

Centennial-scale evolution of Dansgaard-Oeschger events in the northeast Atlantic Ocean between 39.5 and 56.5 ka B.P.

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[1] There is much uncertainty surrounding the mechanisms that forced the abrupt climate fluctuations found in many palaeoclimate records during Marine Isotope Stage (MIS)-3. One of the processes thought to be involved in these events is the Atlantic Meridional Overturning Circulation (MOC), which exhibited large changes in its dominant mode throughout the last glacial period. Giant piston core MD95-2006 from the northeast Atlantic Ocean records a suite of palaeoceanographic proxies related to the activity of both surface and deep water masses through a period of MIS-3 when abrupt climate fluctuations were extremely pronounced. A two-stage progression of surface water warming during interstadial warm events is proposed, with initial warming related to the northward advection of a thin warm surface layer within the North Atlantic Current, which only extended into deeper surface layers as the interstadial progressed. Benthic foraminifera isotope data also show millennial-scale oscillations but of a different structure to the abrupt surface water changes. These changes are argued to partly be related to the influence of low-salinity deepwater brines. The influence of deepwater brines over the site of MD95-2006 reached a maximum at times of rapid warming of surface waters. This observation supports the suggestion that brine formation may have helped to destabilize the accumulation of warm, saline surface waters at low latitudes, helping to force the MOC into a warm mode of operation. The contribution of deepwater brines relative to other mechanisms proposed to alter the state of the MOC needs to be examined further in future studies.

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1. Introduction

[2] Over the last few decades, a large number of studies have provided evidence that the Earth's climate system underwent a series of abrupt oscillations during the last glacial period, with amplitudes approaching those demonstrated for orbital-scale climate variability [e.g., Johnsen *et al.*, 1992; Dansgaard *et al.*, 1993]. These "Dansgaard-Oeschger" (D-O) oscillations were first identified in Greenland ice cores [Johnsen *et al.*, 1992] and have been subsequently correlated to similar abrupt events in numerous other palaeoclimate archives [e.g., Voelker and Workshop Participants, 2002]. They follow an approximate pacing of 1500 years [Schulz, 2002], and are interspersed between the large Heinrich (H) ice-rafting events found in the North Atlantic [Bond *et al.*, 1993; Hemming, 2004]. D-O events have been intensively investigated because of

their large thermal amplitude (exceeding 15°C in the Greenland Summit ice cores [Huber *et al.*, 2006]), abruptness (with cold-warm transitions occurring over a period of only a few decades or less [Austin *et al.*, 2004; Shackleton *et al.*, 2004]) and their association with large-scale reorganizations of both atmospheric and oceanographic circulation regimes [Hendy and Kennett, 1999; Peterson *et al.*, 2000; Wang *et al.*, 2001; Kiefer *et al.*, 2001; Broecker, 2003a].

[3] Although atmospheric teleconnections must have played a key role in the generation and subsequent transfer of the D-O climate signal to widely dispersed ice, marine, speleothem and lake archives, much attention has been paid to the importance of oceanic processes in contributing to D-O events [Boyle, 2000]. The importance of the Atlantic Meridional Overturning Circulation (MOC) in the amplification and transmission of abrupt climate signals has been highlighted because of its ability to affect the global transfer of heat, and because of a close relationship between the activity of the MOC and abrupt climate signals recorded at high northern latitudes [e.g., Boyle and Keigwin, 1987; Broecker, 1994; Clark *et al.*, 2002; McManus *et al.*, 2004]. Many studies point to a reduction in deepwater formation in the Nordic Seas and northern North Atlantic Ocean and the northward spread of southern-sourced deep waters during D-O stadials [Keigwin and Boyle, 1999; Hagen and Hald, 2002], linked to the release of freshwater and icebergs to sites of overturning from the circum-North

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Atlantic ice masses [Bond and Lotti, 1995; Elliot *et al.*, 1998; Lackschewitz *et al.*, 1998; Knutz *et al.*, 2001]. Dokken and Jansen [1999] and van Kreveld *et al.* [2000] suggested that the formation of sea ice brines in the Nordic Seas during D-O stadials could have destabilized stadial boundary conditions by generating an embryonic form of overturning circulation that eventually forced the MOC into a “warm” interstadial mode. However, the volume of brines produced at these times, along with their spatial extent, is not well defined. Additionally, the isotopic data used to identify brines may be complicated by deep sea temperature changes in some instances [e.g., Rasmussen and Thomsen, 2004; Rasmussen *et al.*, 1996; Bauch and Bauch, 2001] therefore having different implications for forcing the MOC between the “on” and “off” modes characteristic of interstadial-stadial oscillations.

[4] Several studies have demonstrated that low-latitude processes may also play an important role in D-O events through changes in the atmospheric transport of heat and moisture to higher latitudes via alternately weakened and strengthened meridional pressure gradients [e.g., Cane and Clement, 1999; Peterson *et al.*, 2000]. These processes may be closely related to changes in MOC activity through the buildup of warm, saline waters in the subtropics during stadials [Chapman and Maslin, 1999; Schmidt *et al.*, 2006; Levi *et al.*, 2007] which eventually would have reached a salinity imbalance with relatively fresh polar waters in the Northern Atlantic, forcing the rapid resumption of interglacial warmth when this imbalance was destabilized.

