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A paleoclimatic evaluation of marine oxygen isotope stage 11 in the high-northern Atlantic (Nordic seas)

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Abstract

A sediment core from the high latitude of the Northern Atlantic (Nordic seas) was intensively studied by means of biogeochemical, sedimentological, and micropaleontological methods. The proxy records of interglacial marine oxygen isotope stage (MIS) 11 are directly compared with records from the Holocene (MIS 1), revealing that many features of MIS 11 are rather atypical for an interglaciation at these latitudes.

Full-interglacial conditions without deposition of ice-rafted debris existed in MIS 11 for about 10 kyr (~ 398–408 ka). This time is marked by the lightest d18O values in benthic foraminifera, indicating a small global ice volume, and by the appearance of subpolar planktic foraminifera, indicating a northward advection of Atlantic surface water. A comparison with MIS 1, using the same proxies, implies that surface temperatures were lower and global ice volume was larger during MIS 11. A comparative study of the ratio between planktic and benthic foraminifera also reveals strong differences among the two intervals. These data imply that the coupling between surface and bottom bioproductivity, i.e., the vertical transportation of the amount of fresh organic matter, was different in MIS 11. This is corroborated by a benthic fauna in MIS 11, which contains no epifaunally-living species. Despite comparable values in carbonate content (%), reflectance analyses of the total sediment (greylevel) show much higher values for MIS 11 than for MIS 1. These high values are attributed to increased corrosion of foraminiferal tests, directly affecting the sediment greylevel. The reason for this enhanced carbonate corrosion in MIS 11 remains speculative, but may be linked to the global carbon cycle.

Keywords: interglacial oxygen isotope stage 11; Northern Atlantic; paleoclimate

1. Introduction

Studies of past interglacial periods may provide an insight into climate modes, which may be relevant to the near future. Investigating the temporal and the spatial changes of marine water masses at high northern latitudes is important to learn more about the complex forcing and feedback mechanisms which drive the earth's climate system today as well as in the recent past, i.e., during glacial–interglacial cycles. The Northern Hemisphere landmasses became repeatedly glaciated and deglaciated during these glacial–interglacial cycles. This waxing and waning in global ice volume, as recorded in oxygen isotope

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records from deep-sea sediments, has been linked to variations in solar radiation (Hays et al., 1976; Imbrie et al., 1984). During the past 600 ka, these orbitally induced variations in insolation show a strong 100 kyr cyclicity (Imbrie et al., 1993; Tiedemann et al., 1994). In the North Atlantic these periodicities were expressed as relatively long intervals having cold sea surface temperatures (SSTs), which were interrupted by brief, much warmer episodes (Ruddiman et al., 1986). During the past 450 ka, at least five warm intervals with low global ice volume and relatively warm SSTs are recognized in sediment records from the North Atlantic and attributed to marine oxygen isotope stages (MIS) 1, 5, 7, 9, and 11.

Today, and also probably during most of the Holocene period (MIS 1), the relatively warm and salty North Atlantic surface water extends far into subpolar and polar latitudes of the Norwegian, Greenland, and Iceland seas (Nordic seas, Fig. 1). The ''effectiveness'' of the thermohaline circulation, that is the northwardly advection and vertical overturn of these surface water masses to form denser deep water, is apparently dependent on the conditions in high latitude surface waters (Broecker and Denton, 1989; Imbrie et al., 1993; Rahmstorf, 1995). Previous micropaleontological investigations, which were carried out on different calcareous fossil groups using longer time series records from the Nordic seas, have shown significant glacial to interglacial contrasts in all of the species assemblages studied (e.g., Kellogg, 1977; Streeter et al., 1982; Bleil and Gard, 1989; Henrich and Baumann, 1994). Recently, it has been shown that interglacial periods can differ significantly from each other (Struck, 1997; Bauch, 1997). These studies further indicated that the records of MIS 11 (362–423 ka) are unusual, revealing many features not known from any of the later interglacial periods, i.e., MIS 1 and 5e.

Using a range of different proxy data, this study investigates MIS 11 in more detail. This period has often been referred to as a particularly intense interglacial (Ruddiman and McIntyre, 1976; Crowley, 1991). Because the intensity record of solar radiation

