

Late Holocene surface ocean conditions of the Norwegian Sea (Vøring Plateau)

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[1] Late Holocene sea surface ocean conditions of the eastern Norwegian Sea (Vøring Plateau) are inferred from planktic stable isotopes and planktic foraminiferal assemblage changes in cores JM97-948/2A and MD95-2011 (66.97°N, 7.64°E). Strong covariance between the planktic stable oxygen isotopic record and abundance changes of *N. pachyderma* (sin) show that major changes in surface ocean conditions are reflected both in the geochemical composition of the foraminiferal tests as well as in the composition of the foraminiferal fauna. Surface ocean conditions warmer than present were common during the past 3000 years. During the so-called Medieval Warm Period, surface conditions were highly variable with marked changes in sea surface temperature. The warmest sea surface temperatures during this period occurred between 800 and 550 years BP (0 BP = AD 2000). Climatic deterioration, recorded as decreases in sea surface temperature, occurred at about 2750, 1550, 400, and 100 years BP. The cooling events at about 2750 and 1550 years BP appear to correlate with increases in ice-rafted debris in the North Atlantic. Based on the results from JM97-948/2A and MD95-2011, the onset of the Little Ice Age cooling trend seems to have occurred around 700–600 years BP. Faunal changes indicate two cooling events during the Little Ice Age (at 400 and 100 years BP) that correspond to decreases in Fennoscandian summer temperatures and increases in ice-rafted debris in the eastern North Atlantic. **INDEX TERMS:** 3030 Marine Geology and Geophysics: Micropaleontology; 4267 Oceanography: General: Paleoclimatology; 4870 Oceanography: Biological and Chemical: Stable isotopes; **KEYWORDS:** paleoclimatology, stable isotope, Holocene, micropaleontology

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1. Introduction

[2] Quantitative information about the range of Holocene climate change and the spatial patterns of principal climatic anomalies, both on land and in the ocean, provide important insights into the behavior of the climate system under preanthropogenic conditions. By collecting and compiling paleoclimatic data from various sources we may better understand the forcing mechanisms behind the temporal and spatial patterns observed, and hopefully be able to predict how future climate will evolve.

[3] Recent investigations indicate that the surface water masses of the subpolar North Atlantic have undergone considerable variability during the Holocene [Bond *et al.*, 1997, 2001]. New sedimentological results from the north-eastern Atlantic have also documented the presence of millennial-scale variability in deep water circulation. Sediment color data [Chapman and Shackleton, 2000] from core NEAP15K show a cyclic variability of Holocene climate at

millennial timescales. This variability, which is tied to changes in thermohaline circulation, seems to be coherent with atmospheric changes over Greenland [Chapman and Shackleton, 2000]. The regional cooling event at 8200 cal. years BP recorded in terrestrial [Dahl and Nesje, 1996; Nesje and Dahl, 2001], ice core [Alley *et al.*, 1997] and marine records [Klitgaard-Kristensen *et al.*, 1998] indicates that oceanic and atmospheric variability in the North Atlantic region was strongly interconnected during the Holocene [e.g., Bond *et al.*, 1993]. Recent studies also suggest that cyclicities in North Atlantic Deep Water (NADW) are driven, to some extent, by variations in solar origin [Chapman and Shackleton, 2000; Bond *et al.*, 2001].

[4] The location of the Nordic Seas within the northernmost limb of the North Atlantic Current makes it an important area for studies of Holocene climate variability. The northward flux of warm and saline surface water masses through the North Atlantic and the Nordic Seas provides northwest Europe with both heat and moisture. Previous studies of the last interglacial, the Eemian, have documented rapid changes in sea surface temperature in the North Atlantic and the Norwegian Sea [Cortijo *et al.*, 1994], which were particularly pronounced in the Norwegian Sea [Fronval and Jansen, 1996].

[5] Even though a link between thermohaline circulation and atmospheric changes in the subpolar North Atlantic is strongly suggested by previous studies, substantial efforts are needed to establish exactly how the

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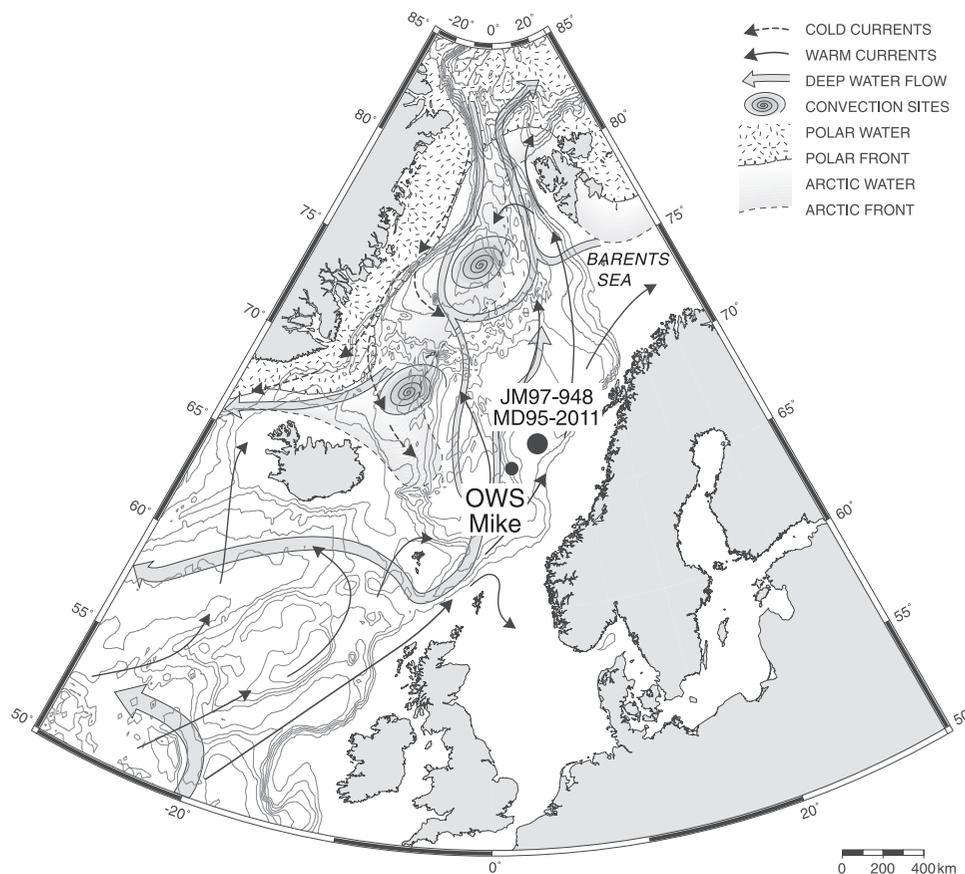


Figure 1. Map showing the location of cores JM97-948/2A and MD95-2011 and a schematic view of the circulation pattern in the Nordic Seas. The position of Ocean Weather Station (OWS) Mike is also shown.

ocean-atmosphere linkage works. Both highly resolved terrestrial and marine records are needed to better understand the ocean-atmosphere system and its different forcings, thresholds, feedbacks, and their effect on climate. For northwestern Europe, changes between marked climatic events, such as the Medieval Warm Period (MWP) and the Little Ice Age (LIA), have been documented in numerous different paleoclimatic archives. However, there is still a need to establish how the ocean-atmospheric system works during periods of such changes.

[6] In this paper we focus on late Holocene climatic change as recorded in marine sediments off Norway. The overall aim is to reconstruct changes in surface ocean conditions in the Norwegian Sea (Vøring Plateau) over the past 3000 years based on geochemical and faunal analysis of planktic foraminifers, and to relate these changes to other paleoclimatic archives, such as glacier fluctuations, tree ring chronologies, ice cores and lake sediment data.