[5] Here we extend these detailed studies with new observations from the northeast Atlantic Ocean using palaeo-oceanographic records of ice-rafted debris (IRD) concentrations, foraminifera-based estimates of sea surface temperature (SST) and stable isotopes. These observations cover the period of stadial-interstadial variations over a full “Bond cycle” [Broecker, 1994] approximately between North Atlantic Ash Zone (NAAZ) 2 and Heinrich event 4 and are characterized by intense climate instability in both surface and deepwater reconstructions. We aim to investigate how changes in surface and deep water conditions were coupled in order to examine the mechanisms forcing the ocean-climate system at these timescales.

2. Study Site and Material

[6] Giant piston core MD95-2006 was retrieved during the International Marine Past Global Change Study (IMAGES) I, R/V *Marion Dufresne* cruise MD101 from the Barra Fan in the eastern Rockall Trough, northeast Atlantic Ocean (Figure 1) (57°01.82N, 10°03.48W, 2130 m water depth). Surface waters at this location are currently influenced by the warm, saline waters of the North Atlantic Current, which is most active during late summer and autumn as it moves northward toward sites of convective overturning in the Nordic Seas. The seabed is overlain by Upper North Atlantic Deep Water (UNADW). This water mass is formed from Nordic Sea overflow waters, which enter the NE Atlantic over the Wyville-Thomson and Iceland-Scotland Ridges and combine with deep waters formed by convection in the Labrador Sea [Van Aken, 2000].

UNADW recirculates northward from the open northeast Atlantic into the eastern side of the Rockall Trough, before turning southward along the western side in a regional gyre. As well as being well placed to monitor key oceanographic variables, the core site is also sensitive to the input of coarse-grained IRD from the British (BIS), Icelandic, and Fennoscandian Ice Sheets [Knutz *et al.*, 2001; Peters *et al.*, manuscript in preparation, 2008]. Reconstructions of glacial-interglacial changes in the circum-North Atlantic Region [e.g., Pflaumann *et al.*, 2003; Kucera *et al.*, 2005; De Vernal *et al.*, 2005] show that the site of MD95-2006 was affected strongly by cold polar waters and seasonal sea ice during the Last Glacial Maximum (LGM), when the polar front shifted southward compared to its present-day location east of the East Greenland margin [e.g., van Kreveld *et al.*, 2000].

[7] The whole-core stratigraphy of MD95-2006 has been presented by Kroon *et al.* [2000] and Knutz *et al.* [2001, 2002]. A greatly expanded glacial section between ~26–18 ¹⁴C ka corresponds to a period of BIS advance, which destabilized the Hebridean shelf edge sediments and caused numerous mass flow deposits to occur over the lower continental slope [Wilson *et al.*, 2002; Bowen *et al.*, 2002]. These deposits are not present in the lower part of the core, which is made up entirely of contourite silts deposited by UNADW-related bottom currents. Close inspection of this sediment section revealed no evidence of coarse debris layers or reworked shallow benthic foraminifera, suggesting that an in situ climate signal could be generated. The studied core section corresponds to the interval between Heinrich layer 4 [Knutz *et al.*, 2001] and NAAZ-2 [Austin *et al.*, 2004], where millennial-scale fluctuations in sediment properties are well pronounced [Wilson and Austin, 2002] and the Greenland D-O cycles are particularly well expressed [Bond *et al.*, 1993; Alley *et al.*, 2001]. The data shown for this study interval should theoretically be reproducible for other millennial-scale D-O cycles in the region, and this could be pursued in other cores not affected by the proximal BIS to the same extent as MD95-2006.

3. Methods

3.1. Core Sampling

[8] The interval of MD95-2006 between 2500 and 2820 cm was sampled every 2 cm using 1 cm thick sample slices. Each sample was dried at 40°C, weighed, and wet sieved over a 63 μ m mesh using deionized water. Sand-sized material was dried, reweighed and stored in clean glass vials for counting foraminifera and mineral grains.

3.2. Isotope Measurements

[9] Planktonic foraminifera oxygen isotopes ($\delta^{18}\text{O}$) were measured on approximately 10 individuals of *Neoglobobulimina pachyderma* (*s*) >150 μ m per sample throughout the study period. This species is common in polar waters <7°C [Be and Tolderlund, 1971] and has commonly been used for isotopic measurements of surface waters at high and midlatitudes during glacial periods [e.g., Bond *et al.*, 1993; Fronval *et al.*, 1995; Lackschewitz *et al.*, 1998; Elliot *et al.*, 1998; van Kreveld *et al.*, 2000]. How-

