Interpreting Glacial-Interglacial Changes in Ice Volume and Climate From Subarctic Deep Water Foraminiferal $\delta^{18}O$

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Benthic foraminiferal $\delta^{18}O$ data from sediment cores of the deep Nordic seas were investigated, in combination with proxy records of carbonate content, iceberg-rafted debris (IRD), and planktic foraminiferal $\delta^{18}O$, to reconstruct and interpret some major ocean changes in this climatically sensitive region over the last 5 climate cycles. In particular, complete interglacial cycles were studied in more detail, i.e., time intervals that always include a glacial maximum, the ensuing peak warm period, and the inception of glacial conditions. Marine isotope stages (MIS) 11, 5e, and 1 have been identified as the three most pronounced interglacial periods. Of the three glaciations (MIS 12, 6, 2) that preceded these warm climate intervals, MIS 12 is recognized as the one when ice volume was largest and, consequently, sea level lowest. Of all peak interglacial periods studied, MIS 5e had the smallest ice volume whereas interglacial intervals MIS 11 and 1 show similar $\delta^{18}O$ values, indicating ice volumes and sea levels of comparable magnitude. The early parts of the interglacial-to-glacial transitions, which followed upon the peak interglacial interval in MIS 11 and 5e, show the first significant increase in benthic $\delta^{18}O$ almost time-coeval with a recurrence of IRD. A similar finding is noted in the proxy records from the youngest Holocene sediments. From this it may be concluded that the present oceanic conditions in the Nordic seas are in such a critical state that significant effects on the thermohaline system cannot be precluded for the time to come.

1. INTRODUCTION

Oxygen isotope records from deep sea sediments reveal that, for the past 600 ka or so, the Pleistocene climate underwent major changes from glacial to interglacial conditions [e.g., *Tiedemann et al.*, 1994]. The waxing and waning of ice sheets caused global sea level to fluctuate, leaving an imprint in the stabel oxygen isotope composition ($\delta^{18}O$) of marine biogenic calcite. The change in $\delta^{18}O$ is about 1.25 ‰ in the subtropical North Atlantic region since the last glacial maximum (LGM) some 20 kyr ago [*Fairbanks and Matthews*, 1978] and has been related to a sea-level rise of between 105-126 m [e.g., *Fairbanks*, 1989; *Peltier*, 1998]. Benthic foraminiferal oxygen isotope records infer that the rise in sea level after the penultimate glaciation, marine isotope stage (MIS) 6, to the highstand in MIS 5e (the last peak interglaciation) must have been at least as large as that after the LGM [*Chappell and Skackleton*, 1986; *Shackleton*, 2000]. U-Th analyses of corals indicate a sea level during MIS 5e several meters higher than in the Holocene [e.g., *Chen et al.*, 1991; *Gallup et al.*, 1994]. For older interglacial and glacial periods there

Earth's Climate and Orbital Eccentricity:
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Geophysical Monograph 137

10.1029/137GM07

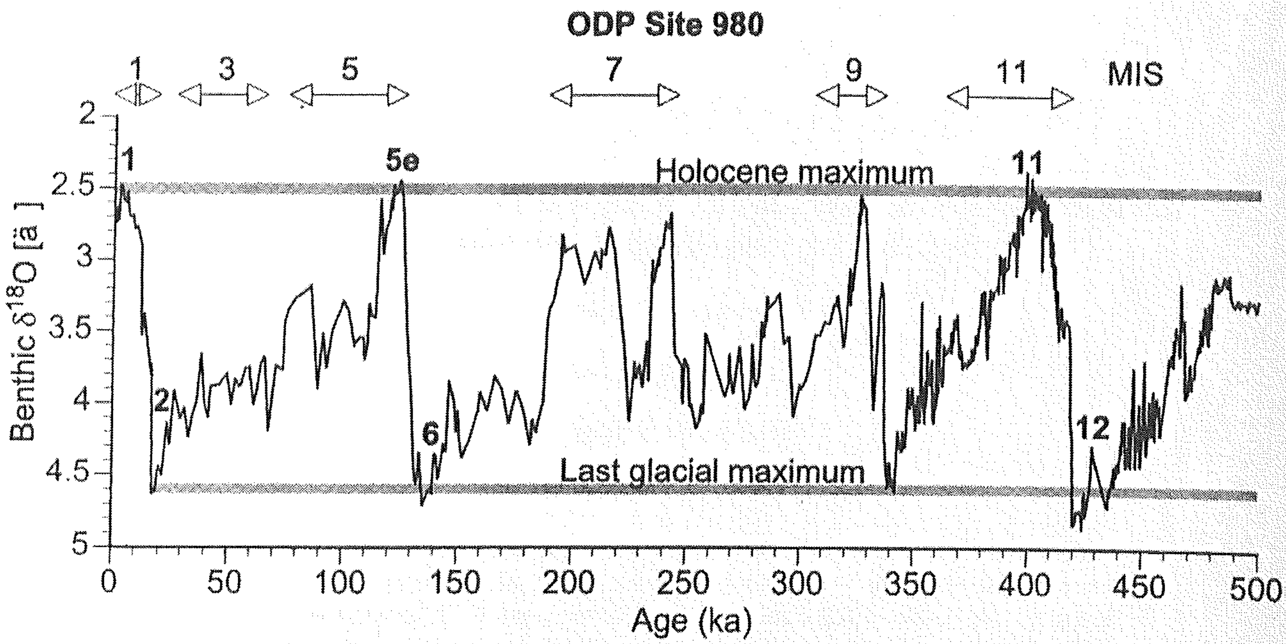

Figure 1. Benthic foraminiferal $\delta^{18}O$ record (*C. wuellerstorfi*) from the northeast Atlantic (55°29'N/14°42'W, 2179 m water depth; from *McManus et al.* [1999]). Note the differences between peak interglacial (MIS 1, 5e, and 11) and peak glacial $\delta^{18}O$ values (MIS 2, 6, and 12).

are only few records that give some insight into past sea-level highstands and estimates of global ice volumes [e.g., *Chappell and Skackleton*, 1986; *Bard et al.*, 1996]. A recent investigation from the Red Sea, which considered the possible sill depth variations, implies major sea-level lowstands for peak glacial periods such as MIS 12, 6, and 2; sea-level lowering during MIS 10 and 8 was comparatively less severe [*Rohling et al.*, 1998].

Many sea-level reconstructions using paleoshoreline evidence (raised reefs and beaches) were made in tectonically unstable regions and, thus, strongly depend on assumptions of the uplift rates. Similar complications are found in those regions that were directly affected by isostatic crustal response to changes in the load of the ice sheets themselves. Moreover, since the major ice sheet fluctuatinos were located in the northern hemisphere, the relative sea level during and after deglaciations did not rise simultaneously on a global scale [*Lambeck and Nakado*, 1992].

In contrast to paleoshorelines, deep-sea $\delta^{18}O$ records not only contain information of local or regional importance. Because of their ocean-wide connection due to global circulation, they should also record global changes in climate, ice volume and sea level. As the amount of continental ice fluctuated due to either growth or melting, the surface and deep ocean also underwent substantial changes in both circulation and water mass properties [e.g., *Oppo and Lehman*, 1995]. The changes in global ice volume over glacial-interglacial timescales are most readily recognizable in benthic foraminiferal oxygen isotopic records (Figure 1). However, deep-sea $\delta^{18}O$ records over glacial-interglacial cycles are also affected by temperature changes [*Mix et al.*, 1999; *Shackleton*, 2000] and a variable gradient in the $\delta^{18}O$ signature between major ocean basins [*Labeyrie et al.*, 1987; *Zahn and Mix*, 1991; *Dokken and Jansen*, 1999].

Because the nearby landmasses became glaciated and deglaciated many times over the last glacial-interglacial cycles, the climatically sensitive surface ocean properties in the northeast Atlantic alternated accordingly [e.g., *Broecker and Denton*, 1989; *Rahmstorf*, 1994]. Long time intervals with cold sea-surface temperatures were interrupted only briefly by much warmer episodes [*Ruddiman et al.*, 1986], conditions which are also reflected in the benthic foraminiferal $\delta^{18}O$ records (Figure 1).