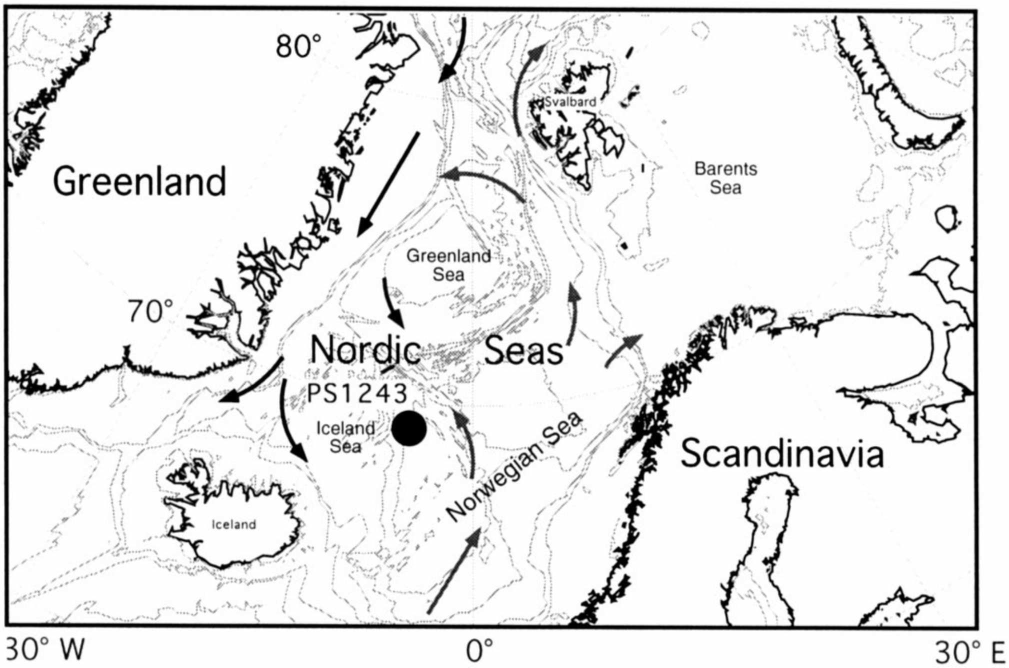

Fig. 1. Generalized surface water circulation in the Nordic seas. Black arrows denote water inflow from the Arctic Ocean whereas grey arrows repesent the flow pattern of warmer water of Atlantic origin. The location of the investigated core PS1243 is indicated (note that in some previous publications this core is also listed as 23243).

of MIS 11 has strong similarities to the Holocene, a study of MIS 11 may offer climatic information which may help predicting the climate of the near future (Howard, 1997a). In order to evaluate the conditions during MIS 11, we compare the records from MIS 11 with data from the much better known Holocene period.

2. Material

Core PS1243 from the western part of the southern Norwegian Sea was selected for this study. This gravity core together with a trigger box core was obtained in 1984 from a water depth of 2710 m (Augstein et al., 1984). Today, this site is located at the western edge of the warm Norwegian current near the Arctic Front (Swift, 1986) where the SST today is about 7°C and the bottom water temperature is around −1°C (Environmental Working Group, 1998). Owing to its position at the eastern slope of the Iceland Plateau, PS1243 shows less variable sedimentation rates than are found in cores further east, along the Norwegian continental margin (Henrich et al., 1989). Previous investigations have revealed that PS1243 extends back to MIS 12 (Birgisdottir, 1991; Bauch, 1997). The average sedimentation rate of the entire core is ≤ 2 cm/kyr. In some intervals rates can be much higher. During peak interglacials, such as the past 9 ka, when sediments are mainly composed of biogenic carbonate, sedimentation rates in this core are about 4 cm/kyr (Bauch et al., 1996).

3. Methods

A series of different measurements were made on PS1243, which can be used as paleoceanographic proxies. The benthic and planktic foraminiferal assemblages were studied (mesh size > 125 μm) in order to gain information on paleoproductivity and past water mass changes. Another similar approach was carried out by measuring the calcium carbonate content of the bulk sediment, assuming that the carbonate content (% weight) mainly reflects the calcitic microfaunal groups of foraminifera and coccolithophorids.

Color reflectance measurements of glacial–interglacial sediments from the Northern Atlantic have shown good correlation with carbonate contents (Grousset et al., 1993; Cortijo et al., 1995). Apparently, this correlation depends upon a relative increase in the production of biogenic carbonate during warmer intervals (lighter color) and less productivity together with an increased amount of iceberg-rafted detritus (IRD) during colder intervals (darker color). In the Nordic seas, changes in sediment lightness mainly depend on the input of biogenic $CaCO_3$, the specific types of IRD, and some ash layers (Helmke, 1996). Reflectance measurements were carried out on PS1243 at discrete 1-cm steps using a hand-held Minolta CM-2002 Spectrophotometer. The results are expressed as relative greylevel values L (0–100%) in accordance with Nagao and Nakashima (1992).

The occurrence of IRD at specific depth levels in late Quaternary sediment of the Nordic seas gives evidence of the spatial and temporal distribution of icebergs during glacial and deglacial times (Baumann et al., 1995; Fronval et al., 1995). In order to take advantage of this important paleoclimatic indicator, IRD grains (> 250 μm) were counted in core PS1243 for MIS 1 and 11; a direct comparison of our results with data obtained previously from PS1243 using the size fraction 125–250 μm (Birgisdottir 1991) could not reveal a difference in core depth between the two size fractions with regard to the cessation and onset of IRD deposition across main interglaciation (i.e., MIS 1, 5e and 11).

4. Results

4.1. Downcore stratigraphy and stable isotopes

The age model of the younger core section is based on AMS radiocarbon analyses measured on the left-coiling variety of the polar planktic foraminiferal species *Neogloboquadrina pachyderma* sin. The derived radiocarbon ages were reservoir corrected and translated to calendar years (Bauch et al., 1996).