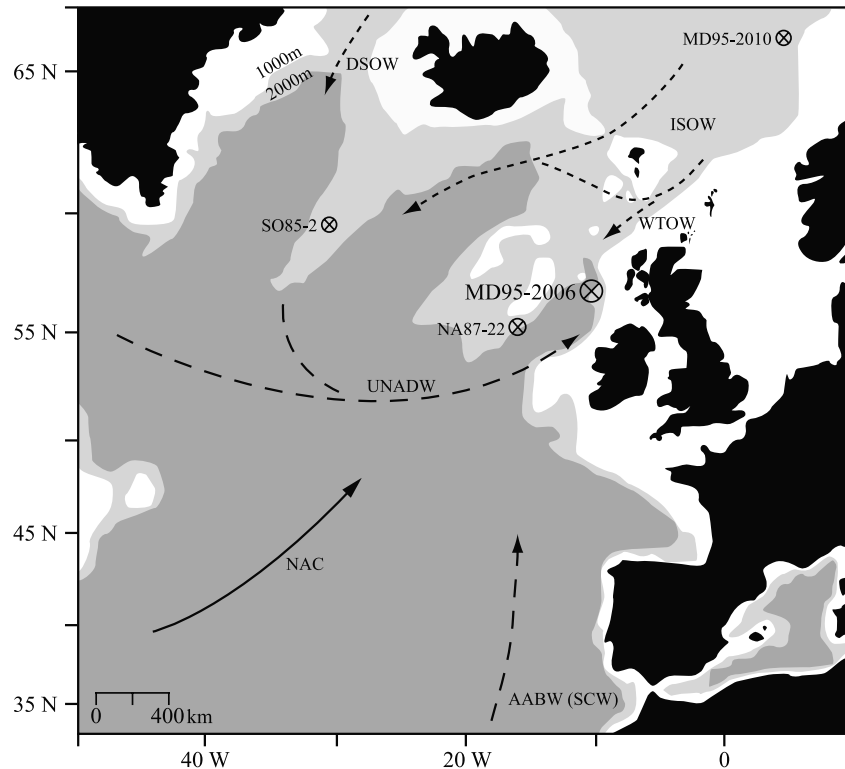


Figure 1. Locations of MD95-2006 and other published cores mentioned in the text. UNADW, Upper North Atlantic Deep Water; NAC, North Atlantic Current; AABW/SCW, Antarctic Bottom Water/Southern Component Water; ISOW, Iceland-Scotland Overflow Water; WTOW, Wyville-Thomson Overflow Water; DSOW, Denmark Strait Overflow Water.

ever, *N. pachyderma* (*s*) calcifies below the surface at depths up to 200 m [Bauch *et al.*, 1997], and so may not reveal additional information on water hydrography at shallower depths. Therefore, samples of the shallow-dwelling planktonic species *Globigerina bulloides* were also measured where they could be picked in high enough abundances. *N. pachyderma* (*s*) is considered to calcify during summer within its ecological range [Be and Tolderlund, 1971; Reynolds and Thunell, 1986], while *G. bulloides* is associated with the spring bloom [Ganssen and Kroon, 2000]. However, because *N. pachyderma* (*s*) is not generally found in waters warmer than $\sim 7\text{--}8^\circ\text{C}$, it may calcify outside of summer when it is at the warmest extent of its ecological range [Schmidt and Mulitza, 2002; King and Howard, 2005].

[10] Benthic isotopes were measured on 1–6 individuals of the epibenthic species *Cibicidoides wuellerstorfi*. Benthic and planktonic measurements were made with a ThermoFinnigan MAT 252 mass spectrometer coupled to a CarboKiel-II carbonate preparation device. Isotope values are referred to the Vienna Pee Dee Belemnite scale (VPDB) through calibration to the NBS-19 carbonate standard. External reproducibility was monitored through repeat measurements of an internal laboratory standard and was 0.03‰ for $\delta^{13}\text{C}$ and 0.06‰ for $\delta^{18}\text{O}$. Benthic $\delta^{18}\text{O}$ values have been shifted by +0.64‰ to accommodate offset of

C. wuellerstorfi from oxygen isotope equilibrium [Shackleton, 1974].

3.3. Sea Surface Temperature

[11] Summer sea surface temperatures (SSTs) at 10 m water depth have been estimated from assemblage counts of planktonic foraminifera using the Artificial Neural Network (ANN) and modern analogue technique using a similarity index (SIMMAX) techniques with the modern North Atlantic faunal data set of Kucera *et al.* [2005]. Eighty-four samples were dry sieved at 150 μm and split to obtain 300 (minimum 216, maximum 660) planktonic individuals, which were identified according to the taxonomic scheme of Pflaumann *et al.* [2003]. SIMMAX estimates are based on the 10 most similar modern analogs, while ANN estimates are based on the 10 networks which were best in learning the SST-faunal relationship in a different subset of the training data set [Malmgren *et al.*, 2001].

3.4. Age Model

[12] Three AMS ^{14}C dates are available for the studied portion of MD95-2006, which are all close to the limits of the radiocarbon technique. The measurement and calibration of the ^{14}C dates has been thoroughly discussed by Wilson and Austin [2002], who developed a polynomial age depth model based on these and other ^{14}C age determinations for shallower core intervals. The identification of