In contrast to the northeast Atlantic, investigations of glacial-interglacial sediments applying a limited number of proxies from further north in the subarctic latitudes (Nordic seas) have provided only little information on times when oceanic conditions were similar to the Holocene [e.g., *Kellogg*, 1980]. So far, detailed studies of peak-interglacial intervals have revealed that MIS 1, 5e, and 11 are the only three intervals of the past 450 ka when specific faunal and sedimentological proxies, such as the appearance of warm surface water foraminifera and the lack of iceberg-rafted debris (IRD), would indicate comparable climatic boundary conditions [*Bauch et al.*, 1996; *Bauch et al.*, 2000b]. However, whereas changes in benthic foraminiferal $\delta^{18}O$

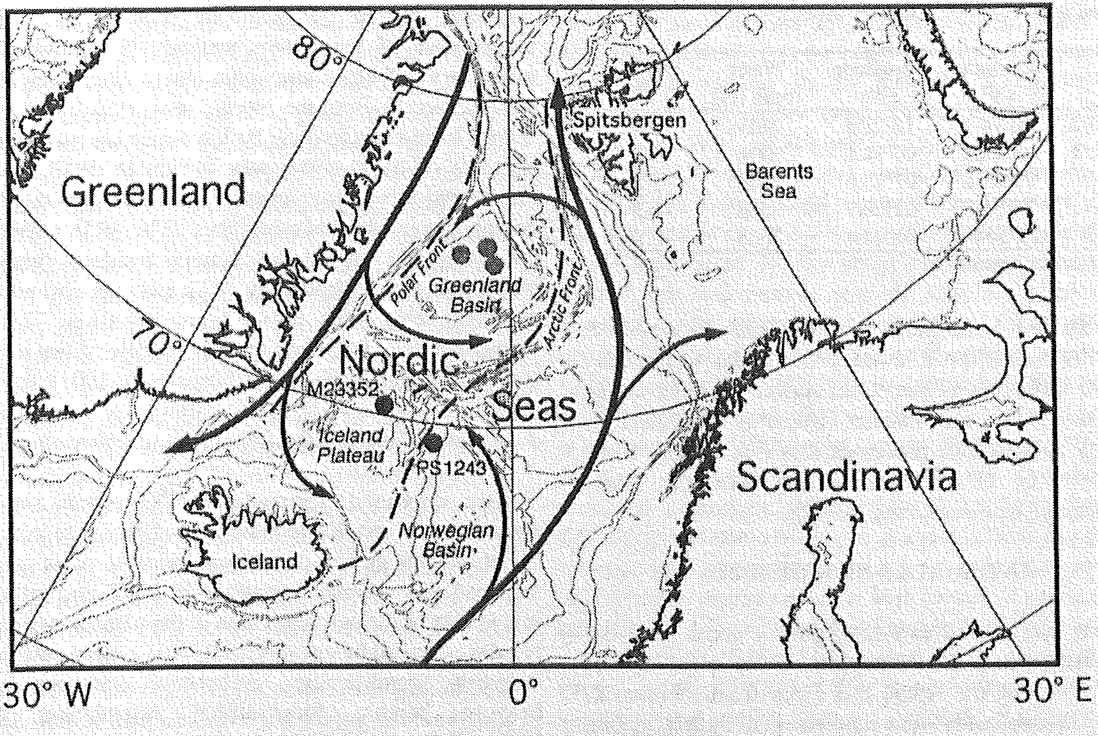

Figure 2. Map of the Nordic seas showing the positions of investigated cores (black dots) and coretop samples (grey dots) from the Greenland Basin (see also Table 1). Also indicated are the modern surface circulation (arrows) and the two major oceanic fronts that divide the water masses of the Nordic seas into three domains.

from sediment cores in the North Atlantic are interpreted as evidence for a variable thermohaline circulation and overall climate [e.g., *Adkins et al.*, 1997], little has been investigated as to this aspect on the basis of deep water $\delta^{18}O$ downcore records from the Nordic seas.

This study is intended to determine whether benthic $\delta^{18}O$ records from the Nordic seas can be used to estimate relative changes in deep water conditions across major interglacial cycles and whether these changes also relate to variations in the overall climate. While glacial-to-interglacial deep water benthic isotope records from outside the Nordic seas usually exhibit $\delta^{18}O$ decreases on the order of 2 ‰ or more due to concomitant increase in bottom water temperature (see also Figure 1), records from the Nordic seas covering the last two major climate transitions (Termination 1 and 2) show much less change, ≤ 1 ‰ [*Labeyrie et al.*, 1987; *Bauch et al.*, 1996; *Bauch et al.*, 2001]. This is intriguing, because measurements of glacial-interglacial porewater $\delta^{18}O$ also indicate a relatively small change in deep-water $\delta^{18}O$ [*Schrag et al.*, 1996; 1997]. The implications are that the relationship between paleoseawater $\delta^{18}O$ and past sea levels appears different from what has been previously estimated, and further, that both paleoseawater and benthic foraminifera from Nordic seas sediments have comparable glacial-to-interglacial changes in deep water $\delta^{18}O$.

2. OCEANOGRAPHY SETTING

The modern surface water pattern of the Nordic seas is characterized by a threefold subdivision based on salinity and temperature variations which decrease towards the north and the west [*Swift*, 1986]. Warm and high-salinity Atlantic surface water flows along the eastern periphery of the Nordic seas into the Arctic Ocean as the Norwegian and West Spitsbergen currents (Figure 2). The western part of the Nordic seas is characterized by polar waters that flow as East Greenland Current from the Arctic Ocean southward into the North Atlantic. Polar and Atlantic waters are separated from each other by the "Arctic Water", which is essentially mixed from the other two. The Arctic Water is bounded by distinct polar and arctic fronts. Deep water formation takes place mainly in the Greenland and Iceland seas, i.e., within the Arctic Water domain. This process is an essential component of the climate-driving thermohaline circulation and forms the main source of North Atlantic Deep Water [e.g., *Broecker*, 1991; *Aagaard and Carmack*, 1994].

In contrast to the steep salinity and temperature gradients between the surface water masses, the physical parameters of the deeper waters are more even throughout the Nordic seas. However, bottom water temperatures (BWT) in areas with water depths below 1,600 m still vary from basin to

Table 1. Sediment cores and coretops investigated.

Cores	Latitude (N)	Longitude (W)	Water depth (m)	BWT* (°C)
M23352-1/2	70°00.4	12°25.8	1819	-0.9
PS1243-1/2	69°22.3	06°32.1	2710	-0.9
PS1895-9	75°24.8	07°18.6	3358	-1.3
PS1900-7	74°31.7	02°20.1	3538	-1.3
PS1901-1	75°56.6°	03°44.4	3588	-1.3

* bottom water temperature

basin. In the Greenland Basin, with water depths around or below 3,500 m, the BWT is about -1.3°C due to intensive deep water formation as a result of surficial cooling during winter. On the Iceland Plateau, which is also strongly affected by convectional processes, BWT is usually of the order of -0.9°C, as it is also in the Norwegian Basin [*Environmental Working Group*, 1998].

3. MATERIALS AND METHODS

We have investigated two long gravity cores from the Iceland Plateau and the Norwegian Basin close to the present position of the Arctic and Polar fronts (Figure 2, Table 1). The two cores were supplemented by trigger box cores from the same sites to have an undisturbed sediment sample of the uppermost Holocene. Measurements of benthic foraminiferal isotopes from the surface sediment at the site in the Norwegian Basin (site PS1243; BWT -0.9°C) were compared with surface sediments from the deep Greenland Basin (BWT -1.3°C). The surface sample from site PS1243 was treated with Rose Bengal so that comparative isotope measurements could be made on stained (living) and unstained (dead) specimens.

Most of the stable isotope analyses were carried out in the Leibniz Laboratory at Kiel University using the fully automated Kiel Carbonate Preparation Device and a Finnigan MAT 251 mass spectrometer system. The analytical accuracy of this system is ± 0.08 ‰ for $\delta^{18}O$ and ± 0.05 ‰ for $\delta^{13}C$. All measurements were calibrated to Pee Dee Belemnite isotope scale (PDB) via the NBS 20 carbonate standard. Benthic foraminifera from the last glacial to Holocene section in core M23352 were measured using the Kiel Device Finnigan MAT 252 system at GEOMAR. On average, 5-7 specimens (size fraction 250-500 μm) of *Cibicidoides wuellerstorfi* and between 20-30 specimens (size fraction 125-250μm) of *Oridorsalis umbonatus* were used for the isotopic analyses. In addition to these multiple-specimen samples, three single-specimen measurements were made on *C. wuellerstorfi* from MIS 2. To support the general downcore isotope stratigraphy of the cores, $\delta^{18}O$ records of the polar planktic species *Neogloboquadrina pachyderma* sinistral (sin.) were produced taking 25-30 specimens per sample.