The further downcore stratigraphy of core PS1243 was constructed using the SPECMAP chronology (Imbrie et al., 1984) and is based on both planktic and benthic stable oxygen isotope records (Fig. 2). For the planktic isotope record the species *N. pachyderma* sin. was used. The benthic isotopic curve is a splice record of the epifaunal species *Cibicidoides wuellerstorfi* and the infaunal species *Oridorsalis umbonatus*. To produce a complete glacial–interglacial $\delta^{18}O$ record this splicing is necessary (Labeyrie et al., 1987; Bauch et al., 1996) because in the Nordic seas *C. wuellerstorfi* rarely occurs in glacial and deglacial sediments (Belanger, 1982; Struck, 1995, 1997). The oxygen isotope data of *C. wuellerstorfi* and *O. umbonatus* were corrected by + 0.64‰ and + 0.36‰, respectively, to account for their well established species-dependent departure from isotopic equilibrium (Duplessy et al., 1988). The correlation of the oxygen isotope record of MIS 11 in core PS1243 to SPECMAP isotope stratigraphy is illustrated in Fig. 3. All three records would indicate that the interval of smallest global ice volume, which presumably would also be the time of full-interglacial conditions, lasted from 398–408 ka.

Based on the two $\delta^{18}O$ records (Fig. 2a,b) a clear stratigraphic distinction between glacial and interglacial stages can be made. One of the striking features in these two downcore records is the good visual covariance of the spliced benthic record with the planktic record of *N. pachyderma* sin. Most $\delta^{18}O$ shifts appear nearly synchronous in both records. A main difference between these two isotope records is the much larger range in the absolute shift in the oxygen isotope amplitude of the planktic record across glacial–interglacial stage boundaries. This is probably due to the additional increase in SSTs during interglacials, which is superimposed on the absolute $\delta^{18}O$ change caused by the global ice-volume effect. Accordingly, the planktic $\delta^{18}O$ record clearly marks the stages 1, 5e, and 11 as the three warmest periods in PS1243. But in the benthic record, the $\delta^{18}O$ values of MIS 11 are heavier than in MIS 1, 5e, and even MIS 7, possibly indicating differences in global ice volume among the interglaciations.

4.2. Faunal and sedimentary records

The indication from the $\delta^{18}O$ records that MIS 1, 5e, and 11 were the three warmest interglacial periods of all past 12 oxygen isotope stages is supported by the carbonate content and the total concentrations of benthic and planktic foraminifera (Fig. 2 e,f,g). Both the carbonate and planktic formininfera data from MIS 11 show values comparable to MIS 1, indicating high carbonate productivity during the two periods. However, the faunal record of benthic foraminifera shows relatively low values during MIS 11. From comparison of the entire downcore pattern, it is suggested that the number of planktic foraminifera seems to have had the greatest influence on the carbonate values. We also note that especially during the MIS 11, maximum foraminiferal abundances slightly lead the highest values observed in the carbonate record.

Comparing the carbonate record with the sediment reflectance, the highest carbonate values observed for MIS 1 and 5e are not reflected by significant changes in sediment color. In fact, these changes do not differ from those observed during glacial periods when only little or no carbonate is found. This is in contrast to MIS 11, which shows a strong shift towards high reflectance and obviously a better visual correlation with the carbonate record, even though carbonate values in MIS 11 are no higher than in MIS 1.

A downcore record of the planktic foraminiferal $\delta^{13}C$ values reveals that in MIS 11 (Fig. 2c) values > 1.0‰ are noted. These values are significantly higher than those of the other two warmest interglaciations, MIS 1 and 5e.

5. MIS 1 versus MIS 11

One of the presumptions of the ice-age theory based on orbital cycles is that glacial to interglacial changes, i.e., major changes in northern hemisphere glaciation, are led by several thousand years by a maximum in insolation (Imbrie et al., 1993; see also Fig. 4b). In the Nordic seas, the detailed phase relationship between an interglacial proxy (e.g., the high proportion of subpolar planktic foraminifera relative to the polar species *N. pachyderma* sin.), indicating enhanced inflow of Atlantic surface water masses, and a glacial–deglacial parameter (e.g., IRD), which reflects the existence of icebergs, has been documented previously (Bauch et al., 1996). Records of these two important proxies in relation to