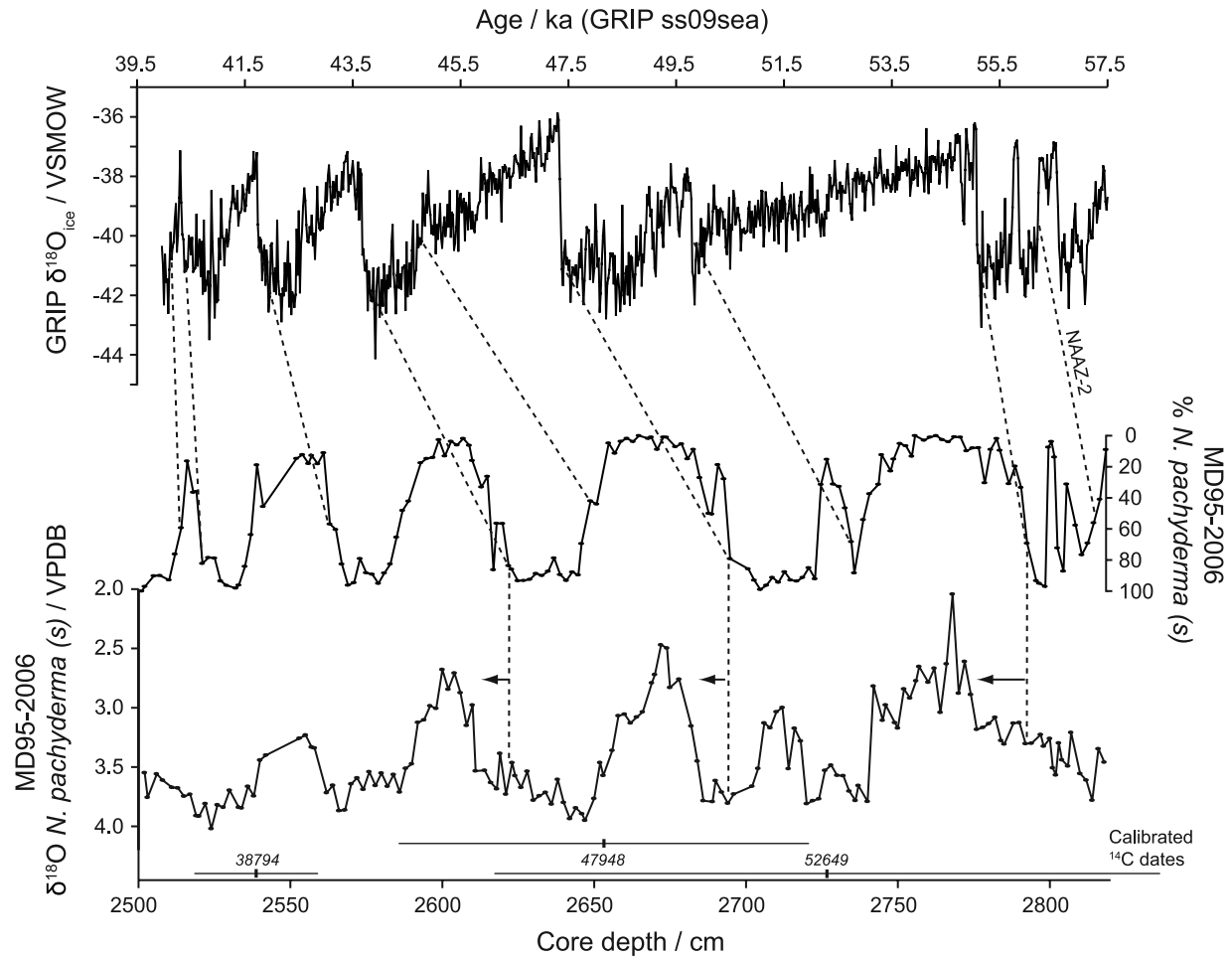


Figure 2. Construction of the age model for the studied core section. Tie points between percent *N. pachyderma* (s) and GRIP $\delta^{18}\text{O}_{\text{ice}}$ plotted on the ss09sea timescale [Johnsen *et al.*, 2001] are shown as dashed lines. Stratigraphic offsets between stadal-interstadial transitions represented in percent *N. pachyderma* (s) and *N. pachyderma* (s). Here $\delta^{18}\text{O}$ are shown as arrows. The 2σ age ranges of previously published rangefinder radiocarbon dates [Wilson and Austin, 2002] are shown at the base of Figure 2 as black lines.

NAAZ-2 at 2817 cm depth [Austin *et al.*, 2004] provides an additional independent age constraint, since this marker has also been found in other sediment cores from the North Atlantic and Nordic Sea region and in the Greenland Ice Core Project (GRIP) and Greenland Ice Sheet Project 2 (GISP2) Greenland ice core records [Gronvald *et al.*, 1995; Meese *et al.*, 1997]. The location of this marker on the transition out of interstadial (IS)-15 in the GISP2 and GRIP $\delta^{18}\text{O}$ records and in the abundance record of *N. pachyderma* (s) in MD95-2006 provides both a guide for fine tuning the marine chronology to the Greenland ice core age depth models and to correlating to other sediment records covering this time period (Figure 2).

[13] The method of transferring ice core chronologies to marine sediment records has been widely applied in the circum-North Atlantic region [e.g., Bond *et al.*, 1993; Elliot *et al.*, 1998; van Kreveld *et al.*, 2000; Pailler and Bard, 2002; Shackleton *et al.*, 2004], usually through the identification of similar patterns in surface temperature data and

the Greenland $\delta^{18}\text{O}$ records. This method presupposes a teleconnection between the two locations, which is most likely a shift in the location of the boundary separating cold polar water (and air masses) from subpolar, more temperature water (air) masses. Therefore, marine proxy indicators which bear a strong connection to the location of this boundary should yield the most reliable results. The exact temperature proxy data used to achieve this correlation differs between studies. Some employ planktonic $\delta^{18}\text{O}$ data [e.g., Bond *et al.*, 1993] while others use either quantitative estimates of SST or the relative abundance of indicator foraminifera species [e.g., van Kreveld *et al.*, 2000; Pailler and Bard, 2002]. In MD95-2006, the transition from stadal to interstadial conditions in *N. pachyderma* (s) $\delta^{18}\text{O}$ lags the warming recorded in percent *N. pachyderma* (s) by 100–200 years (Figure 2). Given the close association of percent *N. pachyderma* (s) with the location of the modern polar front [e.g., Pflaumann *et al.*, 1996], and the observation that NAAZ-2 falls on the IS-15 boundary defined by percent

Table 1. Tie Points Used to Construct the MD95-2006 Age Model^a

MD95-2006 Depth (cm)	GRIP ss09sea Age (Years B.P.)
2515	40,220
2522	40,360
2564	41,740
2623	43,800
2650	44,700
2696	47,420
2736	49,800
2794	55,100
2817 (NAAZ-2)	56,220

^aNAAZ-2, North Atlantic Ash Zone 2.