2. Material and Methods

2.1. Core Location and Oceanographic Setting

[7] Marine sediment cores were recovered from the Vøring Plateau (Figure 1) during IMAGES campaign MD101 in 1995 onboard R/V *Marion Dufresne*, and in 1997 during

a cruise with F/F *Jan Mayen*. During MD101, core MD95-2011 (66.97°N, 7.64°E, 1048 meter water depth, and a length of 17.49 meter) was recovered using a giant piston corer (CALYPSO). Due to overpenetration of the giant piston corer, a box core, JM97-948/2A, was later used at the same position to sample the surface sediments lost during the piston coring.

[8] The current system in the Nordic Seas (Greenland, Iceland and Norwegian Seas) consists of three main currents: the warm Norwegian Atlantic Current (the northern continuation of the Atlantic Current), the cold East Greenland Current, and the East Icelandic Current [Mork and Blindheim, 2000]. There are two limbs of the Norwegian Atlantic Current, one at the shelf edge and the other further west. These limbs merge into a single branch west of Lofoten and Vesterålen [Mork and Blindheim, 2000]. The northward inflow of warm and saline Atlantic water to the Nordic Seas has great impact on climate in this region where the typical temperature of Atlantic Water is 9.5–10.5°C [Hansen and Østerhus, 2000]. Atlantic waters enter the Norwegian Sea through the Faeroe-Shetland Channel as well as between the Faroes and Iceland [Blindheim et al., 2000]. The southward flowing cold East Greenland Current brings both surface and deeper water masses from the Arctic Ocean. The East Icelandic Current carries some of the waters

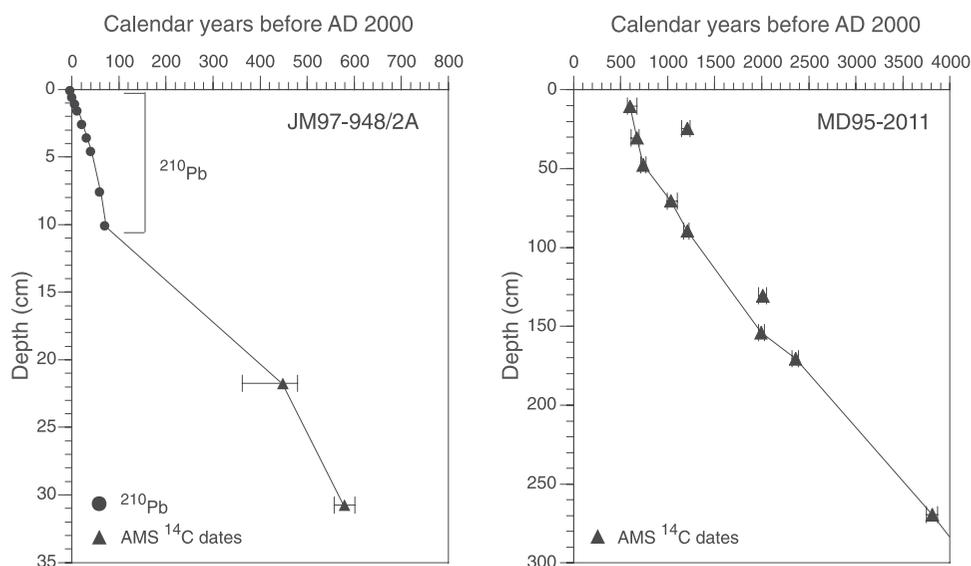


Figure 2. Age model for cores JM97-948/2A and MD95-2011. The age model for box core JM97-948/2A is based on a combination of ^{210}Pb and AMS ^{14}C chronologies. The age-depth relationship for MD95-2011 is based solely on AMS ^{14}C dates. Error bars of AMS dates represent the 1σ deviation. The asymmetric distribution of the error of the AMS dates is explained by changes in the calibration curve. All AMS dates were calibrated using a reservoir correction of 400 years and the INTCAL98 marine calibration curve [Stuiver *et al.*, 1998].

of the East Greenland Current into the Iceland and Norwegian Seas. Water from the East Greenland Current also forms dense overflow water which exits the Nordic Seas through the Denmark Strait and contributes to the deep western boundary current in the North Atlantic [Blindheim *et al.*, 2000].

2.2. Stable Isotopes

[9] All samples were processed at the University of Bergen using standard low-temperature techniques. The working half of the piston core was subject to shore-based sampling every 1-centimeter down to 215 cm. The JM97-948/2A box corer was subsampled every 0.5-cm through its entire length of 30 cm.

[10] The stable isotope measurements were performed on Finnigan MAT 251 and 252 mass spectrometers at the University of Bergen. Stable oxygen and carbon isotope records were generated from analyses of the planktic species *Neogloboquadrina pachyderma* (dextral), *Neogloboquadrina pachyderma* (sinistral) (hereafter referred to as *N. pachyderma* (dex) and *N. pachyderma* (sin), respectively), and the benthic species *Cassidulina teretis* from the $>150\ \mu\text{m}$ fraction. The analytical data are reported with reference to the Vienna Pee Dee belemnite (VPDB) standard via NBS19 [Coplen, 1996] and an in-house laboratory standard. The analytical errors for measurements of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ of carbonate standards are 0.07‰ and 0.06‰, respectively.

2.3. ^{14}C Accelerator Mass Spectrometry and ^{210}Pb Dating

[11] We have constructed an age-depth model for the past 3000 years by integrating results from ^{210}Pb -dating and accelerator mass spectrometer (AMS) radiocarbon dating techniques (Figure 2, Table 1). The ^{210}Pb -dating was

Table 1. Age Control Points of Cores JM97-948/2A and MD95-2011 Based on ^{210}Pb and AMS Dating of the Planktic Foraminifer *N. pachyderma* (Dextral)

Core	Core Depth, cm	^{14}C AMS Age years BP $\pm 1\sigma$	Calibrated Age, ^{a,b} cal years BP, 0 = AD 2000	Age Range, $\pm 1\sigma$	Comment/AMS Laboratory Number ^c
JM97-948/2A	0.25		5		^{210}Pb dated
JM97-948/2A	0.75		9		^{210}Pb dated
JM97-948/2A	1.25		15		^{210}Pb dated
JM97-948/2A	1.75		20		^{210}Pb dated
JM97-948/2A	2.75		30		^{210}Pb dated
JM97-948/2A	3.75		40		^{210}Pb dated
JM97-948/2A	4.75		49		^{210}Pb dated
JM97-948/2A	7.75		68		^{210}Pb dated
JM97-948/2A	10.25		79		^{210}Pb dated
JM97-948/2A	21.75	735 \pm 40	448	362–480	KIA6285
JM97-948/2A	30.75	940 \pm 40	549	558–601	KIA4800
MD95-2011	10.5	980 \pm 60	601	568–675	GifA96471
MD95-2011	24.5	1590 \pm 40	1209	1150–1237	KIA5600 ^d
MD95-2011	30.5	1040 \pm 40	675	614–695	KIA3925
MD95-2011	47.5	1160 \pm 30	739	716–772	KIA5601
MD95-2011	70.5	1460 \pm 50	1037	993–1103	KIA3926
MD95-2011	89.5	1590 \pm 30	1209	1169–1227	KIA6286
MD95-2011	130.5	2350 \pm 40	2012	1965–2052	KIA3927 ^d
MD95-2011	154.0	2335 \pm 25	1992	1965–2047	KIA6287
MD95-2011	170.5	2620 \pm 60	2359	2322–2392	GifA96472
MD95-2011	269.5	3820 \pm 35	3813	3748–3870	KIA10011

^aRadiocarbon ages were calibrated using CALIB 4.3 in conjunction with INTCAL98 [Stuiver *et al.*, 1998].

^bA 400-year correction for the marine reservoir effect was applied.