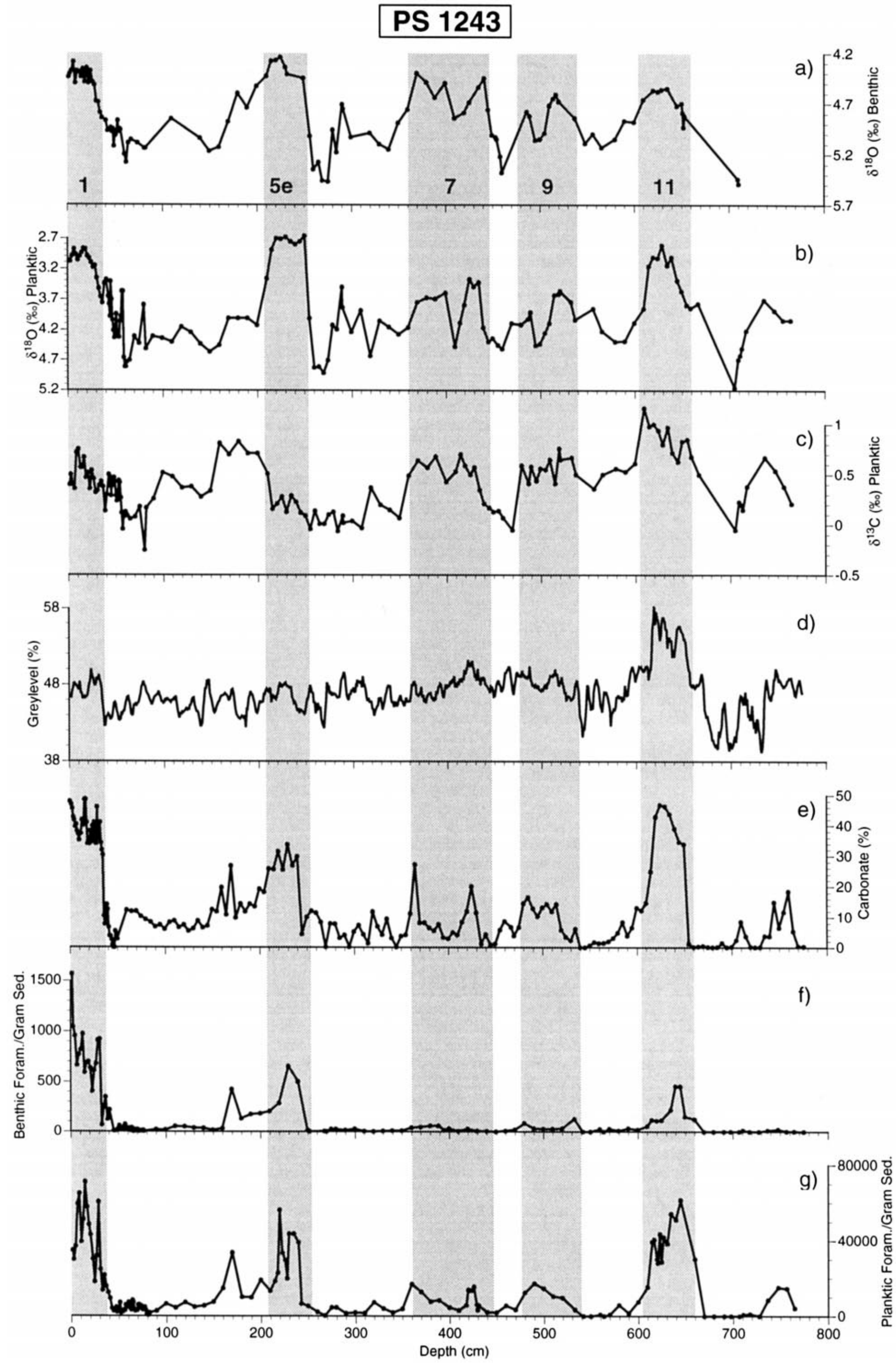

Fig. 2. Downcore records of the main proxy tools applied to PS1243 in this study. Some of these records show distinct differences between interglacial intervals (MIS) which are shaded in grey. Particularly obvious discrepancies between MIS 1 and 11 are observed in graphs c, d, and f whereas other proxies appear to be similar (graphs a, b, e, and g). Record (e) is taken from Birgisdottir (1991).

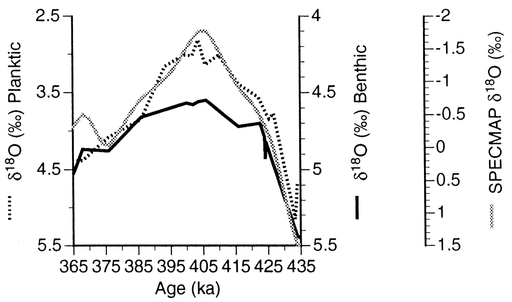

Fig. 3. Correlation of the benthic and planktic $\delta^{18}O$ records of core PS1243 to the SPECMAP isotope stratigraphy (Imbrie et al., 1984) for the time period of MIS 11.

the incoming solar radiation are shown in Fig. 4a. They strongly indicate that full-interglacial conditions associated with a significant decrease in *N. pachyderma* sin. always occurred after the deposition of IRD from melting icebergs had ceased, however, high amounts of IRD are observed during the late glacial phase prior to MIS 1 and 11. During the ensuing transitional phase, which was the time of highest insolation and of primary deglaciation and, thus, was associated with a major input of deglacial meltwater to the Nordic seas from a diminishing Fennoscandian ice sheet (Sarnthein et al., 1995), the deposition of IRD steeply decreased. In MIS 11, the primary deglaciation lasted from 430–410 ka. This is about three times longer than the deglacial period after the last glaciation. In this context it is interesting to note that the duration of the full-interglacial conditions in both stages appears similar, lasting about 10 kyr. The observation that the deglacial process stopped near 408 ka, as indicated by the IRD record, can be also deduced from continuously decreasing planktic and benthic oxygen isotope values up to this time (Fig. 4b,c). This timing agrees reasonably well with absolute dating of terrestrial isotope records (Winograd et al., 1997). With regard to total global ice volume decrease and associated sea-level rise, the consistently heavier benthic $\delta^{18}O$ values observed in the full-interglacial part of MIS 11 relative to MIS 1 would indicate both, a greater global ice volume and a lower sea-level stand during