N. pachyderma (*s*) in MD95-2006 and $\delta^{18}\text{O}$ in the GRIP ice core [Austin *et al.*, 2004], this proxy is suggested to be the most reliable for achieving a graphical correlation to the GRIP $\delta^{18}\text{O}$ stratigraphy. The variable hydrographic influences on the planktonic $\delta^{18}\text{O}$ data mean that a firm temperature influence cannot always be guaranteed, particularly in a region which potentially experienced a range of surface water masses with different temperatures and salinities during the last glacial period [e.g., Vidal *et al.*, 1998; Lackschewitz *et al.*, 1998; Elliot *et al.*, 2002]. The lag of planktonic $\delta^{18}\text{O}$ relative to percent *N. pachyderma* (*s*) is likely to be related both to this hydrographic variability and to ecological changes in the depth and seasonality preferences of *N. pachyderma* (*s*) through time as discussed in section 5.1. This lag also has implications for the construction of other marine chronologies based on matching $\delta^{18}\text{O}$ signatures to the Greenland ice core records, especially in high-latitude settings where these data are generated using *N. pachyderma* (*s*).

[14] The GRIP ss09sea timescale [Johnsen *et al.*, 2001] has been transferred to MD95-2006 by correlating abrupt stadal-interstadial transitions visible in abundance record of *N. pachyderma* (*s*) to similar events recorded in the GRIP $\delta^{18}\text{O}_{\text{ice}}$ record (Figure 2 and Table 1). Ages for intervening samples have been estimated by linear interpolation between age control points. This method reasonably assumes a minimal time lag between the recording of abrupt climate transitions in Greenland ice and at the site of MD95-2006. The dating error associated with this assumption is on the order of a few decades or less [Shackleton *et al.*, 2004]. This is far less than would be associated with an age model based solely on calibrated radiocarbon dates in this time period. The age model suggests mean sedimentation rates of 23 cm per 1000 years, (reaching as high as 50 cm per 1000 years), and a mean time step of 106 years for all data except sea surface temperatures (SSTs), which have been produced at a mean time step of 193 years. The studied time interval ranges from Greenland Interstadials (GIS) 9 to 15 (~39.5–56.5 ka B.P.). All data will be archived within the NOAA National Climatic Datacenter database (<http://www.ncdc.noaa.gov/paleo/paleo.html>).

4. Results

[15] SIMMAX and ANN SST reconstructions parallel each other closely, supporting the accuracy of the inferred

values (Figure 3). SIMMAX dissimilarity coefficients do not fall below 0.83, suggesting that no-analog situations do not significantly affect the data [e.g., Pflaumann *et al.*, 2003; Kucera *et al.*, 2005]. However, SIMMAX estimates do diverge from the ANN estimates to much higher values during the warmest parts of each interstadial. These SST jumps correspond to larger inferred 2σ errors, and imply that they are based on a more heterogeneous subset of modern analogs. For these reasons, the ANN SST estimates are considered more reliable.

[16] The pattern of SST changes in MD95-2006 closely matches the GRIP $\delta^{18}\text{O}_{\text{ice}}$ record of IS-15 to 9, in support of the approach used to develop the age model (Figure 3). The main features of the SST data are reproduced in the SIMMAX, ANN and percent *N. pachyderma* (*s*) records, which all show large changes between stadal and interstadial periods. The magnitude of these SST changes is approximately 7°C , with minimum temperatures of $3\text{--}5^{\circ}\text{C}$ during stadials and maximum temperatures of $\sim 11^{\circ}\text{C}$ during interstadials. This range of stadal-interstadial SST variation is intermediate between the ranges inferred from summer faunal SST reconstructions in the western Atlantic Ocean ($\sim 1\text{--}2^{\circ}\text{C}$) [Vautravers *et al.*, 2004] and from the Iberian margin ($\sim 9^{\circ}\text{C}$) [de Abreu *et al.*, 2003]. The interstadial intervals recorded in MD95-2006 are not characterized by the steady decline in temperatures seen over Greenland, in agreement with the observation of Vautravers *et al.* [2004] in the western North Atlantic Ocean, although some intrainterstadial variability does exist. This variability is most obvious in the transitions into IS-12 and to a lesser extent IS-11, when small temperature reversals can be seen.

[17] *N. pachyderma* (*s*) $\delta^{18}\text{O}$ becomes higher during stadials and lower during interstadials, suggesting that they mainly reflect surface temperature changes over the core site rather than salinity variations due to the input of glacial meltwater with low $\delta^{18}\text{O}$. This differs from sites in the open North Atlantic Ocean where stadal-interstadial changes are often muted compared to Heinrich events, which are marked by sharp reductions in planktonic $\delta^{18}\text{O}$ [e.g., Hemming, 2004]. However, *N. pachyderma* (*s*) $\delta^{18}\text{O}$ is decoupled from the SST data in some cases, indicating that salinity changes were important over short time periods. The most prominent example of this is during the stadal following IS-13, when $\delta^{18}\text{O}$ becomes 0.8‰ lower than during the stadal onset. This event probably reflects the input of glacial meltwater during Heinrich 5 [Hemming, 2004].