^cKIA, Leibniz-Labor for Radiometric Dating and Isotope Research, Kiel; GifA, Centre des Faibles Radioactivités, Gif sur Yvette.

^dSamples not used in the age-depth model.

performed on the topmost sediments of core JM97-948/2A (Figure 2) at the Gamma Dating Center, Risø National Laboratory, Denmark.

[12] Twelve AMS ^{14}C dates were generated on monospecific samples of *N. pachyderma* (dex) (Figure 2). As two of these AMS ^{14}C dates, both in core MD95-2011, deviated considerably from the remaining ten AMS dates, we chose not to include these dates into the age-depth model for core MD95-2011. The two excluded AMS ^{14}C dates gave inverted ages, which are probably related to re-sedimentation. The AMS dating was carried out at the Leibniz Laboratory for Isotope Research at Kiel University. Descriptions of analytical procedures are found in the work of Nadeau *et al.* [1997, 1998, and references therein]. Radiocarbon ages were calibrated using CALIB 4.3 in conjunction with the INTCAL98 marine calibration curve [Stuiver *et al.*, 1998]. A 400-year correction for the marine reservoir age was used for the calibration. All ages are reported as calendar years before AD 2000 throughout this paper unless otherwise stated.

[13] The mean sedimentation rates in JM97-948/2A and MD95-2011 are 0.146 and 0.119 cm/year, respectively. With a sample spacing of 0.5 cm in core JM97-948/2A and 1.0 cm in core MD95-2011, the resulting mean temporal resolution is 3.4 years and 8.4 years, respectively.

2.4. Census Counts and Sea Surface Temperature Estimates

[14] The relative abundances of planktic foraminifers are based on census counts of splits of the $>150\ \mu\text{m}$ fraction containing approximately 300 specimens. In this paper we have used the modern analog technique (MAT) (in conjunction with the ATL916 core top database; X. U. Pflaumann, personal communication, 2000) to estimate sea surface temperatures (SSTs) in the Norwegian Sea. In this area the diversity of the foraminiferal fauna is low and the core tops somewhat unevenly distributed. Even though distance weighted MAT estimates are statistically superior, we have chosen not to apply geographical weighting. The rationale behind this is that since environmental changes lead to faunal migrations [González-Donoso and Linares, 1998], weighting based on geographical distance does not necessarily lead to more realistic SST estimates. Stable isotope, faunal and SST data are available from the World Data Center for Paleoclimatology, 325 Broadway, Boulder, Colorado. (URL, <http://www.ngdc.noaa.gov/paleo/paleo.html>; email, paleo@ngdc.noaa.gov.)

3. Results

3.1. Faunal Assemblage Changes

[15] Changes of the planktic foraminifer fauna of cores JM97-948/2A and MD95-2011 over the past 3000 years are shown in Figure 3. The planktic foraminifer fauna in the studied cores is characterized by polar, subpolar and transitional species. By far the most dominant species is *N. pachyderma* (dex), which make up 40–70% of the total fauna throughout the studied time span. *Globorotalia quinqueloba* is present in relative high abundance throughout the studied period, seemingly without any particular correlation with either *N. pachyderma* (dex) or *N. pachyderma* (sin). The abundances of *Globigerinita glutinata* and *Glo-*

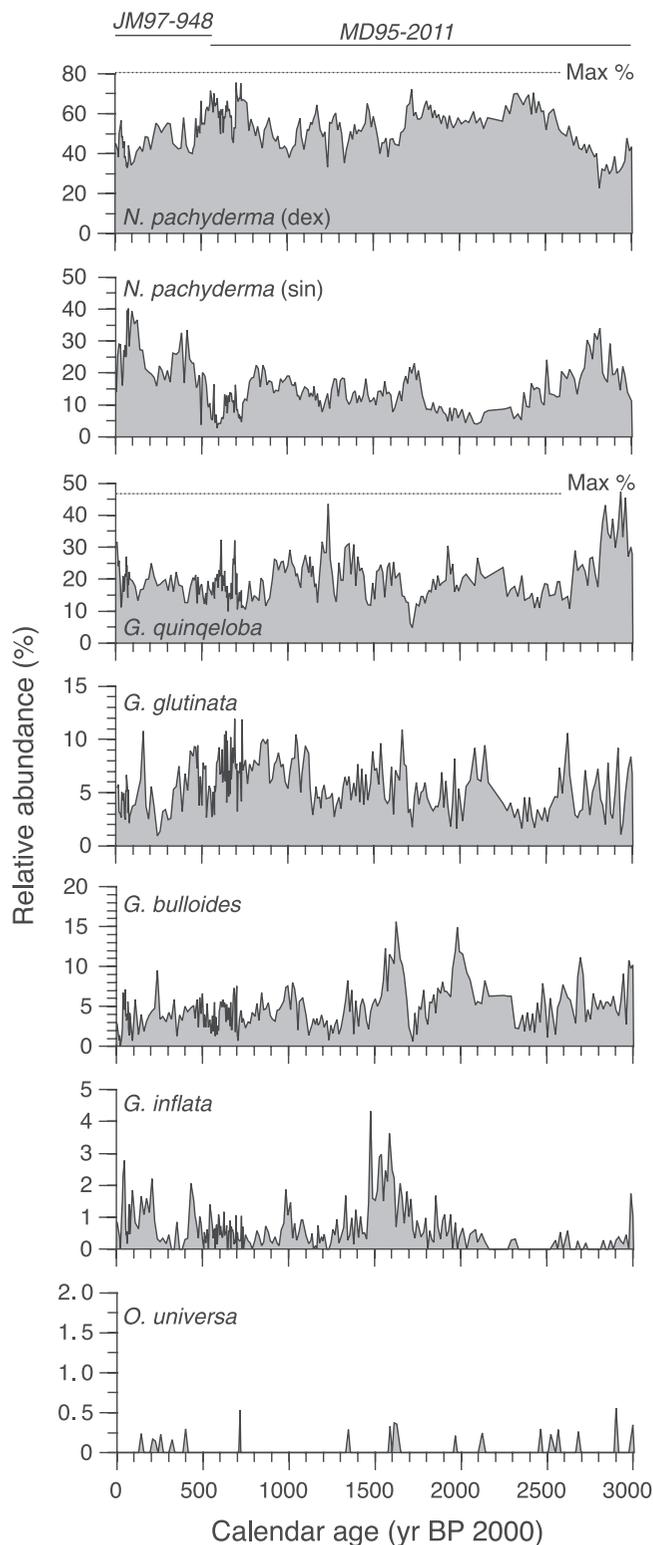


Figure 3. Temporal changes in the relative abundance of selected (major) planktic foraminifer species in cores JM97-948/2A and MD95-2011 over the past 3000 years. The distribution of data from the two spliced cores is shown at the top. Stippled lines show the maximum abundance of *N. pachyderma* (dex) and *G. quinqueloba* in the modern core top database.