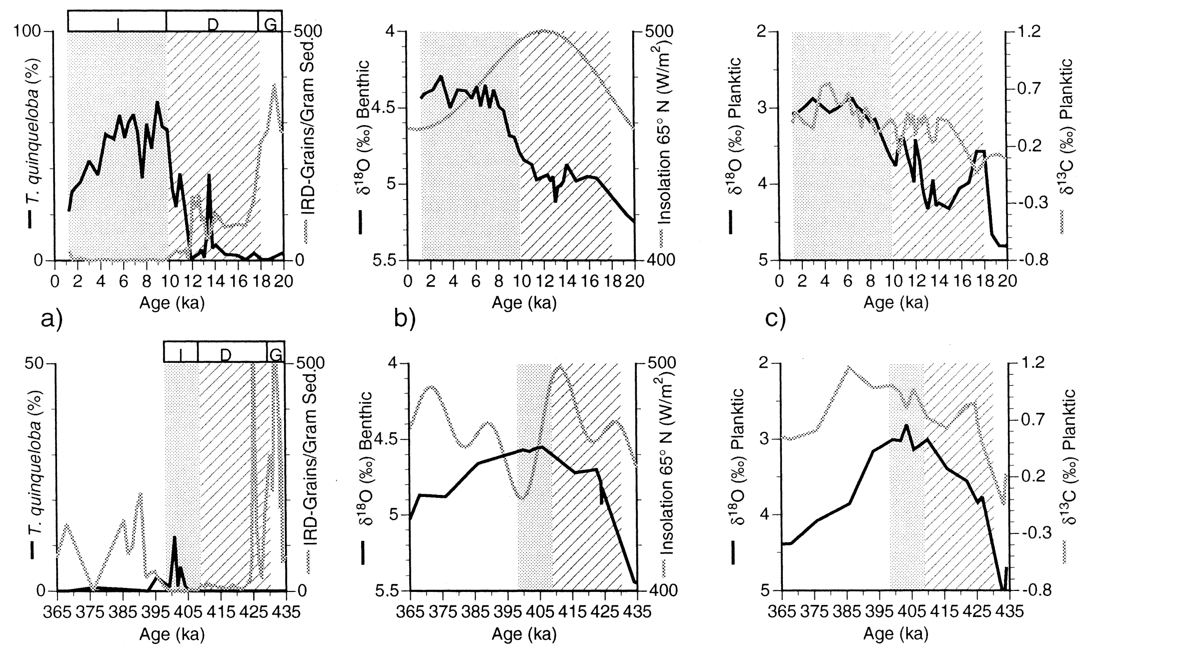

Fig. 4. Direct comparison of records for the past 20 ka and the time period of MIS 11. On the basis of IRD records depicted in 4 a, each studied interval was subdivided into a late glacial (G), a deglacial (D), and a full-interglacial (I) part. These defined sections are shown in all other graphs in this figure and Fig. 5.

the former interglaciation (provided that bottom water temperatures were similar).

5.1. Surface and bottom water conditions

The only small decrease in polar *N. pachyderma* sin. abundance in MIS 11 relative to MIS 1 (Fig. 4a) would suggest cooler near-surface water temperatures (Bé and Tolderlund, 1971) in MIS 11. In fact, taking only the ''true'' subpolar species (Fig. 5a), which in this case is *T. quinqueloba*, the total proportion of warm-water indicators would be only 10% compared with up to 70% during early to mid MIS 1 when SSTs appear to have been warmest (Koç et al., 1993; Bauch, 1997). The occurrence of about 5% of specimens of the right-coiling variety of *N. pachyderma* have no temperature-related significance, because such proportions are present in core PS1243 well before 408 ka and after 398 ka and appear to be ''normal'' for other glacial intervals in the Nordic seas and even for Holocene sediments from the Central Arctic Ocean (Bauch, 1997, 1999). About 75% of *N. pachyderma* sin. are observed in our youngest sample. Since the present day SST in the area of PS1243 is about 7° C, the maximum decrease of *N. pachyderma* sin. down to just 85% would infer a SST below 7°C for the warmest part of MIS 11. Even cooler SSTs must have prevailed until 408 ka, because it has been shown that an assemblage consisting of more than 95% *N. pachyderma* sin. is associated with SSTs well below 5°C (Bé and Tolderlund, 1971; Kellogg, 1980). In MIS 1 and 5e, the warmest period evolved directly after the period of deglaciation and highest insolation and seems to have expanded over a large area of the Nordic seas (Fronval et al., 1998; Bauch et al., 1999). For MIS 11 a similar oceanographic evolution cannot be inferred (Bauch, 1997).

The confinement of *T. quinqueloba* occurrence to the IRD-free interval (between 398–408 ka) is in strict agreement with the present day and Holocene situation (Johannessen et al., 1994; Bauch et al., 1996). However, when comparing the total number of planktic foraminifera (including *N. pachyderma* sin.) with that of the benthic foraminifera (Fig. 5b), there is a disagreement between MIS 1 and 11. During early MIS 1, planktic and benthic test concentrations are low, but both steadily increase after