[18] Interstadial decreases in the $\delta^{18}\text{O}$ record of *N. pachyderma* (*s*) lag increases in SST by approximately 100–200 years, particularly in IS-14, IS-12 and IS-11 (Figures 2 and 3). Interestingly, this lag occurs at the same time as maximum differences between the $\delta^{18}\text{O}$ composition of *N. pachyderma* (*s*) and *G. bulloides*, and the continued deposition of IRD into the early part of each interstadial. $\delta^{18}\text{O}$ values of each planktonic species converge during the warmest part of each interstadial before diverging again at the start of the following stadal. Similar differences in the response of the isotopic and SST proxies are not as clear during the cooling transitions at the end of each interstadial, when declines in percent *N. pachyderma* (*s*) and SST, and

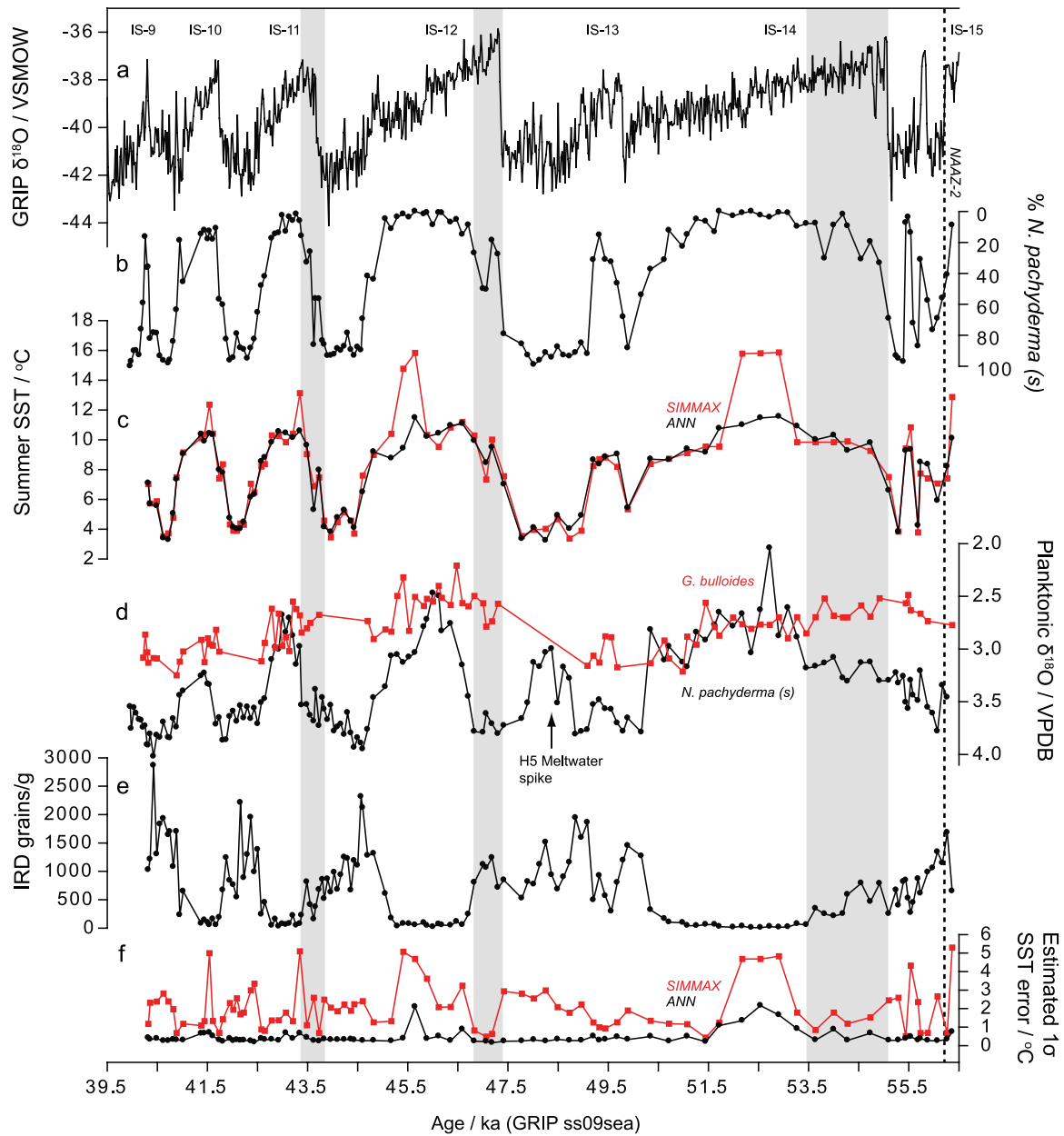


Figure 3. Proxy records from MD95-2006 and the GRIP ice core. Dansgaard-Oeschger interstadials are numbered IS 9-15, and the position of Heinrich 5 is marked. (a) GRIP $\delta^{18}\text{O}_{\text{ice}}$. (b) Percent *N. pachyderma* (s). (c) Summer SST (10 m depth) estimates, using the SIMMAX modern analog technique and the Artificial Neural Network technique. (d) *N. pachyderma* (s) and *G. bulloides* $\delta^{18}\text{O}$. (e) Abundance of IRD grains $>150\ \mu\text{m}$. (f) Estimated 1σ SST reconstruction errors for SIMMAX and ANN methods. Shaded bars denote periods of shallow surface water warming at the start of IS-11, -12 and -14. The dashed vertical line indicates the position of North Atlantic Ash Zone 2 in MD95-2006 and the GRIP ice core.

increases in *N. pachyderma* (s) $\delta^{18}\text{O}$ and IRD occur either within or in adjacent sampling intervals.