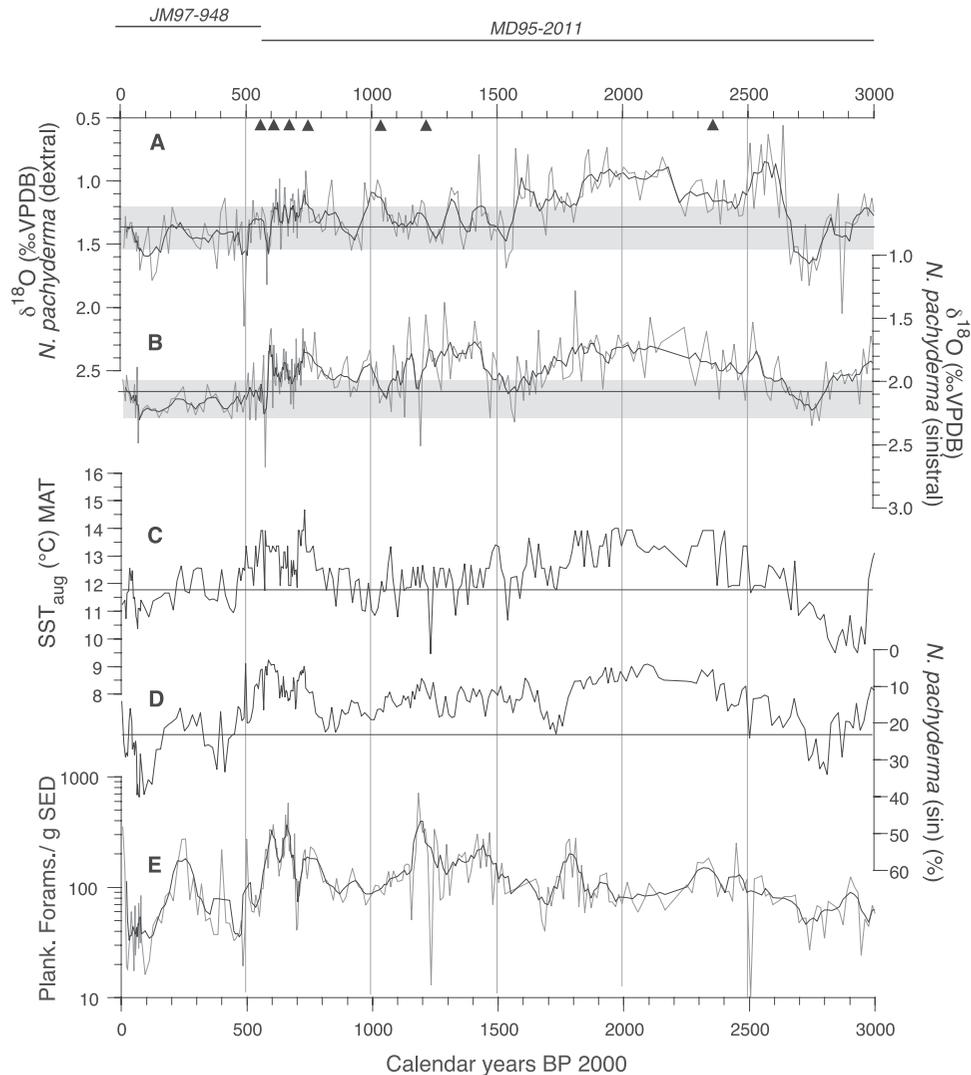


Figure 4. Temporal changes in surface ocean variability at the Vøring Plateau over the past 3000 years. The distribution of data from the two spliced cores is shown at the top. Comparisons between (a and b) changes in the oxygen isotope composition of *N. pachyderma* (dex) and *N. pachyderma* (sin), (c) August SST, (d) relative abundance of *N. pachyderma* (sin) (note the inverted scale), and absolute abundance of planktic foraminifers (>150 μm) per gram dry bulk sediment (smoothed by a 3-point moving average). The absolute foraminiferal abundance data is plotted on a logarithmic scale to emphasize the variability at low abundances. Figures 4a, 4b, and 4e show raw data (gray line) and a 5-point moving average (black line) of the raw data. Shaded bars denote typical core top oxygen isotope values of *N. pachyderma* (dex) and *N. pachyderma* (sin) taken from Johannessen [1992]. Thin, solid lines across the shaded bars represent average $\delta^{18}\text{O}$ values for the past 50 years in JM97-948/2A. Filled triangles denote radiocarbon age control points.

bigerina bulloides generally vary between 5 and 10% over the past 3 kyr on the Vøring Plateau. The subpolar species *G. bulloides* thrives in both cool high-latitude environments as well as in low-latitude upwelling regions [Hilbrecht and Thierstein, 1996]. In modern samples from the Norwegian Sea, the highest abundances of *G. bulloides* are found in the southeastern part, along the path of warm inflow of Atlantic Water [Johannessen et al., 1994]. The transitional species *G. inflata* occurs sporadically in the oldest part of the record, from 3000 to about 2000 years BP. After this period *G. inflata* accounts for between 5 and 10% of the assem-

blage. Only scattered occurrences of *Orbulina universa* are recorded throughout the studied time interval.

[16] There is a strong correlation between major features in the MAT-derived SST and percentage *N. pachyderma* (sin) records, as well as temporal changes in the absolute abundance of planktic foraminifers (>150 μm) (Figure 4). Visual examination of the planktic foraminifers shows that dissolution of planktic foraminifers is low throughout the studied 3000 years. Since there is no visible evidence of down core changes in the degree of dissolution, at this relatively shallow site, the recorded changes in foraminifer

abundance can be interpreted to reflect changes in productivity related to SST. This is particularly evident after 1000 years BP where decreased foraminifer production is clearly linked to cooler SSTs.

[17] The degree to which individual species exhibit a strong correlation to SST varies between species and also within an individual species' temperature range. When working with a relatively narrow temperature range such as the expected range of SST variability in the late Holocene of the Vøring Plateau, the scatter associated with the relationship between abundance and temperature becomes increasingly important. *N. pachyderma* (sin) appear to carry the most unequivocal relationship between abundance and temperature, at least when considering the temperature interval 5.5–15°C. This makes the abundance variations of this species particularly interesting as a proxy for temperature change in the eastern Norwegian Sea. Furthermore, the successful correlation between records of relative abundances of *N. pachyderma* (sin) from sites in both the North Atlantic and in the Norwegian Sea [e.g., Bond et al., 1993; McManus et al., 1994; Fronval et al., 1995] and the Greenland ice core records, justify the use of the relative abundance of *N. pachyderma* (sin) as a temperature proxy.

3.2. The *N. pachyderma* Stable Isotope Record

[18] The stable oxygen isotope record of JM97-948/2A and MD95-2011 for the past 3000 years, based on measurements of *N. pachyderma* (dex) and *N. pachyderma* (sin), is shown in Figure 4. There is a strong correlation between major features of the two records. Both $\delta^{18}\text{O}$ records display increasing values in the oldest 250 years of the record (3000–2750 years BP). This increase amounts to about 0.4–0.5‰ in both records. This is followed by a marked $\delta^{18}\text{O}$ decrease (~ 0.8 ‰) in *N. pachyderma* (dex) between 2750 and 2600 years BP. This decrease is expressed by only a slight decrease in $\delta^{18}\text{O}$ of *N. pachyderma* (sin) and does not have any counterpart in the percentage *N. pachyderma* (sin), the SST record or the total abundance of planktic foraminifers. Some of the lightest $\delta^{18}\text{O}$ values of the entire *N. pachyderma* (dex) record were recorded at 2600 years BP, after which $\delta^{18}\text{O}$ values increase, but not to values as enriched as at 2750 years BP. The significance of the considerable changes in the $\delta^{18}\text{O}$ of *N. pachyderma* (dex) during this period is not known. Between 2600 and 1600 years BP both $\delta^{18}\text{O}$ records display values lower than modern values. A cooling trend, initiated between 1900 and 1800 years BP, ended around 1600–1500 years BP with $\delta^{18}\text{O}$ values similar to those at present for both morphotypes. Beginning about 1000–900 years BP, both morphotypes record a trend toward lower $\delta^{18}\text{O}$ values that peaked about 700–600 years BP. After this time, both records display an increase in $\delta^{18}\text{O}$. This increase most likely corresponds to the cooling associated with the Little Ice Age (LIA). At the Vøring Plateau, the LIA is recorded as two discrete cooling events centered at 400 and 100 years BP (Figure 4).