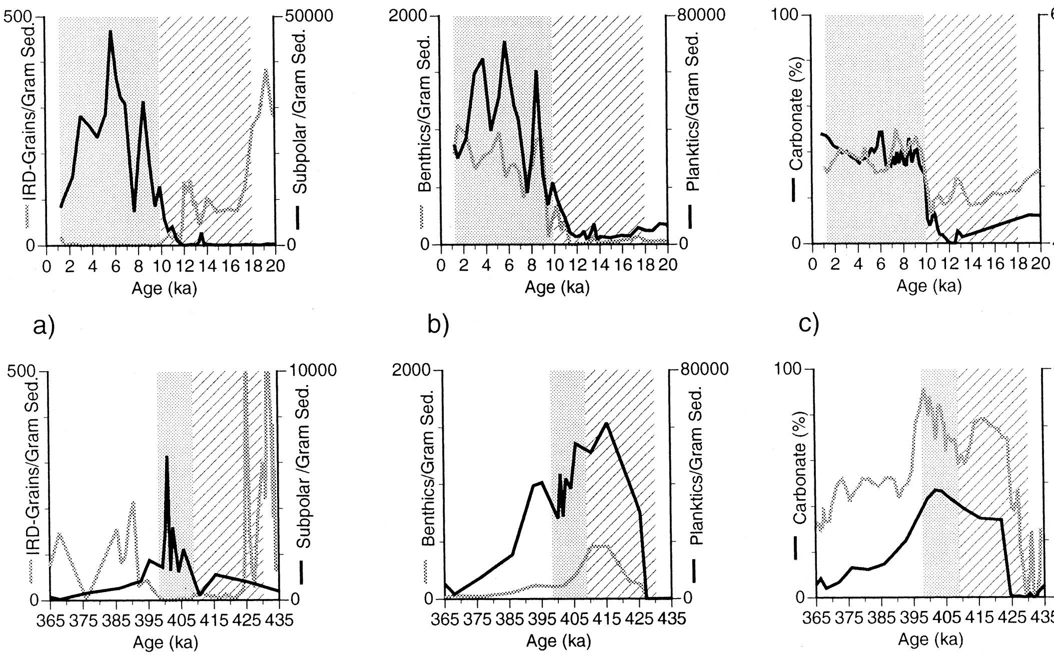

Fig. 5. Direct comparison of records for the past 20 ka and the time period of MIS 11 (see also Fig. 4).

the end of Termination I. This is in contrast to MIS 11, where both groups show highest abundances during the deglacial and early full-interglacial parts, and are clearly leading the main occurrence of subpolar abundance (see Fig. 5a). In general, planktic foraminiferal species can be related to changes in physical and chemical parameters such as surface temperature, salinity, and nutrients (Reynolds-Sautter and Thunell, 1989), whereas benthic foraminifera are more apt to reflect surface ocean primary productivity of a given water mass, i.e., the specific circumstances of how and in what quantities food is exported to the sea bottom (Gooday, 1988). The ratio between the two groups, therefore, may be used as productivity index (Berger and Diester-Haass, 1988). Comparing the planktic and benthic abundance records between the two investigated time intervals results in a significantly higher ratio for MIS 11 than for MIS 1. This may imply that although surface water bioproductivity was very high during MIS 11, there was not a proportional increase in total benthic specimens. This is important, because the productivity at the surface represents the main food source for the benthic community via downward transportation of organic matter (Graf, 1989). However, the entire situation in MIS 11 is even more complicated. As has been demonstrated in previous studies, certain epifaunal species such as *C. wuellerstorfi* can be regarded as typical indicators for interglacial periods in the Nordic seas (Streeter et al., 1982; Haake and Pflaumann, 1989). Moreover, the dominance of epifaunal species over infaunal species as a result of Holocene-like water-mass circulation, i.e., advection of warm Atlantic surface water and vertical convection, has been recently shown (Struck, 1995).

It is puzzling that almost the entire benthic foraminiferal assemblage in MIS 11 is made up of infaunal species, or species which feed on detrital organic material (Struck, 1997) rather than filtering their food supply from the water itself like *C. wuellerstorfi* (Lutze and Thiel, 1989; Linke and Lutze, 1993). The very few specimens of *C. wuellerstorfi* that have been recognized in PS1243 appeared time-coeval with *T. quinqueloba*. Coincidentally, this part of the core is also marked by highest abundances in coccoliths (Bleil and Gard, 1989), providing further evidence that only during this period was there a notable influence of relatively warm Atlantic surface water. Unfortunately, the general lack of *C. wuellerstorfi* during most of MIS 11 makes it impossible to obtain a more complete epifaunal benthic $\delta^{13}C$ record. Therefore, interpretations concerning possible deep-water mass linkages between the Nordic seas and the North Atlantic will remain speculative. However, all our other records clearly indicate that surface and bottom water mass conditions during MIS 11 were distinctly different from the Holocene situation.