Beside stable oxygen isotopes, the downcore stratigraphy of the two gravity cores was compared with records of carbonate content and sediment reflectance. A previous comparison of these two methods has provided good evidence for their stratigraphic use in correlating between sediment cores from the Nordic seas [*Bauch and Helmke*, 1999]. In addition to these bulk sedimentary parameters, the contents of lithic grains was studied in detail. This aids interpretation of the climatic implications of the IRD in comparison with the other data. The IRD counts were conducted on the dried sample residues (grain size >250μm), supplemented by IRD analyses carried out in one-centimeter steps using X-ray negatives. This latter methods allows for counting lithic particles down to a grain size of 500μm [*Grobe*, 1987; *Birgisdottir*, 1991].

3.1. Benthic Foraminiferal Background Information

Deep sea benthic forminiferal $\delta^{18}O$ records are usually considered to be reliable indicators of glacial to interglacial variations in global ice-volume because it is generally assumed that regional variations in BWT and salinity are less severe at greater depth than in the planktic realm. From Nordic seas sediment cores it has proven difficult to establish monospecific, epibenthic downcore records because benthic foraminiferal assemblages change drastically over glacial-interglacial cycles. In particular, epibenthic species are often missing in glacial core intervals [e.g., *Struck* 1995]. To overcome this problem, previously published long benthic oxygen isotope records from the deeper parts of the Nordic seas were commonly spliced together using the epibenthic species *C. wuellerstorfi* for the 'warmer' core sections (where this species is more common) and the shallow-endobenthic species *O. umbonatus* for the glacial intervals [*Labeyrie et al.*, 1987; *Bauch et al.*, 1996]. Both species depart from isotopic calcite equilibrium with apparently constant values [*Graham et al.*, 1981; *Belanger et al.*, 1981; *Labeyrie et al.*, 1987].

More recently, glacial-interglacial downcore studies on *O. umbonatus* and *C. wuellerstorfi* have shown that particularly the epifaunal species *C. wuellerstorfi* exhibits a kind of 'vital effect' that causes a much larger offset between the two species during peak glacial and early deglacial periods than had been reported before [*Veum et al.*, 1992; *Bauch et al.*, 2000a; *Bauch et al.*, 2001]. However, both species maintain a constant offset to each other during peak interglacial times as well as the subsequent post-interglacial transitions, when glacial conditions were not yet fully developed [*Vogelsang*, 1990; *Didié and Bauch*, in press].

4. RESULTS

4.1. Downcore Records

The general downcore stratigraphy of the two gravity cores PS1243 and M23352 has been described in previous studies [*Bauch and Helmke*, 1999; *Bauch et al.*, 2000a].

The records indicate that the two cores date back to MIS 12 (Figure 3); core PS1243 is used here as reference because more downcore parameters are available than for M23352. In both cores, three major interglacial peaks are recognized and identified as MIS 11, 5e, and 1. These three peak interglaciations are marked by increased carbonate content with sediment reflectance values being highest in MIS 11. These high values are due to corroded foraminiferal tests [*Bauch and Helmke*, 1999], making the entire sediment to reflect lighter in MIS 11 than in any other peak interglacial core sections, e.g., MIS 5e and 1. The prominent MIS 11 reflectance peak is a useful stratigraphic marker and can be traced to many other core sites in the southern Nordic seas, including M23352. In M23352, chronological interpretation on the basis of planktic oxygen isotopes alone is not straightforward beyond MIS 7 because the record seems to be affected by meltwater overprints probably due to the proximity of the core site to polar surface waters.

In the two cores, the carbonate-rich peaks of MIS11, 5e, and 1, always appear together with low IRD; however, it is noted that substantial IRD is still found in core M23352 during MIS 11. No other core intervals reveal a similar relation between these two proxy records. Peak glacial core intervals as well as their deglacial transitions (terminations) are often marked by high IRD values. The highest numbers of lithic grains are recognized in MIS 12, followed by MIS 6. In contrast, MIS 2 is characterized by comparatively low IRD content. In combination with the IRD records, the high benthic $\delta^{18}O$ values found in MIS 12 and 6 give evidence of extensive glaciations during these two time intervals. This latter finding seems to be consistent with the major glacial trends observed in the benthic $\delta^{18}O$ record from the northeast Atlantic (Figure 1).

4.2. *Species-Dependent Oxygen Isotopic Offsets*

The already published benthic $\delta^{18}O$ record shown in Figure 3 [*Bauch et al.*, 2000b] had a low spatial resolution and was spliced together using data of *C. wuellerstorfi* and *O. umbonatus* [cf. *Labeyrie et al.*, 1987]. Therefore, this record does not resolve the inconsistency in the species-dependent $\delta^{18}O$ offset of the two species during times with increased IRD deposition as it was previously described for MIS 6 and Termination 2 in core M23352 [*Bauch et al.*, 2000a]. Such a relationship between enhanced deposition of IRD and significant decreases indeep water endobenthic and epibenthic $\delta^{18}O$ has also been found now for parts of MIS2 and Termination 1 in core PS1243 [*Bauch et al.*, 2001]. Similar findings, but based on endobenthic species only, have been reported from shallower areas (<1000 m water depth) of the Nordic seas [e.g., *Rasmussen et al.*, 1996; *Dokken and Jansen*, 1999]. While some authors believe that formation of brines ejected from sea ice is mainly responsible for the occurrence of unusually low benthic $\delta^{18}O$ anomalies during glacial times, others would relate them primarily to a change in overall circulation that could have resulted in an increase of bottom water temperatures [*Rasmussen et al.*, 1996; *Bauch et al.*, 2000a; *Bauch et al.*, 2001; *Bauch and Bauch*, 2001].

The densely sampled and well-dated upper portion of core PS1243 demonstrates some of the important changes that occurred in the deep Norwegian Basin during the past 30 cal. ka (Figure 4). For better comparison, the isotopic data of the two benthic species were corrected onto a common scale using their well-known departures from isotopic equilibrium (+0.64 ‰ for *C. wuellerstorfi*, +0.36 ‰ for *O. umbonatus*). Despite this correction, the two $\delta^{18}O$ records reveal a new offset which starts at about 26 cal. ka, at a time when IRD input steeply increased. The $\delta^{18}O$ difference between *C. wuellerstorfi* and *O. umbonatus* remained significant until about 10 cal. ka when IRD deposition had come to an end. Since that time, site PS1243 changed into a predominantly pelagic type of depositional environment, characterized by high carbonate sedimentation and relatively constant benthic foraminiferal $\delta^{18}O$ levels.

Comparison of the $\delta^{18}O$ record of *O. umbonatus* with that of the planktic foraminifer *N. pachyderma* sin. shows a similar trend which persists through the last glacial maximum (around 20 cal. ka) and the ensuing deglaciation. Both species, for instance, exhibit heaviest values near 20 cal. ka and several time-coeval $\delta^{18}O$ depletions during the glacial-interglacial transition. The LGM $\delta^{18}O$ value of ~5.3 ‰ for *O. umbonatus* in the Nordic seas is comparable to maximum values found in *C. wuellerstorfi* for this time in many cores from the North Atlantic [*Sarnthein et al.*, 1994]. In contrast to *O. umbonatus*, the $\delta^{18}O$ values of *C. wuellerstorfi* from peak glacial core sections of the Nordic seas are incompatible with data from other regions, limiting its use as an estimator of relative global ice volumes. However, $\delta^{18}O$ values of *C. wuellerstorfi* and *O. umbonatus* correspond well when the glacial climate mode is only weakly developed, e.g., during the interval MIS 5d-5a [*Vogelsang*, 1990; *Didié and Bauch*, in press].

The IRD records show that the two different counting methods applied produce quite comparable results (Figure 4). While in both records IRD deposition nearly ceased at about 10 cal. ka, the $\delta^{18}O$ values in the two benthic foraminifera continue to decrease until about 7-8 cal. ka. As is known from studies around the Laurentide ice sheet and Caribbean coral data, the northern hemisphere deglaciation finally came to an end near 8 cal. ka, after which the global sea-level rise slowed [*Fairbanks*, 1989; *Barber et al.*, 1999]. Therefore, the decrease in benthic $\delta^{18}O$ observed between 10 and 7 cal. ka is probably related to a continuing rise in the sea level during the first half of the Holocene. In more recent time, we note a consistent recurrence of lithic grains in both IRD records after ~1.5 cal. ka, which was preceded by a slight but detectable increase in benthic $\delta^{18}O$ (Figure 4).