[19] During the period between IS-13 and the start of IS-12, no benthic data could be generated because *C. wuellerstorfi* was entirely absent. Benthic foraminifera $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ both display a series of gradual increases and decreases that have a very different signature compared to stadial-interstadial oscillations seen in the surface water proxy data (Figure

4). Benthic $\delta^{13}\text{C}$ varies between $\sim 1.1\text{‰}$, which is indicative of “modern” NADW sourced from open-ocean overturning at high latitudes [Kroopnick, 1985] and $<0.5\text{‰}$, which is similar to the values used to infer the presence of southern source deep waters at intermediate depths during the last glacial period [Oppo and Lehman, 1993; Elliot et al., 2002]. Benthic foraminifera $\delta^{18}\text{O}$ follows a similar pattern of gradual changes from $\sim 3.9\text{‰}$ at the start of stadial periods

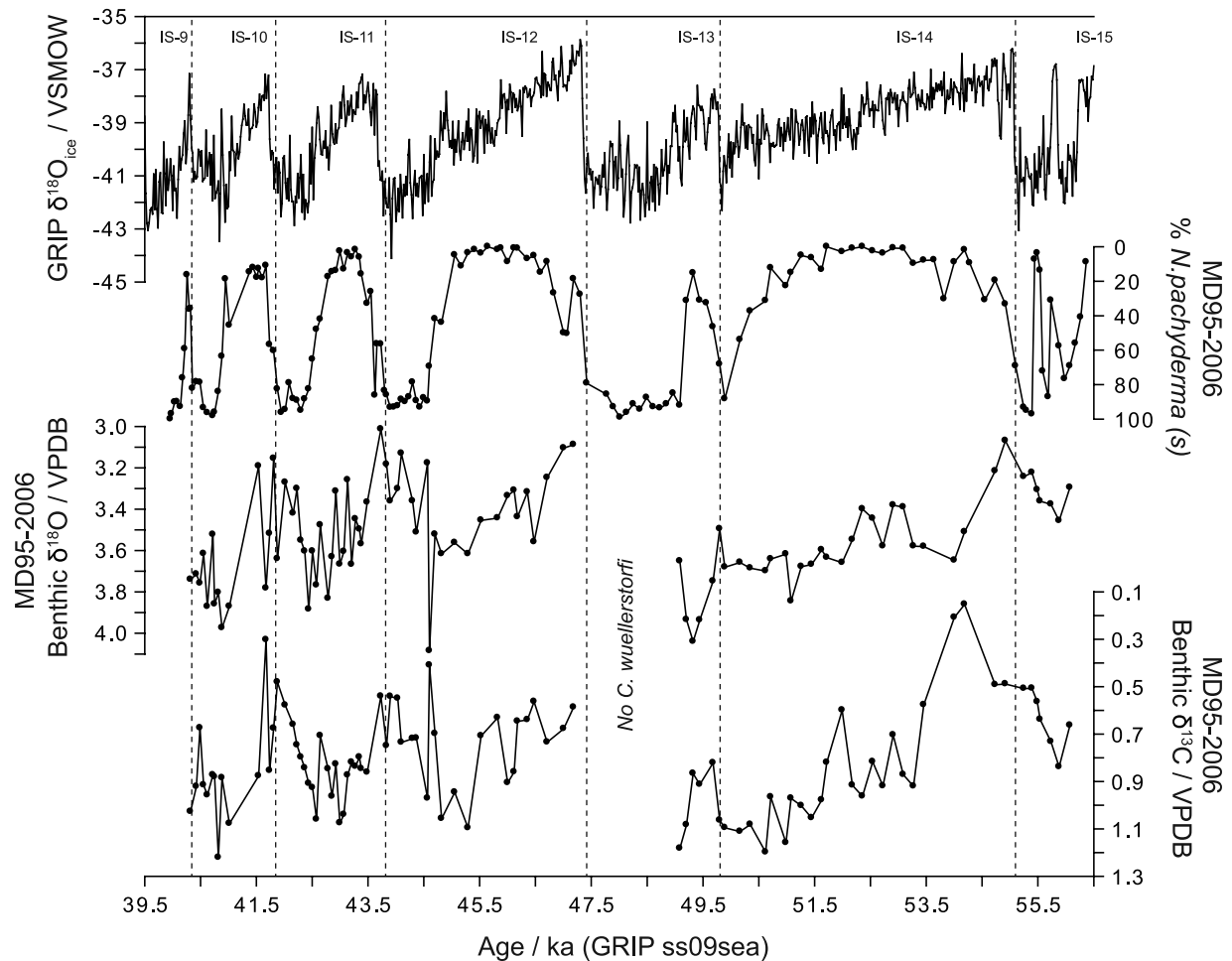


Figure 4. Comparison of benthic foraminifera $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ isotope data to percent *N. pachyderma* (*s*) in MD95-2006 and GRIP $\delta^{18}\text{O}_{\text{ice}}$. Stadal-interstadial transitions are marked with vertical dashed lines.

to $\sim 3.2\%$ during stadal-interstadial transitions. Shorter fluctuations in benthic $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ are also superimposed on the longer-term millennial-scale cycles.

5. Discussion

5.1. Millennial-Scale Fluctuations in the MOC

[20] Stadal-interstadial shifts in SST over the site of MD95-2006 together suggest that surface waters were alternately affected by the northward flow of warm salty waters entrained within the North Atlantic Drift during interstadials, and cold, fresh polar waters during stadials, when the polar front must have migrated south of the core site. These stadial migrations are well expressed in the percent abundance of *N. pachyderma* (*s*), which reached values $>95\%$ which is characteristic of modern subsurface waters north of the polar front [e.g., *Be and Tolderlund, 1971; Darling et al., 2006*]. This scenario is in good agreement with studies that have addressed the manner in which the activity of the Atlantic MOC changes between D-O stadials and interstadials [e.g., *Keigwin and Boyle, 1999; Hagen and Hald, 2002; Vautravers et al., 2004*], and model results that qualitatively equate interstadial with

interglacial boundary conditions and stadials with full glacial conditions [e.g., *Seidov and Maslin, 2001*].

