\delta^{13}C$ therefore suggest that mechanisms other than melting icebergs can cause severe surface water changes.

Climatic Implication

As evidenced by our above findings the surface water instabilities of events 1 and 2a-b can be directly linked to the final melting of the northern hemisphere ice sheets, which commenced after 15 ka (Fig. 5). Synchronous with deglaciation, a major variability in atmospheric and oceanic dynamics started to develop, as observed in both, marine and terrestrial high latitude records (Lehman and Keigwin, 1992; Taylor *et al.*, 1993). These caused the well known two-stepped features in marine $\delta^{18}O$ records, i.e. Termination Ia and Ib. Since IRD deposition in the Nordic Seas ceased at about 8.5 ka, a connection of Event 2c at 7–8 ka with iceberg release from the Greenland and Fennoscandian ice sheets seems improbable. Based on many cores from the low latitude Atlantic, Mix and Ruddiman (1985) recognized a three-stepped Termination I, introducing the term Ic for the time period between 8–6 ka. More recently, a rapid increase in sea level of 6.5 m drowning Caribbean coral reefs at ~7 ka (7.6 ka calendar years) has been reported (Blanchon and Shaw, 1995). According to Blanchon and Shaw (1995), this catastrophic rise event (CRE 3) has no obvious link to northern hemisphere ice sheets either and may be caused by a marine ice sheet instability around the Antarctic. Interestingly, the two central Greenland ice cores GISP II and GRIP show a distinct cooling for this time period (Grootes *et al.*, 1993).

Because of the apparent synchroneity of CRE 3 with our Event 2c and Termination Ic, it may be tempting to hypothesize a global event as direct cause. But, it should be considered that also the northern European region underwent major changes between 8–7 ka, which possibly strongly influenced Norwegian Sea circulation. A key point could be, therefore, the initiation of the coastal circulation which began during the flooding of the North Sea at ~8 ka and the ensuing invasion of the Baltic basin by the Litorina Sea (e.g. Stabell and Thiede, 1985). This process should have led to an increased lowering of the salinity in the surface water of the northern North Sea and together with the near synchronous opening of the English Channel enhanced North Sea circulation (Stabell and Thiede, 1985), forcing this low salinity water to flow northward. This assumption gains support from a strong ^{18}O depletion at ~8 ka in the TROLL 3.1 core (Lehman and Keigwin, 1992), which is ideally situated just south of Norway within the Norwegian Trench to monitor such an event. As a further consequence of the water mass evolution in the North Sea-Baltic Sea region, the Norwegian Coastal Current developed (Bjørklund *et al*., 1985).

Based on all this evidence, we believe that our Event 2c is most likely linked to these regional processes, rather than the entrainment of meltwater in the North Atlantic from a distal Laurentide ice sheet as described from similar earlier situations (Andrews *et al*., 1994). The final stage of the melting of remnant Norwegian ice caps occurred during the onset of the Hypsithermal after ~8 ka, when according to Nesje and Dahl (1993) todays' Mid-Norwegian glaciers started to disappear totally (Fig. 5). The Norwegian coastal surface circulation was then much more prone to receive larger amounts of low salinity surface water from the fjords and North Sea area than possibly later. This could have led to a westerly expansion of the NCC across the eastern part of the NC, which in turn may have resulted in a suppression of the intensity of the latter. This also may be the reason why Event 2c is not clearly documented within the planktic carbon isotopes of the western Norwegian Basin. But still there seems to have been an impact on the subpolar and polar planktic foraminiferal fauna at about the same time (Fig. 2).

Our correlation between late glacial and early Holocene surface water changes and the Fennoscandian ice sheet history does not allow us to speculate on the cause of the most youngest observed carbon isotope depletion, Event 3. Although a significant feature in all our presented Norwegian Sea cores, Event 3 is less obvious in other parts of the Nordic Seas (Bauch and Weinelt, *unpublished data*). It may, therefore, be connected only to ^{13}C variations of total dissolved CO_2 within the surface water of the Norwegian Current itself due to changes in the degree of surface water stratification. An increase in bioproductivity as cause seems to be unlikely, because the nutrient content at todays well pertubated Arctic Front is as high as in the warmer Atlantic layer where the $\delta^{13}C$ values of the surface water and in *N. pachyderma* sin. are comparatively much more depleted (Johannessen *et al*., 1994). A more defined stratification may have been either induced by the process of widespread freshening of the upper surface water of the northern North Atlantic (Dickson *et al*., 1988) or by higher surface temperatures of the North Atlantic layer. At least a freshening at about 3 ka can be also recognized in paleosalinity estimates from the NE Atlantic (Maslin *et al*., 1995).

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