4. Discussion

4.1. Significance of the Stable Oxygen Isotopic Signal

[19] In the absence of glacial meltwater discharge during the late Holocene, the oxygen isotope signal recorded by

planktic foraminifers in North Atlantic water masses is expected to monitor changes in SST and salinity. The potential influence on the planktic $\delta^{18}\text{O}$ by temperature and salinity can be estimated by looking at instrumental data for the upper 50 m of the water column from Ocean Weather Ship Station M (66°N, 2°E) [Østerhus et al., 1996] (Figure 1). Since the measurements at Station M started in 1948, the average seasonal (August–February) temperature range is 2.1°C. The average seasonal sea surface salinity range at 50m over the same period have is only 0.14‰. Based on the slope of the North Atlantic mixing line (0.6‰ $\delta^{18}\text{O}_w$ per practical salinity unit (PSU)), a 0.14‰ average seasonal salinity range corresponds to a 0.2‰ change in $\delta^{18}\text{O}_w$. However, due to the presence of freshwater sources in the Nordic Seas with less negative oxygen isotopic composition relative to the North Atlantic freshwater end-member, the Nordic Seas mixing line gives a gradient of only 0.22‰ $\delta^{18}\text{O}_w$ per PSU [Østbø, 2000]. In this context, a 0.14‰ average seasonal salinity range at Station M corresponds to a 0.05‰ change in $\delta^{18}\text{O}_w$. Hence we conclude that changes in sea surface temperature are likely to be responsible for most of the variability of the oxygen isotope record from the Vøring Plateau. This conclusion is also supported by the strong similarity between the planktic oxygen isotope records and the MAT-SST record from this locality.

[20] The location of the cores discussed in this study implies that the recorded isotopic signal is derived from changing properties of North Atlantic Waters. However, eastward migration and/or enhanced frontal mixing of the Arctic Front could also introduce cooler waters to the Vøring Plateau. The different surface water masses in the Nordic Sea can be characterized by their $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ signature recorded by planktic foraminifers [Johannessen, 1992; Johannessen et al., 1994]. For the past 3000 years, the $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values of both *N. pachyderma* (sin) and *N. pachyderma* (dex) compare well with typical values of North Atlantic Water (Figure 5).

4.2. Surface Ocean Variability at the Vøring Plateau 3000–1200 Years BP

[21] Foraminiferal summer SST estimates for the past 3000 years BP are shown in Figure 4. Modern August SSTs at the location of JM97-948/MD95-2011 are about 11.4–11.6°C. Compared with modern August temperatures, the Vøring Plateau experienced August SSTs warmer than at present during most of the past 3000 years. Comparisons between modern (i.e., core top sediments spanning the last 50 years) values and past changes in the other individual proxies show that SSTs warmer than present were common throughout the past 3000 years BP (Figure 4).

[22] The combined isotope and faunal records in Figure 4 indicate a cold event around 2750 years BP, followed by a warming trend between 2750 and 2500 years BP. Glacier advances equivalent in age to the inferred cold event at 2750 years BP are reported from northern Sweden [e.g., Denton and Karlén, 1973; Karlén et al., 1995], and from southern Norway [e.g., Dahl and Nesje, 1994]. The period of Scandinavian glacier expansion thus correlates well with the inferred temperature changes in the south-eastern

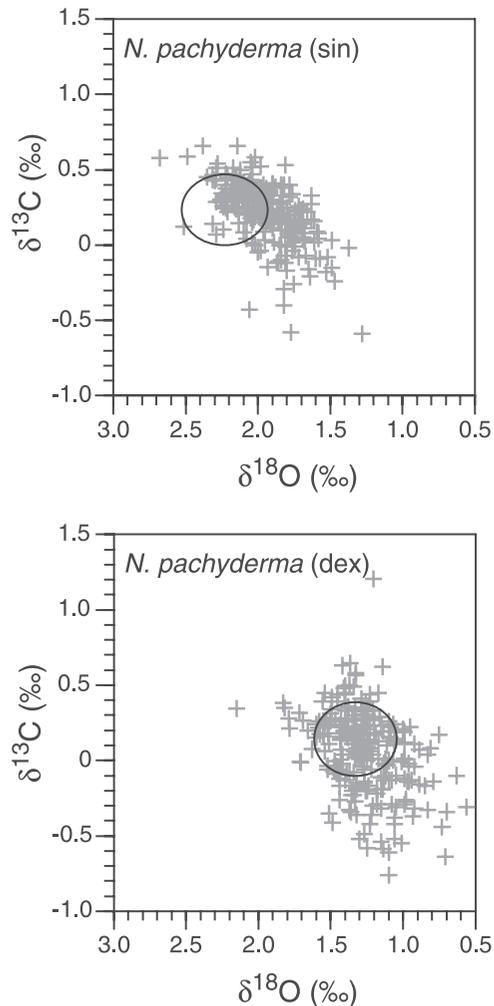
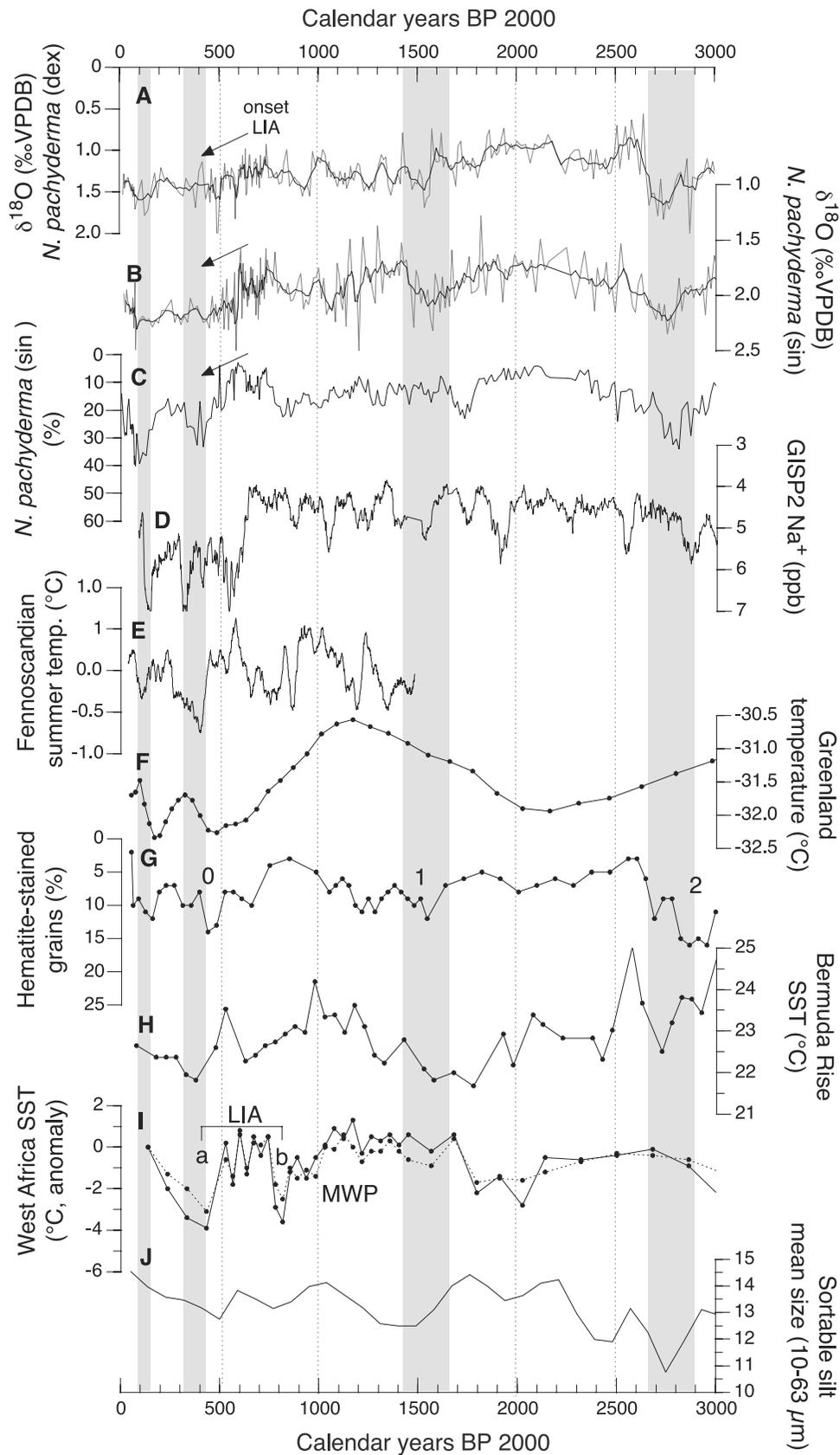


Figure 5. Characteristics of surface waters in the eastern Norwegian Sea over the past 3000 years based on planktic foraminiferal $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ measurements. Circles denote modern $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values of Atlantic Waters in the Nordic Seas [Johannessen, 1992].