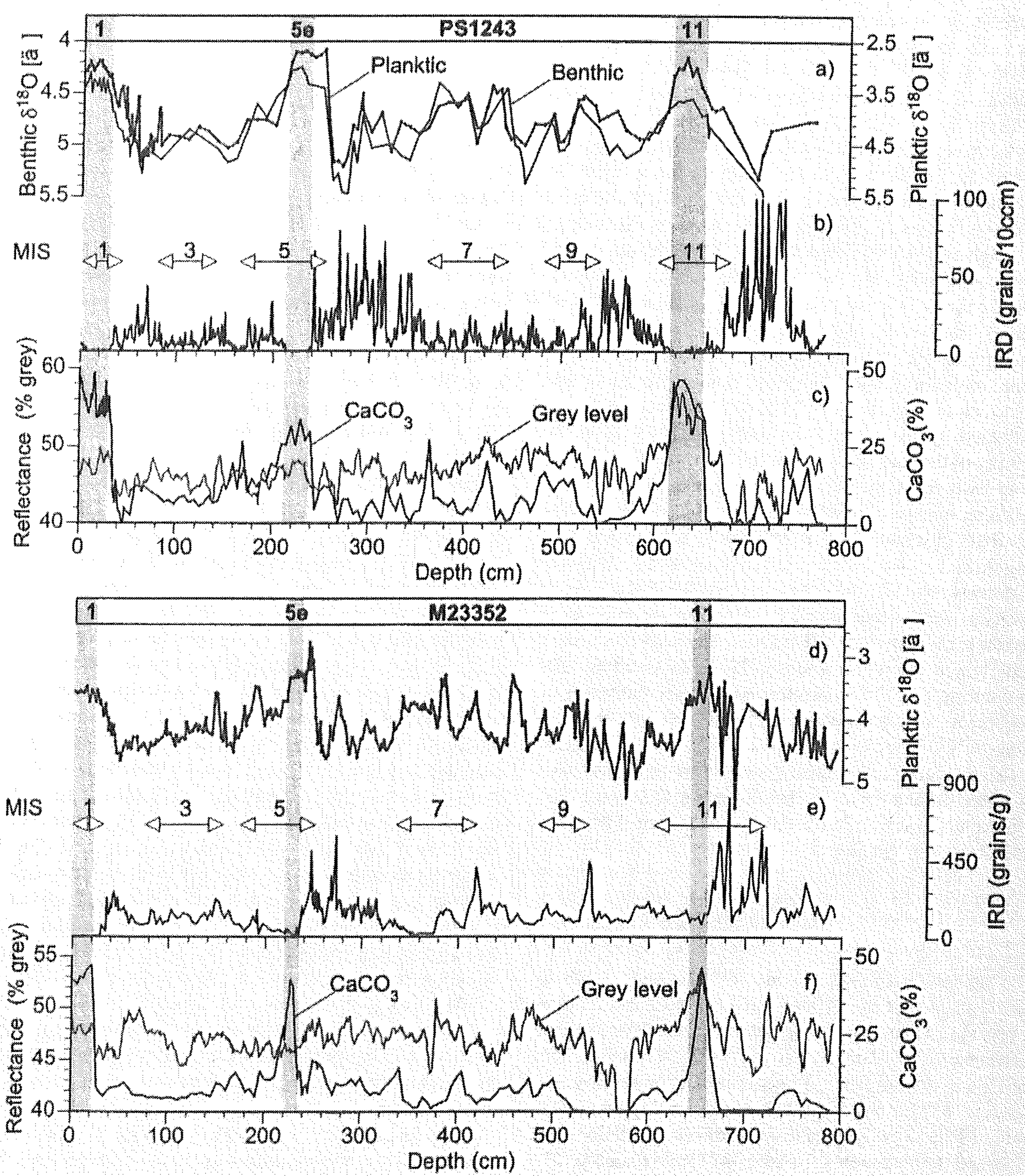

Figure 3. Downcore proxy records from cores PS1243 and M23352 showing the stratigraphic subdivison as marine isotope stages (only odd numbers shown). The grey vertical bars indicate the three most dominant interglacial intervals. These are characterized by pelagic depositional regimes with low IRD input and high sedimentation of microfossil carbonate. (a) Spliced benthic foraminiferal $\delta^{18}O$ record (grey line) of *C. wuellerstorfi* and *O. umbonatus* (species-dependent isotopic offset corrected by +0.64 ‰ and +0.36 ‰, respectively) in comparison with the $\delta^{18}O$ record of the polar planktic foraminifera *N. pachyderma* sin. (from *Bauch et al.*, 2000b). (b) Number of lithic grains (grains >500μm per 10 ccm of total sediment) counted from X-ray negatives in one centimeter steps. (c) Records of carbonate content (weight %) and sediment reflectance expressed on the greylevel scale (0 = black; 100 = white). (d) As in (a), (e) Number of lithic grains per gram of total sediment (>250μm), (f) as in (c).

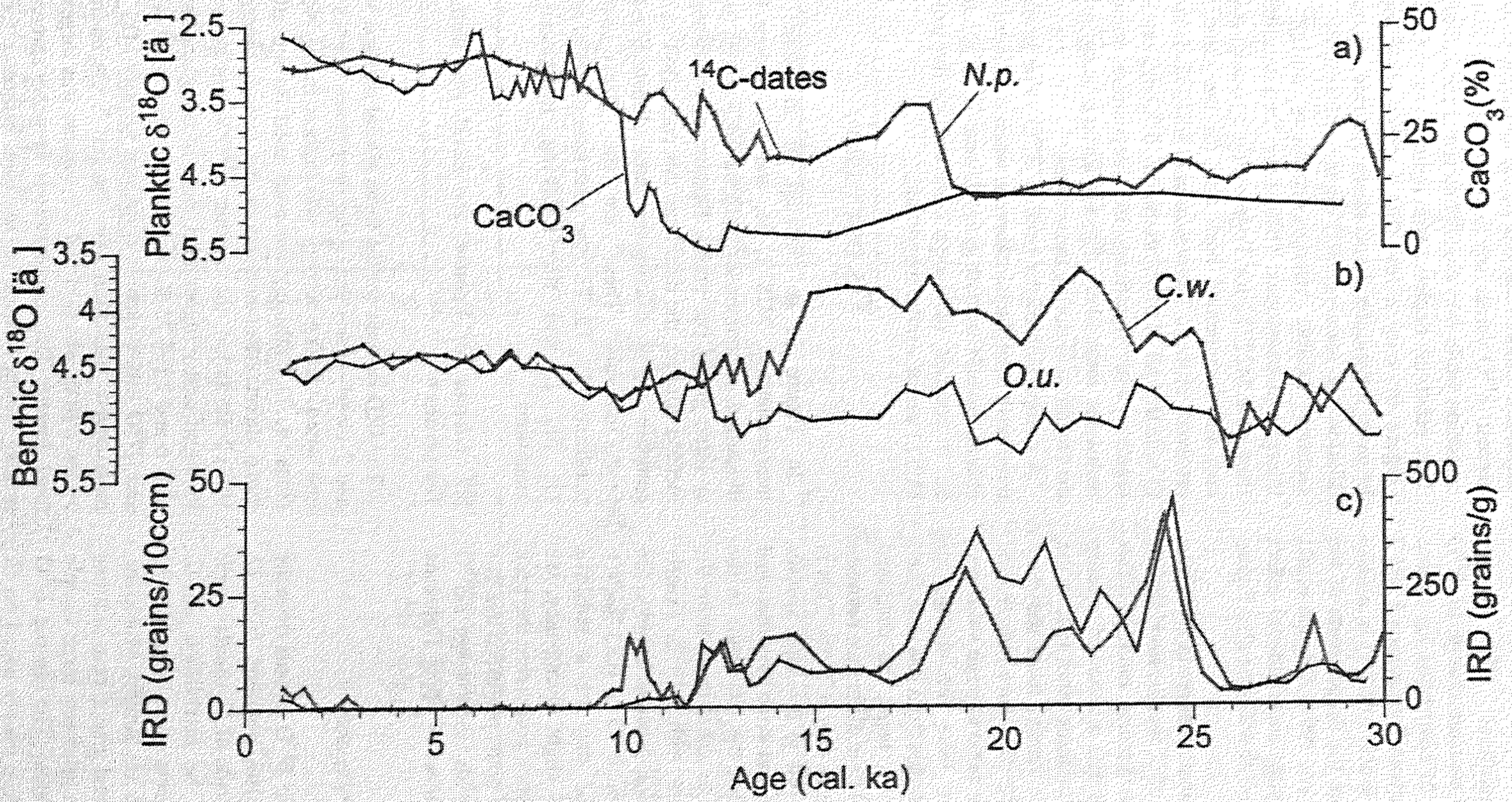

Figure 4. Proxy records from the AMS-radiocarbon dated upper section of core PS 1243. Black dots in (a) indicate measured AMS ^{14}C-dates (chronology from *Bauch et al.* [2001]). (b) Comparison of the two $\delta^{18}O$ records from epibenthic species *C. wuellerstorfi* and endobenthic species *O. umbonatus* (records are on a common scale by adjusting for species-dependent offsets shown in Figure 2). (c) Comparison of the two methods for counting IRD. Grey line indicates the X-ray method (grains >500μm per 10 ccm of total sediment), black line was counted on the basis of washed sample residues (grains >250μm per gram of total sediment).