5.2. Marine carbonate system

Changes in marine surface water salinity due to freshwater release from a decaying Fennoscandian ice sheet via melting icebergs or bursting proglacial lakes may affect not only the stable isotope chemistry of the upper ocean layer (Bauch and Weinelt, 1997), but also surface ocean bioproductivity. Most coccolithophorids and planktic foraminifera prefer to live under full-marine conditions. In the Nordic seas these two groups are responsible for most of the carbonate deposited during the past 500 ka (Henrich, 1992; Henrich and Baumann, 1994). Analyses of the reflectance of glacial–interglacial sediments from the North Atlantic reveal a high correlation with carbonate content (e.g., Cortijo et al., 1995). In the Nordic seas during MIS 1 and 11, equally high carbonate contents are observed (about 50%), however, they are not associated with equally high changes in sediment reflectance. Across the MIS12/11 transition greylevel changes are of the order of 20% whereas during the last glacial to Holocene transition this change is not more than 7% (Fig. 2). This difference in sediment reflectance between the two interglacials could be shown in a study, which involved analyses of many sediment cores from the Nordic seas (Bauch and Helmke, 1999). In fact it could be also shown that MIS 11 is always the interval with highest sediment reflectance throughout the last 500 ka. Based on several sediment cores from the Nordic seas, carbonate records obtained from the two fractions >63 and <63 μm reveal that the smaller fraction makes up only minor proportions of the bulk carbonate during MIS 11 and 5e (in a core from the Iceland Sea between 6 and 11% is found <63 μm compaired with 40 and 37% in the fraction > 63 μm) whereas up to 25% are found in MIS 1 (Henrich, 1992; Bauch et al., 1999). Therefore, coccoliths cannot account for the changes observed in sediment reflectance. A direct comparison of the reflectance of the washed sediment residues > 125 μm (i.e., foraminifera only) from all three interglacials shows that the whitest foraminiferal tests always occur in MIS 11 (Helmke and Bauch, 1999). As revealed by photographs (SEM), severe corrosion of the surface of the foraminiferal tests probably caused this increase in foraminiferal test reflectance (Bauch and Helmke, 1999). Although all tests in MIS 11 appear rather fragile they are by no means fragmented. A visual check of the sand-size fraction 63–125 μm even reveals small-sized specimens (juvenile) of *T. quinuqeloba*, implying that even the relative proportions of species in the total assemblage of this core may not be significantly altered by preferential dissolution of less calcified species taxa (Berger and Piper, 1972).

It is intriguing to speculate on the cause of such an obvious carbonate dissolution event in MIS 11, considering the relatively shallow depth range of the studied cores. At site PS1243 high plankton bioproductivity during MIS 11 can be interpreted from the occurrence of high abundance of planktic foraminifera. Therefore, high rates in vertical flux of fresh total organic carbon (TOC) may be assumed for this time. This could have changed bottom water CO_2 via remineralization and, thus, calcite solubility. In fact, enhanced vertical flux rates of TOC may be inferred from the presence of certain species of benthic foraminifera in MIS 11 (Struck, 1997), but such high rates cannot be verified by increased TOC concentrations of the bulk sediment (Birgisdottir, 1991).

In other parts of the world ocean, MIS 11 is marked by increased growth rates in coral reefs and high carbonate accumulation in shallow seas (Droxler et al., 1996, Davies et al., 1997), whereas enhanced carbonate dissolution is recognized in sediment records from deep ocean basins (Peterson and Prell, 1985; Bassinot et al., 1994; Howard, 1997b). If the world ocean experienced a massive production of biogenic calcite during MIS 11, this increase could have led to carbonate-ion undersaturation in the deeper ocean and to a shallowing of the world ocean's lysocline. Presumably, changes in glacial to interglacial seawater-pH, i.e., dependence of the pH on the carbonate-ion concentration, may affect the fractionation of carbon isotopes in biogenically precipitated calcite in the way that an increasing pH leads to lower $\delta^{13}C$ values (Sanyal et al., 1995; Spero et al., 1997). Because planktic $\delta^{13}C$ records from the Nordic seas commonly show $\delta^{13}C$ depletion during glaciation and $\delta^{13}C$ enrichment during interglaciations, the Nordic seas records may be linked to global changes in ocean chemistry rather than to regionally confined biological processes. It could still be that the relatively high planktic $\delta^{13}C$ values observed in many Nordic seas sediment records for MIS 11 (Bauch, 1997) were related to increased phytoplankton production in combination with rapid, vertical downward transportation of organic matter, which would have caused a preferential removal of ^{12}C from the surface water. However, $\delta^{13}C$ measured on *N. pachyderma* sin. from surface sediments of the Nordic seas and the Arctic Ocean reveal high values in the seasonally ice-covered western Nordic seas (Johannessen et al., 1994; Sarnthein et al., 1995), and by far the highest values ($> 1.0‰$) are found in the perennially ice-covered Central Arctic Ocean (Spielhagen and Erlenkeuser, 1994). Because of ice coverage and rather cold SSTs (between -1 and -1.5°C), which prevail in these two regions today, it seems that the rate in air–sea exchange of CO_2 as well as the low water temperature may also play a significant role on the $\delta^{13}C$ signal in planktic foraminifera (Johannessen et al., 1994). If true the high $\delta^{13}C$ observed in MIS 11 would in fact argue for cold SSTs in the Nordic seas, in accordance with out interpretation based on subpolar foraminiferal abundance.

6. Summary and implications

An intensive multiparameter investigation based on biogeochemical, sedimentological, and micropaleontological methods was carried out on core PS1243 from the Norwegian Basin in the Nordic seas. The paleoceanographic implications of these proxy data were used for a detailed comparison between two interglaciations, namely, MIS 1 (Holocene) and MIS 11. The study revealed many features of MIS 11, which seem to be rather atypical for an interglaciation at these latitudes. The major results and their interpretation are summarized below:

• The planktic $\delta^{18}O$ record of core PS1243 marks MIS 11, 5e, and 1 as the three warmest intervals during the past 5 climatic cycles (c. 450 ka). However, a direct comparison of the benthic oxygen isotope values reveals heavier oxygen isotope ratios in MIS 11, implying a larger global ice volume during this time.