Norwegian Sea. According to Denton and Karlén [1973], the glacier advances during this period were approximately equal to advances recorded during the Little Ice Age. At the Vøring Plateau, data from all records in Figure 4 support this assertion. Additional evidence that this period of cooler SST and glacier expansion was a widespread occurrence comes from measurements of sea salt in the GISP2 ice core [O'Brien et al., 1995]. In Figure 6 we let Na^+ represent sea-salt concentrations because this species is the most conservative sea-salt species in the GISP2 record [Kreutz et al., 1997]. Increases in sea-salt concentration are documented for the period 3150–2450 years BP. O'Brien et al. [1995] suggested that changes associated with expansions of the north polar vortex or intensified meridional airflow were responsible for these increases. The GISP2 Na^+ record has recently been calibrated to the Icelandic Low sea level pressure [Meeker and Mayewski,

2002]. The results of O'Brien et al. [1995], however, do not imply that climatic deterioration during this time was as severe as during the LIA. Other marine evidence for colder climates during this time period comes from both faunal as well as sedimentological data in the North Atlantic [Bond et al., 1997, 2001]. Bond et al. [1997] documented distinct increases in the concentration of ice-rafted debris (IRD) at 2850 years BP in sediment cores from both sides of the Atlantic Ocean. The one hundred-year discrepancy in the dating of this cold event in the North Atlantic and at the Vøring Plateau is within the combined error of the radiocarbon dated records. New data from cores MC52 and VM29-191 in the eastern North Atlantic show that the cooler conditions recorded in the eastern Norwegian Sea around 2750 years BP correspond well with IRD-cycle 2 [Bond et al., 2001] (Figure 6). The overall warming of the eastern Norwegian Sea continued from 2750 to about 2500 BP. If the decrease in $\delta^{18}\text{O}$ between 2750 and 2000 years BP is explained by changing SSTs alone, the warming during this time was about 2–2.5°C.

[23] Between 1600 and 2500 years BP surface ocean conditions at the Vøring Plateau indicate relative warmth with the highest SST recorded around 2000 years BP. Lacustrine sedimentary sequences from northern Jostedalbreen ice caps, as well as from glaciers in central Jotunheimen, show evidence of glacier expansion around 2000 years BP, which suggests that the warm surface waters at this time increased the moisture available for winter precipitation in central and western Norway [Matthews et al., 2000; Nesje et al., 2000b]. After about 2000 years BP, climate, as judged by the SST proxy records from the Vøring Plateau, started to cool. Both the record of estimated SST and the oxygen isotope record show that this cooling trend culminated around 1550–1500 years BP (Figure 4). Within this cooling trend, however, both faunal data, and to some extent the oxygen isotope data from *N. pachyderma* (dex), indicate a prominent event of cooler SST centered around 1700 years BP. While the faunal and geochemical responses to this event at the Vøring Plateau appear to be of different magnitude, this may imply some type of decoupling between biological and geochemical processes. There is evidence suggesting that cooling in the Norwegian Sea around 1550–1500 years BP is a widespread phenomenon. Bond et al. [1997] reported a climatic shift toward cooler SST and more IRD at 1450 years BP in North Atlantic sediments. New data by Bond et al. [2001] also support earlier evidence suggesting a cooling of the eastern North Atlantic during this time (Figure 6). If we consider the isotopic SST reconstruction of the Sargasso Sea [Keigwin, 1996], we find that SSTs in general were cooler in that region between 1850 and 1350 years BP (Figure 6). However, the cooler climates during this period seem to have peaked between 1750 and 1550 years BP, somewhat before the 1450 years BP cool peak of Bond et al. [1997]. Direct comparisons over such short timescales as this are difficult, however, because of differences in sedimentation rates, dating control, and marine reservoir age corrections. Additional evidence suggesting a complicated pattern of climate change during this period comes from data off West Africa [deMenocal et al., 2000] where



transfer function SST estimates imply a cooling event at about 1950 years BP, i.e., 500 years older than the cold event at 1450 years BP suggested by *Bond et al.* [1997] (Figure 6).

[24] After about 1550 years BP surface ocean conditions in the southeastern Norwegian Sea warmed over a period of about one to two hundred years. Relatively light $\delta^{18}\text{O}$ values occurred 1450–1400 years BP, while faunal changes, as manifested in the record of estimated SST, suggest a peak in temperature closer to 1500 years BP. This relatively short period of inferred warmth correlates with terrestrial evidence for a period of warm climate at ~ 1500 ^{14}C years BP (i.e., about 1410 years BP) [*Karlén et al.*, 1995]. Another short period of warmer SST about 1200 years BP is suggested by a decrease in the percentage of *N. pachyderma* (sin) and an increase in the abundance of planktic foraminifers in general (Figure 4). The records of estimated SST and the oxygen isotope record, however, do not display any evidence supporting this observation. In contrast, SSTs and $\delta^{18}\text{O}$ of *N. pachyderma* (sin) indicate cooler conditions around 1200 years BP. This is also a period of well-documented glacier advances [e.g., *Karlén*, 1988; *Dahl and Nesje*, 1994; *Hughes and Diaz*, 1994; *Karlén et al.*, 1995]. In conclusion, the period 1800–1400 years BP seems to have been a time span where climate changes of shorter duration were superimposed on a more gradual long-term shift toward relatively cool conditions.

4.3. The Medieval Warm Period and the Little Ice Age

[25] The Medieval Warm Period (MWP) has been suggested as an interval of unusual warmth between roughly the ninth to fourteenth century [*Hughes and Diaz*, 1994, and references therein]. *Lamb* [1977, 1982] notes that northern Europe, Iceland, Greenland as well as North America were warmer during the tenth to thirteenth centuries than the early twentieth century. *Grove and Switsur* [1994] suggested that the MWP could be bracketed by dates of moraines formed during glacier advances preceding and following the MWP. The first of these glacier expansions occurred 1200–1100 years BP in Europe and some other parts of the world [*Grove and Switsur*, 1994]. In northern Sweden a period of glacier advance is dated to ~ 1210 ^{14}C years BP [*Karlén et al.*, 1995]. The most recent period of glacier advance started around 750–700 years BP [*Grove and Switsur*, 1994], which correlates with the onset of the LIA in its broadest sense. Hence the MWP according to *Grove and Switsur* [1994] would have occurred between

1000 and about 750 years BP, with a possible period of glacier readvance around 950–850 years BP. *Briffa et al.* [1990] reconstructed very cold summer temperatures in Fennoscandia between 900 and 850 years BP, and very warm summer temperatures from 850–800 years BP. The reconstruction of *Briffa et al.* [1990] does not indicate overall warmer summer temperatures between 1000 and 700 years BP. *Jennings and Weiner* [1996], on the other hand, found evidence in foraminifer and lithofacies records from the Nansen Fjord (east Greenland) suggesting stable climatic conditions warmer than at present between 1270 and 890 years BP.