4.3. Holocene and Coretop Benthic $\delta^{18}O$

Deep water benthic foraminiferal $\delta^{18}O$ records commonly reveal a much higher absolute change in $\delta^{18}O$ than can be attributed to melting of continental ice masses alone. The benthic $\delta^{18}O$ change across Termination 1 and 2 (glacial-interglacial transitions MIS6/5, MIS2/1) is usually of the order of 2 ‰ in cores from the deeper parts of the North Atlantic [e.g., *Sarnthein et al.*, 1994; *Keigwin et al.*, 1994; *Oppo et al.*, 1997], implying an additional effect caused by an increase in BWT of ~3 to 4°C (see also Figure 1).

In order to investigate whether small changes in BWT can be detected in the $\delta^{18}O$ signature of *C. wuellerstorfi* and *O. umbonatus* and to determine the precise modern offset between the two species, we analyzed a series of coretop samples from core PS1243 and from three other sites in the Greenland Basin (Figure 5). The results from the deep Greenland Basin show a mean $\delta^{18}O$ difference between the two species of 0.36 ‰. Exactly the same average value was determined for living specimens from the coretop of PS1243 and for the topmost 19 samples from this core (i.e., the postdeglacial section). However, using the unstained samples from the coretop of PS1243 the difference between *C. wuellerstorfi* and *O. umbonatus* is only 0.23 ‰. This is due to lighter $\delta^{18}O$ values observed in the latter species alone (Table 2), which could imply that some bioturbational bias is involved. On the other hand, the difference in the mean $\delta^{18}O$ values of *O. umbonatus* (unstained) from the Greenland and Norwegian basins is with 0.13 ‰ exactly what can be expected from the BWT gradient between the two regions (assuming a change in $\delta^{18}O$ of 0.23 ‰/°C). This suggests that *O. umbonatus* may be a sensitive temperature recorder. Despite this slight discrepancy, we decided to use the 0.36 ‰ offset when comparing the peak interglacial sections on the basis of both species; to use this larger offset is justified because it is also the average $\Delta\delta^{18}O$ value calculated from the large number of upper Holocene samples.

4.4. Comparison of Downcore $\delta^{18}O$ Records

For directly comparing the benthic $\delta^{18}O$ records from the various core sections all $\delta^{18}O$ data obtained from *O. umbonatus* were calculated to the scale of *C. wuellerstorfi;* no complete $\delta^{18}O$ record could be established on *C. wuellerstorfi* alone because this species occurs in MIS 11 only during its later phase [*Bauch et al.*, 2000b]. Moreover, *C. wuellerstorfi* is rare in deglacial and glacial core sections and specimens are found only when relatively large sample volumes are taken [*Bauch et al.*, 2000a]. The benthic $\delta^{18}O$ data depicted in Figure 6 are therefore mainly based on *O. umbonatus*. We used isotopic data of *C. wuellerstorfi* only for the IRD-free section of MIS 5e in core PS1243 (above 237 cm core depth).

The three investigated climate intervals reveal several important features that can be recognized in both cores. The highest $\delta^{18}O$ values are found during MIS12 and its early transition while the values observed at the end of MIS 6 and 2 are relatively depleted with lightest values found in MIS 2 (Figure 6); only a few data are available from MIS 12 because foraminiferal specimens are rare, probably due to dilution by massive IRD input. All benthic $\delta^{18}O$ core sections exhibit strong depletions during glacial-to-interglacial transitions. These have been described in detail from core M23352 for Termination 2 [*Bauch et al.*, 2000a] and are documented here also for the same, but better resolved, climatic interval in core PS1243. It is significant to note that all of these prominent $\delta^{18}O$ depletions occurred together with significant IRD deposition, indicating that meltwater most likely affected surface ocean properties in the Nordic seas during these times.

A comparison of the benthic $\delta^{18}O$ value levels from the three peak interglacial intervals, when IRD input was low or absent, reveals in both cores lowest $\delta^{18}O$ values for MIS 5e. In contrast, the values observed for the Holocene and MIS11 are of similar magnitude.

5. DISCUSSION

In Figure 7, a close-up of the three dominant interglaciations, MIS 1, 5e, and 11, in core PS1243 is illustrated in comparison with the IRD record (now shown as lithic grains per gram). Because of the dense sampling method applied, the data from this core allow us to directly compare and to discuss the three climatically interesting time intervals. These are the glacial-to-interglacial transitions (T1, T2, T5), the peak interglacial phases of MIS 11, 5e, 1, and the following interglacial-to-glacial transitions.

5.1. Glacial Ice Volume, Sea Level, and Bottom Water Temperature

The IRD records for the three glacial maxima MIS 12, 6, and 2 clearly characterize the final phase of MIS 12 as the one when iceberg activity was highest in the Nordic seas (Figs. 6, 7). Although the units used do not allow us to make direct statements about the overall severity of the MIS12 glaciation, the benthic $\delta^{18}O$ values are with ~5.10 ‰ significantly higher than those of the glacial maxima of MIS 6 and 2 (~4.75 ‰ and 4.55 ‰, respectively). The IRD and $\delta^{18}O$ proxy records combined identify MIS 12 as the most severe glaciation over the last 5 glacial-interglacial cycles during which global ice volume was largest and sea level lowest. This conclusion is in good agreement with similar findings based on sea-level estimates for the Red Sea [*Rohling et al.*, 1998], sea water $\delta^{18}O$ calculations for the North Atlantic [*McManus* et al., 1999], as well as other deep-sea benthic $\delta^{18}O$ records [*Shackleton*, 1987].