• In the studied area, full-interglacial conditions without IRD deposition but increased abundance of subpolar foraminifera existed in MIS 11 for about 10 kyr (~ 398–408 ka). The occurrence of subpolar planktic foraminifera (mostly *T. quinqueloba*) during this time indicates the advection of Atlantic surface water. However, in contrast to the Holocene warm SSTs never have expanded over large areas of the Nordic seas. The Atlantic surface waters were probably confined to the eastern part of the Nordic seas.

• The total number of planktic foraminifera (per gram sediment) in MIS 11 is comparable to those in MIS 1. Yet the ratio between planktic and benthic foraminifera, as well as the species composition of the benthic fauna, differ greatly between the two time intervals. This suggests that the linkage between the ocean surface and the deep sea was different also.

• The general lack of all epibenthic-living foraminiferal species in MIS 11, in contrast to their relative abundance in MIS 1 and other interglaciations, may suggest a different mechanism in the downward flux of food supply through the water column.

• Despite comparable values in carbonate content (% weight), reflectance measurements of the total sediment (% greylevel) show much lighter values for MIS 11 than for MIS 1. This higher reflectance of sediment in MIS 11 is attributed to increased corrosion of foraminiferal tests. The reason for this enhanced carbonate corrosion remains speculative, but may be linked to the global marine carbonate system during MIS 11.

The oceanic evolution throughout MIS 11 does not match the conditions that have prevailed in the Holocene for the past 10 kyr. Particularly the timing of the faunal proxies relative to the insolation record, seems a very striking difference between the two interglaciations (Fig. 6). The species composition of the planktic foraminiferal assemblage clearly reveals that comparatively cold SSTs prevailed in the investigated area during MIS 11. The few data available from planktic foraminiferal assemblages further east also confirm the presence of only small proportions of warm-water indicating species in MIS 11 (Bauch, 1997), implying relatively little advection of warm Atlantic surface water towards the Nordic seas. Previous SSTs estimates for the Northeast Atlantic indicated similar temperatures for past interglaciations, e.g., MIS 1, 5e, and 11 (Ruddiman et al., 1986). However, recent high resolution studies in this area reveal, for instance, the total disappearance of the polar species *N. pachyderma* sin. during MIS 5e

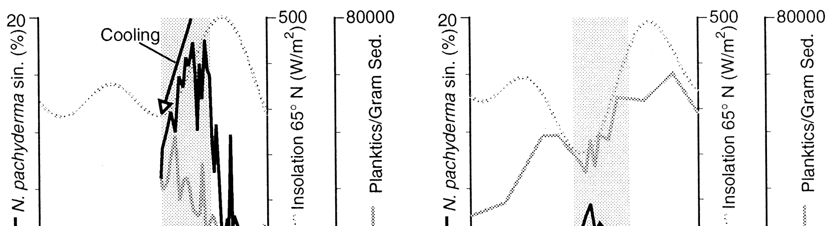

Fig. 6. Comparison of planktic foraminiferal records of the two investigated interglaciations relative to insolation forcing. The shaded area denotes the full-interglacial phase and is the time of no IRD deposition. Based on the proportion of the polar species *N. pachyderma* sin. relative to the warmer water species, the interval of smallest global ice volume (see Fig. 4b) also coincides with the lack of IRD and can be regarded as time of a Holocene-like thermohaline circulation. Taking the record of MIS 11 and MIS 1 at face value the future climate of the Nordic seas is directly heading towards polar conditions (inferred from the stippled curve of *N. pachyderma* sin.). In such a case the intermediate level in insolation noted for the next 20 ka would not at all counteract this cooling tendency.

(e.g., Oppo et al., 1997), whereas in MIS 11 this species still shows some significant fluctuations of up to 10% in its record (Oppo et al., 1998). It, therefore, appears as if comparatively less surface ocean heat was tranported to the Nordic seas during MIS 11. It would be interesting to know to what extent this decreased northward transfer of ocean heat affected deeper water circulation in general, and the climatic conditions on land in particular.

Because of the profound discrepancies between MIS 1 and 11, as recorded by most of our proxy data, it seems unlikely that MIS 11 can be taken as an analogue for the Holocene period in general. Considering the fact that the Nordic seas is a key region for the climate system of MIS 1, it appears unlikely that this region can be neglected when evaluating the question of whether MIS 11 can serve as analogue for MIS 1. However, if we did adopt MIS 11 as an analogue for MIS 1, then we would predict that the present climate in the Nordic seas region would continue to deteriorate, as it has for the past 6 kyr (Fig. 6). In this ‘‘natural’’ future scenario, insolation at intermediate values would not halt this cooling process.

The glacial–interglacial climate system has evolved in a complex way throughout the past 1 Ma. Mechanisms which we regard as highly important today are assumed to have also acted as major driving forces in former times. However, given the complexity of each interglaciation it may not be possible to forecast the future climate by comparison with a time interval, such as MIS 11, which appears most like the present. Detailed studies of other interglacial periods are necessary to define boundary conditions of the general interglacial climate system, and to identify as many modes and varieties of interglacial conditions as possible. This information can be used to assess the present climate in a way that will support the modelling of future climates.

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