[26] Compilations of different evidence (from tree rings, ice cores, historic sources, and glacial and geological records) of climate change between 1200 and 600 years BP suggest a large spatial differentiation [*Hughes and Diaz*, 1994], which complicates comparisons. Looking closely at the climatic proxy records from the Vøring Plateau (Figure 6) between 1200 and 600 years BP, we see evidence suggesting considerable climatic change during this period. Between 1200 and 800 years BP conditions were slightly warmer or as warm as today, which is in agreement with *deMenocal et al.* [2000] who suggested that the MWP was only marginally warmer than at present. *Jennings and Weiner* [1996] documented two cold intervals centered at about at 850 and 630 years BP. The first event correlates with an increase at the Vøring Plateau in the relative abundance of *N. pachyderma* (sin) and a relatively low total abundance of planktic foraminifers in general (Figure 4). There is also some overlap between this event and the period between 950 and 850 years BP of minor glacier advances proposed by *Grove and Switsur* [1994]. Moreover, Fennoscandian summer temperatures [*Briffa et al.*, 1992] show markedly cooler conditions between 890 and 850 years BP. Paradoxically, this event apparently coincided with the Medieval Solar Activity Maximum [*Jirikowic and Damon*, 1994].

[27] The period 800–550 BP was characterized by warmth at the Vøring Plateau, which is also evident in the benthic stable oxygen isotope record from this locality [*Risebrobakken et al.*, 2003]. However, superimposed on this period of warmer climate is a short period of colder SST at about 700–600 years BP. This event, as expressed in the records of estimated SST and abundance variations of *N. pachyderma* (sin), coincides with the second interval of cold climate in the Nansen Fjord that culminated at 630 years BP [*Jennings and Weiner*, 1996].

Figure 6. (opposite) Late Holocene variations in surface ocean conditions in the Norwegian Sea from JM97-948/2A and MD95-2011 compared with ice core, terrestrial and marine paleoenvironmental records on a calendar year BP/AD 2000 basis. Upper two panels (a and b) show the oxygen isotope composition of *N. pachyderma* (dex) and *N. pachyderma* (sin), (c) relative abundance of *N. pachyderma* (sin) (on an inverted scale), (d) the GISP2 Na^+ record [*O'Brien et al.*, 1995] (smoothed by a 21-point moving average and plotted on an inverted scale), (e) smoothed record of Northern Fennoscandian summer temperature anomalies [*Briffa et al.*, 1992], (f) GRIP past surface temperature [*Dahl-Jensen et al.*, 1998], (g) drift-ice record based on the percentage of hematite-stained grains (on an inverted scale) from core MC52 and VM29-191, eastern North Atlantic [*Bond et al.*, 2001] (number 0 to 2 denote the IRD events of *Bond et al.* [1997, 2001]), (h) Bermuda Rise SST based on oxygen isotope variations of *G. ruber* [*Keigwin*, 1996], (i) West African faunal SST anomalies (cold; stippled, warm; solid) [*deMenocal et al.*, 2000] (LIA; Little Ice Age, MWP; Medieval Warm Period), (j) grain size data representing the speed of Iceland-Scotland Overflow Water [*Bianchi and McCave*, 1999]. Raw grain-size data have been smoothed by Gaussian interpolation using a 90-year sampling interval and a 300-year window.

[28] Evidence for warmer conditions in the period 800–550 years BP is also found in the GISP2 ice core [O'Brien *et al.*, 1995] and in Fennoscandian tree ring chronologies [Briffa *et al.*, 1992] (Figure 6). In the GISP2 ice core, low Na^+ concentrations indicate milder climates in particular between 850 and 650 years BP, while the reconstructed Fennoscandian temperature record exhibits warm conditions between 650 and 460 years BP. In the latter record, peak summer warmth occurred 590–570 years BP, which is consistent with results from the Vøring Plateau.

[29] During the Little Ice Age (LIA), climate cooled on a seemingly worldwide basis [e.g., Lamb, 1977, 1982; Broecker, 2001]. Various authors define the start of this period differently. Lamb [1977] restricted the LIA to the period from 450 to 150 years BP. Grove and Switsur [1994] suggested glacier expansion between about 750–700 and 150–110 years BP, but defined only the last four centuries of this period (i.e., 550–150 years BP) as belonging to the LIA. According to Porter [1986], the LIA spanned the time interval between 750 years BP to about 80 years BP. However, discrepancies, especially related to the start of the LIA, may be the result of the different proxies used to delimit the LIA. Glacier advances, for example, can occur not only during times of colder climate but also during times of enhanced winter precipitation [Dahl and Nesje, 1996].

[30] The results from the Vøring Plateau indicate that over the past one thousand years climatic conditions in the southeastern Norwegian Sea have experienced several notable oscillations between warmer and cooler conditions (Figures 4 and 6). Based on the combined results from JM97-948/2A/MD95-2011, the initiation of the LIA in this region seems to have occurred around 700–600 years BP. Over the next 200 years, SSTs cooled up to 2°C based on the stable oxygen isotope record, if salinity changes were negligible. From 600–550 years BP to 400 years BP, the relative abundance of *N. pachyderma* (sin) increased by close to 30%. An increase of this magnitude implies a cooling of about 1.8–1.9°C during this time span. Magnetic susceptibility and Na^+ data from GISP2 also suggest a rather abrupt start of the LIA between 675 years BP [Domack and Mayewski, 1999] and 650 BP [O'Brien *et al.*, 1995]. Both proglacial lacustrine and terrestrial data from south-central Norway suggest that glacier expansion prior to the LIA began close to 650 years BP [Dahl and Nesje, 1994, and references therein]. For the eastern North Atlantic, the IRD record of Bond *et al.* [2001] suggests that the onset of the LIA occurred closer to 800 years BP in this region (Figure 6).

[31] Off West Africa, deMenocal *et al.* [2000] documented two discrete 3 to 4°C cooling events over the past 1000 years, which were suggested to represent the LIA (Figure 6). The first event, LIA_B, had its maximum around 800 years BP, which is too old relative to the LIA as defined in NW-Europe [e.g., Grove and Switsur, 1994]. Regardless of whether or not this event took place during the earlier phase of the LIA or the later part of the MWP (depending on definition), this cooling correlates with the relatively cool period about 750 years BP at the Vøring Plateau and with cooler Fennoscandian summer temperatures [Briffa *et al.*, 1992]. The observation of the LIA as being a two-step event

[deMenocal *et al.*, 2000] is interesting. Two events of lower temperature during the past 1000 years, possibly belonging to the LIA, are also evident in data from the Sargasso Sea [Keigwin, 1996], the eastern North Atlantic [Bond *et al.*, 2001], Greenland [Dahl-Jensen *et al.*, 1998] and Fennoscandia [Briffa *et al.*, 1992] (Figure 6). At the Vøring Plateau both the isotopic data, and especially the faunal data, indicate two cold phases centered at 400 years BP and around 100 years BP. Both events correspond to decreases in Fennoscandian summer temperatures as recorded by Briffa *et al.* [1992] (Figure 6). The timing of the two cooling events at the Vøring Plateau also corresponds well with the IRD data of Bond *et al.* [2001] from the eastern North Atlantic. However, the results from the North Atlantic [Bond *et al.*, 2001], Sargasso Sea [Keigwin, 1996], Greenland [Dahl-Jensen *et al.*, 1998], off West Africa [deMenocal *et al.*, 2000] and this study show that there are discrepancies in the timing of the two LIA events between different regions. Data from Dahl-Jensen *et al.* [1998] suggest that the two LIA cooling events on Greenland occurred slightly earlier, but still within dating uncertainty, of events on the Vøring Plateau. Data from more southerly locations in the Sargasso Sea [Keigwin, 1996] and off West Africa [deMenocal *et al.*, 2000] show larger discrepancies compared to the timing of these events in the Norwegian Sea, eastern North Atlantic and Fennoscandia (Figure 6). This is in contrast with deMenocal *et al.* [2000] who suggested strong in-phase linkages between high- and low-latitude paleoclimate records from the North Atlantic region. The two LIA events recorded off West Africa are not synchronous with the two LIA cool periods on Greenland, in the eastern North Atlantic, in the Norwegian Sea and in Fennoscandia. LIA event A, as defined by deMenocal *et al.* [2000] is possibly coherent with the oldest of the two LIA events recorded at high northern latitudes, while LIA event B clearly predated the LIA in this region (Figure 6). Dating uncertainties associated with marine AMS dates make detailed comparisons between marine, terrestrial and ice core records difficult. Hence the phase offsets between different records can be extremely difficult to quantify due to dating uncertainties, as well as the low temporal resolution of many records, and this may explain the discrepancies in the timing of LIA events in the North Atlantic. Alternatively, the onset of LIA events was time-transgressive from low to high latitudes in the North Atlantic as a strict interpretation of published dates would appear to suggest.