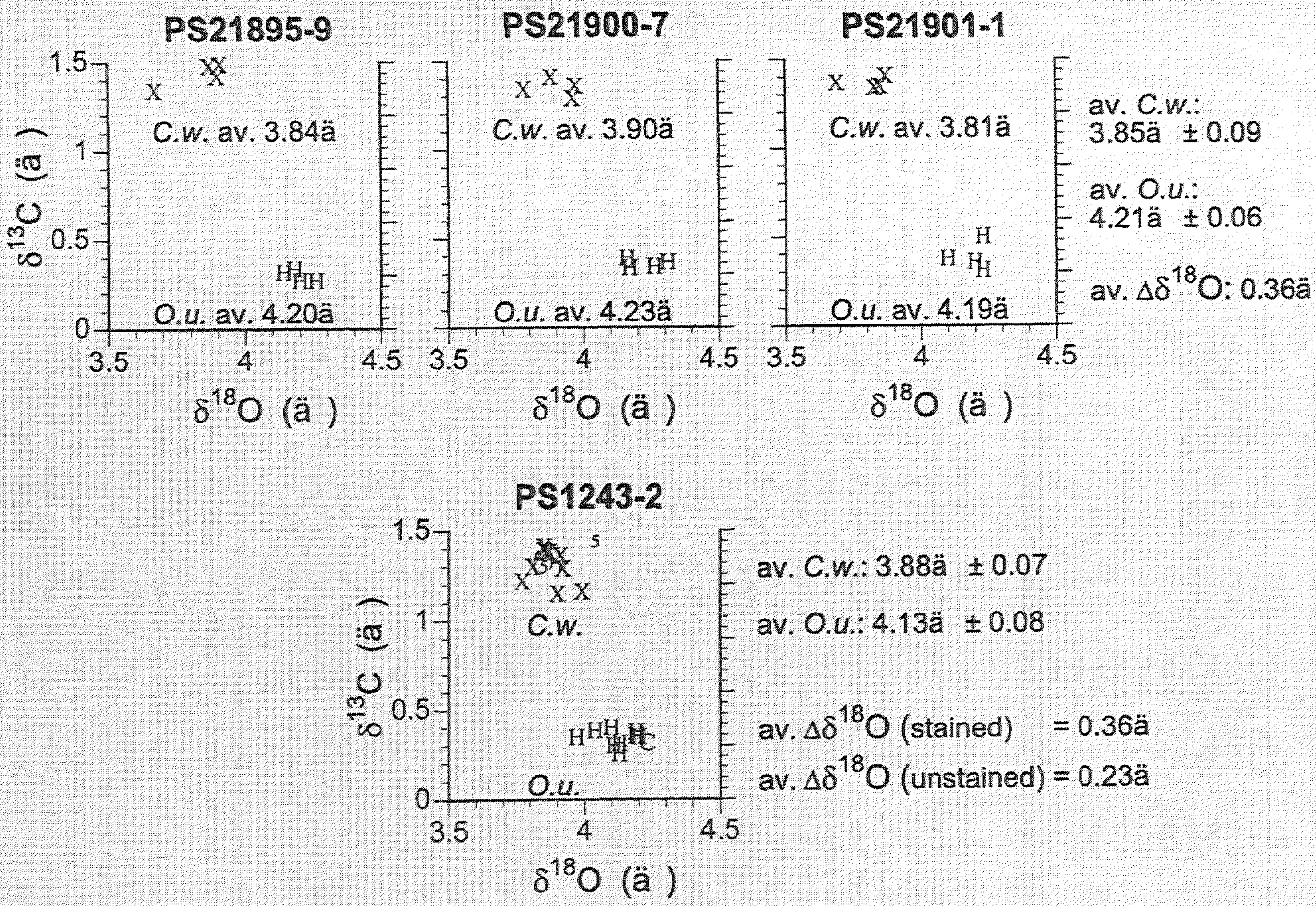

Figure 5. Benthic $\delta^{18}O$ values (uncorrected) from coretop samples of the Greenland (top) and Norwegian basins shown in $\delta^{18}O/\delta^{13}C$ space (compare with Table 1 and 2). Mean values for each species and the calculated offsets between the two species on the basis of stained (living) and unstained (dead) specimens are also shown. Light $\delta^{18}O$ values refer to *C. wuellerstorfi* (J = unstained single specimen; E = stained bulk specimens; 5 = unstained bulk specimens), heavy $\delta^{18}O$ values refer to *O. umbonatus* (H = unstained bulk specimens; C = stained bulk specimens).

Table 2. Benthic foraminiferal $\delta^{18}O$ from coretop samples (C.w. = *Cibicidoides wuellerstorfi*; O.u. = *Oridorsalis umbonatus*).

Core	PS1243-2				PS1895-9		PS1900-7		PS1901-1	
Species	O.u.	C.w.	C.w.	C.w.	O.u.	C.w.	O.u.	C.w.	O.u.	C.w.
	4.19^	3.88*	3.92°	4.04^	4.18^	3.90^	4.31^	3.78^	4.10^	3.85^
	4.10^	3.82*	3.91°	3.85^	4.14^	3.91^	4.17^	3.97^	4.23^	3.69^
	3.97^	3.91*	3.90°		4.26^	3.87^	4.16^	3.88^	4.20^	3.87^
	4.04^		3.85°		4.20^	3.67^	4.26^	3.96^	4.23^	3.83^
	4.20^		3.81°							
	4.19^		3.87°							
	4.13^		3.86°							
	4.11^		3.77°							
	4.13^		3.99°							
	4.23*		3.85°							

* stained bulk specimens; ° unstained single specimen; ^ unstained bulk specimens

The $\delta^{18}O$ difference between MIS 2 and 6 in core PS1243 is relatively large (Figure 7). In a previous work on the LGM in the eastern Nordic seas, two slightly higher $\delta^{18}O$ values were measured for *O. umbonatus* - a mean of 4.72 ‰ when corrected into the *C. wuellerstorfi* scale [*Vogelsang*, 1990]. The actual LGM value in core PS1243 may be obscured due to low sedimentation rates at this time [*Bauch et al.*, 2001] giving some bias to bioturbational admixture of specimens having lighter $\delta^{18}O$ values. Taking the mean LGM value of ~4.72 ‰ found by *Vogelsang* [1990] for *O. umbonatus* to be more realistic, then there is only a minor difference in $\delta^{18}O$ between MIS 6 and 2 in our record. Such a small difference between the two glacial intervals is in good agreement with other estimates [*McManus et al.*, 1999; *Rohling et al.*, 1998; *Shackleton*, 1987].

Using the maximum value of 4.72 ‰ as a reference and a total change of ~1.25 ‰ in $\delta^{18}O$ that was attributed to the global ice volume at the LGM [*Fairbanks*, 1989], then global sea level in MIS 12 must have been at least 30 % lower than during the LGM. But the 1.25 ‰ of global ice volume effect for the LGM relative to the Holocene has been recently challenged by $\delta^{18}O$ records from interstitial porewaters. These new data indicate a total change in $\delta^{18}O$ of only 0.8-1 ‰ [*Schrag et al.*, 1996; 1997]. Assuming that the estimate of at least 120 m of global sea-level fall at the LGM is reliable [e.g., *Shackleton*, 2000], the sea level during MIS 12 relative to the LGM would have been even lower if the results of *Schrag* were used as basis for the calculation.

Estimating relative global ice volumes based on benthic foraminiferal $\delta^{18}O$ strongly depends on precise assumptions of past BWTs at the particular site [*Shackleton*, 2000]. The similar benthic $\delta^{18}O$ values noted for MIS 6 and 2 at Site 980 and PS1243 (figs. 1, 7) are evidence that the BWT between the two regions was homogeneous during these two glacial maxima, i.e., the glacial BWT at Site 980 was about 4°C colder than today. Because our estimate of about 0.9 ‰ change in the $\delta^{18}O$ signature of *O. umbonatus* since the LGM is in good accordance with the change in porewater $\delta^{18}O$ implies that the BWT at site PS1243 during the LGM was not much different from today. However, this does not rule out the possibility of warmer bottom waters in the Nordic seas during cold time intervals other than the actual glacial maxima [*Bauch and Bauch*, 2001].

5.2. *Glacial-to-Interglacial Transitions*

All three terminations investigated document a rather uniform evolution. Following each glacial maximum, the benthic $\delta^{18}O$ values are light in comparison with the ensuing peak interglacial phase (Figure 7). It has been shown that these anomalous isotopic events occurred simultaneously with Atlantic-type planktic and benthic foraminiferal indicator species [*Rasmussen et al.*, 1996; *Bauch et al.*, 2000a; *Bauch et al.*, 2001]. The origin of these light benthic $\delta^{18}O$ spikes, which are so pronounced in the early terminations but which also appear during times of less intensive glaciations, is still controversial. Some authors relate them solely to enhanced brine ejection upon sea-ice formation [e.g., *Veum et al.*, 1992; *Dokken and Jansen*, 1999]. Others have linked them to subsurface inflow of more saline North Atlantic water [*Rasmussen et al.*, 1996] that could have also warmed the bottom waters [*Bauch et al.*, 2001]. Because meltwater-diluted surface waters inhibits the deep convectional processes required to produce very cold bottom waters in the Nordic seas, deep water warming appears to be a likely mechanism that could have contributed to the formation of these light benthic $\delta^{18}O$ anomalies [*Bauch and Bauch*, 2001].

The IRD data from core PS1243 show that massive inputs usually occurred during times of glacial maxima, indicated by highest benthic $\delta^{18}O$ values, and during the deglaciation (Figure 7). Although the IRD records in Termination 1 and 2 show similar trends, with highest values always occurring during the glacial maximum and the early transitional phase, Termination 5 has a massive IRD spike only during the