4.4. Wider Climatic Implications

[32] Over the past 3000 years, drift-ice events in the North Atlantic [Bond *et al.*, 2001] seem to be negatively correlated to SST in the eastern Norwegian Sea (Figure 6). However, stable isotopic and faunal data from JM/979-48 and MD95-2011 spanning the entire Holocene fail to indicate that the drift-ice events reported by Bond *et al.* [1997, 2001] are consistently correlated to temperature cycles in the Norwegian Sea throughout the Holocene [Risebrobakken *et al.*, 2003]. Nevertheless, the large millennial-scale changes in surface ocean hydrography over the past 3000 years, as inferred from the SST reconstruction in this study,

and the drift-ice cycles of *Bond et al.* [1997, 2001] do correlate with reported high-latitude changes in terrestrial climate and atmospheric circulation (see discussion above). The millennial-scale cycles in drift ice have a period of about 1500 years [*Bond et al.*, 2001]. *Bond et al.* [1997, 1999] related this cyclicity to Dansgaard/Oeschger type cyclicity and suggested that the North Atlantic oscillated in a 1–2 kyr mode during the last glacial, as well as in an attenuated mode of some cyclicity during the past 10 kyr. *Schulz and Paul* [2002] questioned the approximate 1500-year spacing between Holocene cool events, instead suggesting a possible correlation between the recurring changes in North Atlantic IRD abundance and the 900-year cyclicity of air-temperature changes (as inferred from $\delta^{18}\text{O}$ measurements) in the GISP2 ice core record.

[33] *Bianchi and McCave* [1999] presented grain-size data from core NEAP15K showing systematic (1500-year) changes in thermohaline circulation in the Holocene North Atlantic (Figure 6), which suggests a link between millennial-scale climate change and thermohaline circulation. However, the reductions in North Atlantic Deep Water production inferred from the grain-size data only correlate with some of the drift-ice events of *Bond et al.* [2001]. In addition, *Schulz and Paul* [2002] recently showed that the grain-size time series of *Bianchi and McCave* [1999] are difficult to distinguish from white noise, which raises questions as to the significance of comparisons between grain-size and other paleoceanographic proxy data. Nonetheless, sediment color data, from the same core (NEAP15K), also used as a proxy for changes in NADW, show high-frequency fluctuations with periodicities equivalent to 550 and 1000 year, as well as 1650-year cyclicity [*Chapman and Shackleton*, 2000]. The latter is comparable to the 1500-year cycle of *Bond et al.* [1997, 2001]. The 550- and 1000-year cyclicities are identical to the dominant cyclicities in the Holocene GISP2 ice core record [*Chapman and Shackleton*, 2000], suggesting a strong interconnection between atmospheric and oceanic variability in the North Atlantic realm. *Stuiver et al.* [1995] suggested a solar origin of the 530-year cycle, which indicates a possible connection between solar variability and NADW circulation [*Chapman and Shackleton*, 2000]. More recently, *Bond et al.* [2001] investigated a possible solar influence on Holocene North Atlantic climate. They compared Holocene changes in the drift-ice records with production rate changes of cosmogenic nuclides (^{10}Be and ^{14}C), the latter being related to solar activity and solar winds. The results show a coherency between changes in production rate of cosmogenic nuclides and millennial- and centennial-scale changes in the drift-ice records. However, there are difficulties relating the correlation between past surface ocean conditions and solar variability with modern observations of North Atlantic climate variability. At present, the North Atlantic Oscillation (NAO) is the dominant mode of winter climate variability in the North Atlantic region, with a typical dipole temperature anomaly pattern [*Hurrell*, 1995]. Modern glacier fluctuations in western Norway are, for example, highly coherent with the NAO index [*Nesje et al.*, 2000a, 2001]. On longer timescales, periods of low winter precipitation appear to be synchronous with periods of

enhanced drift ice during the Holocene. Still, the surface ocean temperature anomalies in the North Atlantic region during the last three drift-ice cycles are not consistent with a NAO type of temperature anomaly pattern [*Bond et al.*, 2001]. To explain the temperature anomaly patterns during these cycles, *Bond et al.* [2001] suggested that decreases in surface ocean salinity, inferred from decreases in surface ocean $\delta^{18}\text{O}$, accompanied the drift ice increases and led to reductions in NADW. Although solar variability alone does not seem to drive climate variability in the North Atlantic region, it may be responsible for triggering system feedbacks.

5. Conclusions

[34] Planktic foraminiferal faunal and stable isotopic data from sediments recovered at the Vøring Plateau show that sea surface temperatures in the eastern Norwegian Sea during the past 3000 years were highly variable. From the foraminiferal faunal and stable isotopic data we draw the following conclusions:

[35] 1. Covariance between temporal changes in the foraminiferal assemblage and the foraminiferal stable oxygen isotopic records reflect a concurrent response to sea surface temperature change.

[36] 2. Significant cooling events, during which sea surface temperatures were lower than present, occurred at about 2750, 1550, 400 and 100 years BP. The decrease in SSTs around 2750 years BP was especially marked and suggests that climate at this time was at least as cold as the cooling during the Little Ice Age. The observed decreases in SST at 2750 and 1550 years BP appear to correspond to periods of increase in the presence of drift ice in the North Atlantic [*Bond et al.*, 1997, 2001].

[37] 3. Surface ocean conditions during the so-called Medieval Warm Period were highly variable with notable changes in sea surface temperatures. The period 800–550 years BP was characterized by relatively warm conditions at the Vøring Plateau.

[38] 4. A cooling trend leading to the LIA began around 700–600 years BP. This correlates well with both proglacial lacustrine and terrestrial data from south-central Norway which indicate that glacier expansion prior to the LIA started close to 650 years BP [*Dahl and Nesje*, 1994]. Data from the Vøring Plateau indicate two cold phases during the LIA centered at 400 years BP and around 100 years BP, both corresponding to decreases in Greenland air temperature and Fennoscandian summer temperatures [*Briffa et al.*, 1992] and increases of drift ice in the eastern North Atlantic [*Bond et al.*, 2001].

[39] 5. The two distinct cooling events during the LIA appear to be synchronous at high northern latitudes. However, comparisons between high-latitude records from the eastern Norwegian Sea (this study), Greenland [*Dahl-Jensen et al.*, 1998], the eastern North Atlantic [*Bond et al.*, 2001], and Fennoscandia [*Briffa et al.*, 1992], and records from more southerly locations, i.e., the Sargasso Sea [*Keigwin*, 1996] and off West Africa [*deMenocal et al.*, 2000], show no apparent consistency between the timing of the two cool events. The phase offsets between these

locations may in part be due to dating uncertainties and/or low temporal resolution. If real, however, the temporal offsets would tend to indicate that the onset of the two LIA cool events was time-transgressive from low to high latitudes in the North Atlantic.

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