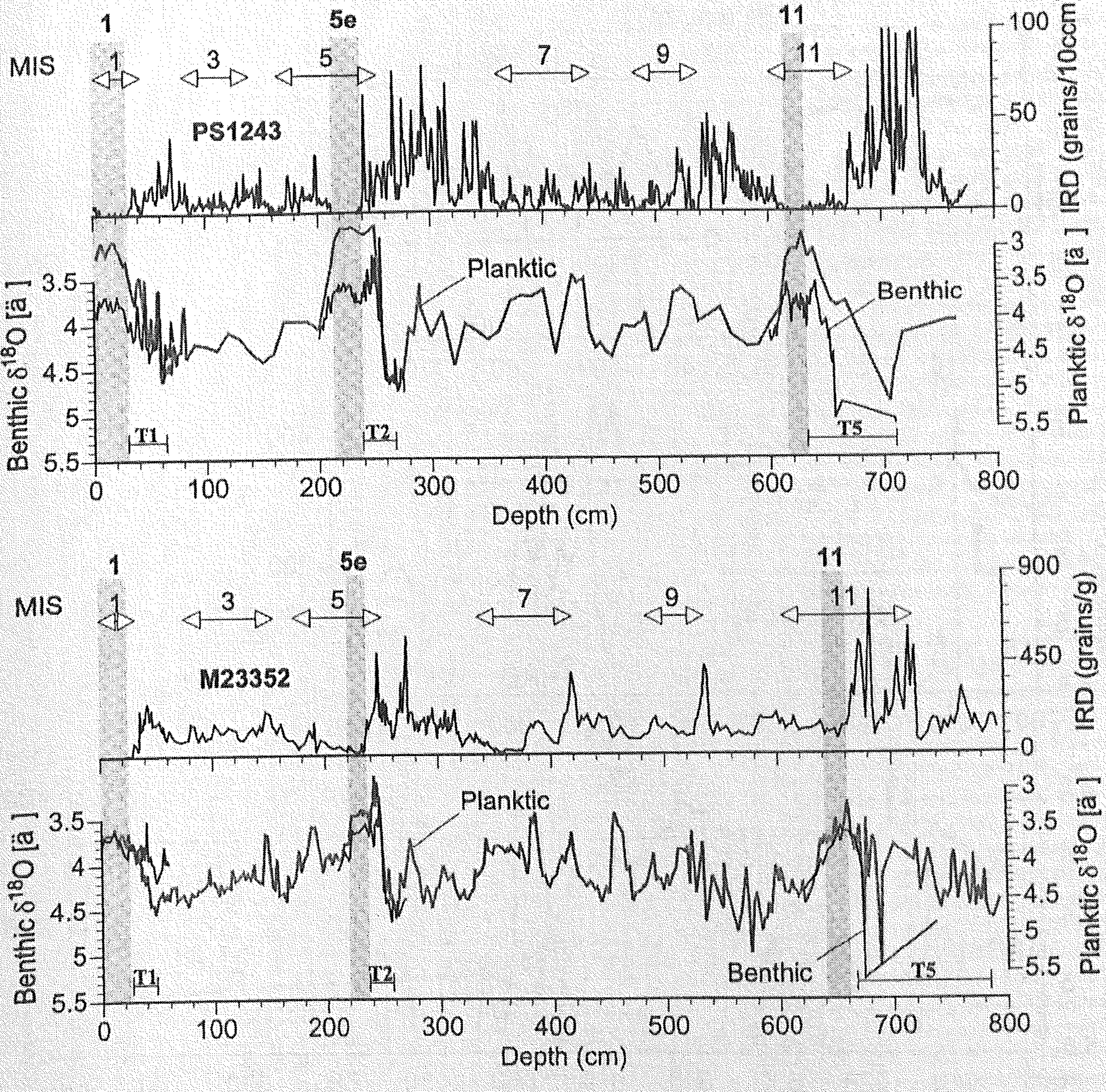

Figure 6. Downcore benthic $\delta^{18}O$ records of cores PS1243 and M23352, using mainly values from *O. umbonatus* (black line), in comparison with IRD and planktic $\delta^{18}O$ (grey line). Note that the upper part of MIS 5e in core PS1243 (above 237 cm) is based on *C. wuellerstorfi* values. All other benthic $\delta^{18}O$ values are illustrated on the *C. wuellerstorfi* scale, i.e., 0.36 ‰ was subtracted from the original values of *O. umbonatus* (see also Figure 5).

early phase of deglaciation (centered around 659 cm core depth) while most of the transition has significantly low but consistent IRD content (between 650-637 cm core depth). This portion of the core contains very high concentrations (specimens/g) of planktic foraminifera composed mainly of polar *N. pachyderma* sin. Warm water indicative subpolar species were only found within the IRD-free section of MIS11 [*Bauch et al.*, 2000b]. This is in accordance with other data showing that the appearance of increased amounts of subpolar species in the Nordic seas always coincides with times when IRD is absent, such as found during the Holocene and MIS 5e [*Bauch et al.*, 1996].

The finding of low but consistent IRD during substantial parts of Termination 5, before proper interglacial conditions

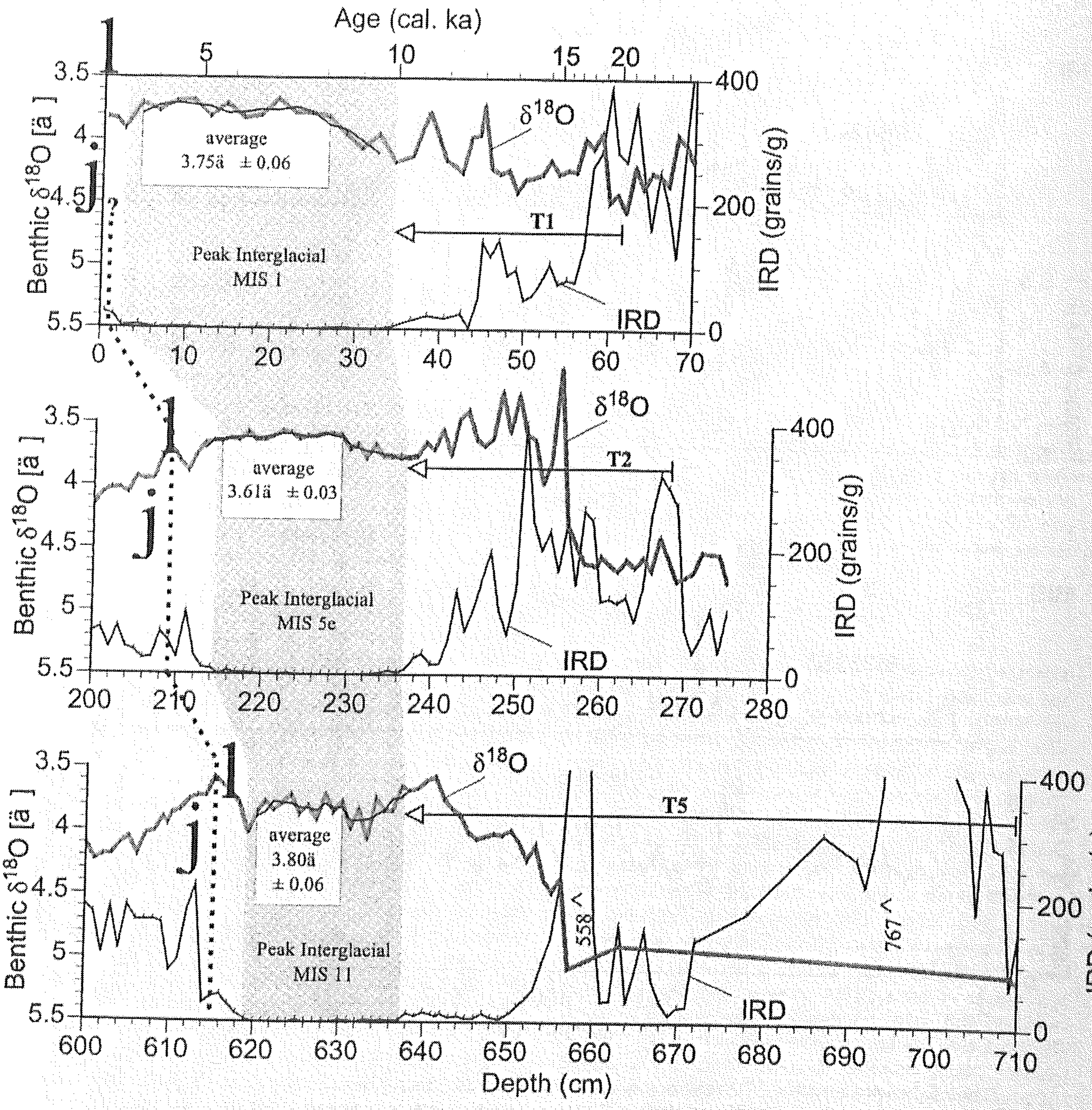

Figure 7. Detailed comparison of MIS 11, 5e, and 1 on the basis of IRD (grains > 250μm per gram of total sediment) and benthic $\delta^{18}O$ records. Thin black lines superimposed on the $\delta^{18}O$ records represent smoothed $\delta^{18}O$ values. Average peak interglacial values shown in the white boxes were calculated using the original data points framed by each underlying box. Arrows mark significant changes in deep water $\delta^{18}O$ during early interglacial-to-glacial transitions.

were reached in the Nordic seas, bears important implications for the interpretation of the climatic evolution of MIS 11. The presence of the IRD indicates that the surface water above core PS1243 experienced meltwater stratification that likely cut off vertical convection. This was a time when regions further south in the North Atlantic were already ice-free and in a state of peak interglacial conditions [e.g., see *Oppo et al.* 1998].

5.3. Peak Interglacial Ice Volume and Sea Level

In all three core sections, the intervals immediately following cessation of IRD deposition are always characterized by a trough in the benthic $\delta^{18}O$ curve followed by a further decrease (Figs. 4, 7). Only during the later phase of each peak interglacial interval did benthic $\delta^{18}O$ values become relatively steady, probably due to the length of time it took for the global ocean to become well mixed [*Lambeck and Nakado*, 1992]. Directly comparing the benthic $\delta^{18}O$ values from these peak interglacial sections indicates that the global ice volume was highest during MIS 5e (Figure 7). For MIS 5e, a sea level higher than the Holocene has been described on the basis of many other records [e.g., *Chappell and Skackleton*, 1987; *Chen et al.*; 1991, *Gallup et al.*, 1994].

The mean $\delta^{18}O$ value calculated for MIS 11 indicates the lowest peak interglacial sea level of all three investigated intervals. However, BWT during MIS 11 may have been as low as today in the deep Greenland Basin at the site of PS1243 if the main convectional gyre had shifted or expanded southward into the Norwegian Basin. This may be concluded from the high proportions of the *N. pachyderma* sin. (>85 %) that persisted at this core site throughout the IRD-free section [*Bauch et al.*, 2000b]. That surface water conditions in the southern Nordic seas during MIS 11 differed significantly from those during MIS 5e and 1 can be also inferred from the persistent occurrence of IRD in core M23352.

Although the determination of past BWTs is crucial for a precise assessment of past global ice volumes, we conclude from our benthic $\delta^{18}O$ and IRD records that the global ice volume during the peak interglacial conditions of MIS 11 was no smaller than in MIS 1. This is in agreement with interpretations based on benthic isotope data from the North Atlantic [*McManus et al.*, 1999] but contradicts other reports indicating for MIS 11 a much smaller ice volume and a higher sea level, respectively [e.g., *Scherer et al.*, 1998; *Rohling et al.*, 1998; *Hearty et al.*, 1999]. It is important to note that the duration of MIS 11 was considerably longer than for any of the later peak interglaciations [*Imbrie et al.*, 1984], and the preceding glacial sea level in MIS 12 was probably the lowest of all later glaciations. Both of these factors combined should have had a particularly strong effect on shoreline deposition during MIS 11, in both isostatically stable and unstable regions. Moreover, it implicates that a direct comparison of the sea-level highstand in MIS 11 relative to the later interglacial highstands is difficult when only based on shoreline evidence [*Lambeck and Nakado*, 1992].

5.4. Interglacial-to-Glacial Transitions

The last interglacial climatic cycle (MIS 5e) is often taken as a case study to learn more about the Holocene and its possible fate in the near future. A number of studies in the northern North Atlantic have recognized a good similarity of these two interglaciations in terms of surface and deep water properties [e.g., *Kellogg*, 1980; *Labeyrie et al.*, 1987; *Bauch et al.*, 1996; *Fronval et al.*, 1998]. The transition from MIS 5e towards MIS 5d, when climate conditions switched back to a glacial-like mode with increasing global ice volume, is of particular interest [*Shackleton*, 1987; *Stirling et al.*, 1998]. Climate cooling at the end of a peak interglacial period should be noticeable first in proxy records from the polar regions due to the sensitivity of the environmental system at such extreme latitudes [*Dickson*, 1999]. There is evidence from the western Nordic seas that the surface water temperatures in MIS 1 started to deteriorate about 6 cal. ka ago [*Koç et al.*, 1993; *Bauch et al.*, 1996] when glaciers in Norway reappeared and grew after they had vanished during the Holocene hypsithermal period [*Nesje and Kvamme*, 1991]. During MIS 5e, surface conditions in the Nordic seas also began to cool already during the peak interglacial phase. IRD recurred somewhat later, at approximately the same time when benthic $\delta^{18}O$ values started to increase but significantly earlier than had been noted in sediment records from further south in the North Atlantic [*Bauch et al.*, 2000a].

In all interglacial-glacial transitions of core PS1243, including the most recent time, we observe this systematic relation between IRD recurrence and increasing benthic $\delta^{18}O$ values (Figure 7). The increase in benthic $\delta^{18}O$ is always followed by a notable isotopic depletion. In MIS 11, this depletion is as high as 0.5 ‰, in MIS 5e 0.2 ‰, and in MIS 1 about 0.1 ‰. While the increase in benthic $\delta^{18}O$ is probably also associated with growing global ice volume, it would be hard to reconcile the $\delta^{18}O$ depletions that followed with a reversal in this process. Recent data from the MIS 5e/5d transition show abrupt cooling events that may have been caused by a weakening of the thermohaline circulation in the North Atlantic [*Adkins et al.*, 1997; *Chapman and Shackleton*, 1999]. The benthic $\delta^{18}O$ depletions observed by us during early interglacial-glacial transitions could directly result from reduced vertical convection in the Nordic seas. This process was then probably also related with an increase in BWT [cf. *Budeus et al.*, 1998]. That the $\delta^{18}O$ depletions were caused by brine ejection [cf. *Dokken and Jansen*, 1999] seems rather unlikely at such an early stage of glacial inception with still relatively low global ice volume. The same would apply to processes that involve an interaction between ice shelves and inflowing surface water masses [cf. *Bauch and Bauch*, 2001] because of a general lack of available ice shelves.

For the fate of the present climate status it is intriguing to speculate on the causes of the recurrence of IRD in the topmost part of core PS1243. Today, this site is usually sea-ice free. But during the latest Holocene cold event, the Little Ice Age, the sea ice margin in the Nordic seas extended much further to the east than at present [*Lamb*, 1979]. Most likely, this eastward sea-ice expansion was also associated with icebergs because sea ice in the Nordic seas usually does not carry lithic grains of such sizes as identified by us [*Pfirman et al.*, 1990]. The $\delta^{18}O$ increase in both benthic species that began approximately 2000 years ago (Figure 4) may be taken as another piece of evidence for an impending change of the present interglacial climate system towards increasing glacial conditions.

6. SUMMARY

In this study, $\delta^{18}O$ measurements on two different benthic foraminiferal species, for which the species-dependent isotopic offset was precisely determined, were carried out using deep-sea sediment core samples from the Norwegian Basin and the Iceland Plateau (southern Nordic seas). Assuming that similar bottom water temperatures always prevailed during glacial and during interglacial climate modes respectively, the benthic $\delta^{18}O$ records have been used to interpret relative changes in global ice volume (sea level) and bottom water conditions for some major contrasting climate intervals since marine isotope stages (MIS) 12. In combination with carbonate content and iceberg-rafted debris (IRD), a clear distinction was made between several climatic modes: (1) postglacial transitions (terminations) marked by decreasing ice volumes and rising sea levels; (2) peak interglacial phases with relatively stable ice volumes and sea levels; (3) interglacial-to-glacial transitions characterized by a return to glacial conditions with growth of continental ice.

Three dominant interglaciations have been identified in the Nordic seas, MIS 11, 5e, and 1. MIS 12 is recognized as the glaciation when global ice volume was largest (or conversely, when sea level was lowest) relative to the last glacial maximum (LGM). In comparison, ice volume was probably only slightly larger in MIS 6 than in the LGM.

The glacial-to-interglacial transitions are usually characterized by significant amounts of IRD in their early parts, gradually decreasing during the later deglacial phases. However, the MIS12/11 transition (Termination 5) differs from the other two (Termination 2 and 1) in showing a steep decrease in IRD during the early part of the deglaciation when benthic $\delta^{18}O$ values were still at a glacial-like level. Most of the remaining part of Termination 5 reveal small but noticeable amounts of IRD implying that surface water freshening from melting icebergs affected the water mass properties of the Nordic seas for a much longer time than further south in the North Atlantic.

In all three interglacial core sections investigated from the Norwegian Basin IRD eventually vanished from the records while benthic $\delta^{18}O$ values gradually reached a relatively steady level. These intervals without IRD are defined by us as peak interglacial intervals. These were times when global ice volume was lowest and, consequently, sea level highest.

The mean benthic $\delta^{18}O$ values during these peak interglacial phases identify the last interglaciation, MIS 5e, as a time with lowest global ice volume. In comparison, sea levels during peak MIS 1 and MIS 11 appear to have been of similar magnitude. On the Iceland Plateau, close to the modern Polar Front, IRD remained a significant feature during peak MIS 11. This implies that surface water conditions during MIS 11 were significantly colder than during MIS 5e and 1.

Based on a detailed examination, the transitions following the peak interglacial phases in MIS 11 and 5e were used as analogue to evaluate the modern status of climate conditions in the Norwegian Basin. In both of these interglacial-to-glacial transitions an early increase in benthic $\delta^{18}O$ is observed concomitant with a reappearance of IRD. Because the topmost samples in the same core (i.e., past 2 cal. ka) also reveal an increasing trend in benthic $\delta^{18}O$ values and a recurrence of IRD since about 1.5 cal. ka, it is suggested that the water mass conditions in the Nordic seas, which already show a consistent surface cooling trend for the last 6 cal. ka, have come to a critical state during more recent time.

Acknowledgments. Many thanks to all scientists and crews onboard RV *Polarstern* and RV *Meteor* who were responsible for obtaining the two studied cores. Jerry McManus kindly provided the isotope record from Site 980 and Bill Hay improved the English text. We are grateful to the isotope team from the Leibniz Laboratory (Kiel) for their assistance. Three anonymous reviewers are thanked for their comments on the manuscript.

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