[21] The different response times of the various surface water proxies to stadal-interstadial transitions suggests that the switch to a vigorous on mode of the Atlantic MOC did not occur instantaneously, but rather in two distinct stages. In the first stage, sharp decreases in the percent abundance of *N. pachyderma* (*s*) occur in tandem with rapid rises in SST. These changes occur before a distinct shift in *N. pachyderma* (*s*) $\delta^{18}\text{O}$ to lower values. In contrast, *G. bulloides*, which calcifies in near-surface waters during spring in the modern northeast Atlantic [*Ganssen and Kroon, 2000*], becomes more abundant in the early part of each interstadial, coincident with each rise in SST (Figure S1). Oxygen isotope values for *G. bulloides* are lower than for *N. pachyderma* (*s*) at these times. In the second stage of each interstadial transition, *N. pachyderma* (*s*) $\delta^{18}\text{O}$ decreases by $\sim 1\%$, and converges with $\delta^{18}\text{O}$ values for *G. bulloides*. At the same time, IRD abundances fall to negligible levels (Figure 3).

[22] Isotopic differences between different planktonic foraminifera have previously been interpreted as a proxy for surface water stratification between species living at

different depths in the water column [e.g., *Mulitza et al.*, 1997; *Mortyn and Charles*, 2003; *Rashid and Boyle*, 2007]. However, changes in the season during which planktonic species calcify may also have a pronounced effect on the $\delta^{18}\text{O}$ of foraminifera shells, particularly if seasonal changes generate warming close to the ecological thresholds for certain species [*Schmidt and Mulitza*, 2002]. In Figure 3, decreases in abundance of *N. pachyderma* (s) could be interpreted as an obvious sign that interstadial warming of the upper water column over MD95-2006 placed this species under stress, with the response that a large proportion of individuals were unable to survive. In contrast, the $\delta^{18}\text{O}$ isotopic composition of a small number of individuals remained virtually unchanged. Given that summer SST estimates for the early part of each interstadial are close to 10°C, which is warmer than the suggested upper temperature limit for the survival of *N. pachyderma* (s) [*Schmidt and Mulitza*, 2002], it would seem likely that these individuals were able to survive by calcifying either in deeper, colder waters, or during the winter season. However, winter SST estimates based on the ANN technique indicate temperatures of 7–8°C during peak interstadials, which is also at the ecological limit of *N. pachyderma* (s), whose $\delta^{18}\text{O}$ values may therefore primarily reflect a subsurface/deep temperature signal.

[23] From Figure 3, the convergence of *N. pachyderma* (s)/*G. bulloides* $\delta^{18}\text{O}$ is driven mainly by a 1‰ reduction in *N. pachyderma* (s) $\delta^{18}\text{O}$, suggesting that interstadial warming extended abruptly into deeper waters after the initial transition recorded by SST proxy data. Seasonal temperature differences between winter (*N. pachyderma* (s)) and spring (*G. bulloides*) over the modern day core site do not amount to more than 1°C at any level within the upper 150 m of the water column. Assuming that this pattern is indicative (though not exactly the same) of interstadial conditions, a 1‰ shift in *N. pachyderma* (s) $\delta^{18}\text{O}$ cannot be accounted for entirely by a change in the seasonal preferences of this species. However, recent studies have suggested that D-O stadials were characterized by a greater seasonal contrast, and some effect of this may be preserved in the difference between *G. bulloides* and *N. pachyderma* (s) $\delta^{18}\text{O}$. A slight change in depth habitat cannot be entirely ruled out as a partial influence on the *N. pachyderma* (s) $\delta^{18}\text{O}$ signal, because percent abundance of this species falls to <5% during IS-11, -12 and -14, and so may represent morphotypes of the shallower-dwelling *N. incompta* (or *N. pachyderma* (d)) [*Darling et al.*, 2006]. However, the consistency of the data through other interstadials when the abundance of *N. pachyderma* (s) remains above 5% means that this effect is probably small on the overall record. The extension of surface warming into deeper waters is also reflected in the delayed increase in the abundance of the deep-dwelling temperate foraminifera *Globorotalia inflata* after the initial surface warming during IS-11, -12 and -14 (supplementary material).¹

[24] The build up of saline surface waters at low latitudes has been hypothesized to trigger the resumption of the

interstadial mode of the Atlantic MOC through the creation of a salinity “imbalance” with relatively fresh northern polar waters [*Schmidt et al.*, 2004; 2006; *Levi et al.*, 2007]. The surface water data from MD95-2006 suggest that the early part of each interstadial was characterized by the rapid warming of near-surface waters, while deeper waters remained relatively cold. This warming may have been related to the sudden northward advection of warm, salty waters over the core site, in support of these previous studies. As each interstadial progressed, surface water warming extended into deeper (100–200 m) subsurface waters, at times when low abundances of IRD and high SST suggest “full” interstadial conditions (Figure 3).

5.2. Deep Water Mass Changes

[25] The deep water stable isotopic signals recorded in MD95-2006 follow a different millennial-scale pattern to those recorded in surface waters. This observation is not unique, in that numerous studies show similar differences between palaeoceanographic signals recorded in the surface and deep ocean [e.g., *Charles et al.*, 1996; *Shackleton et al.*, 2000; *Pahnke and Zahn.*, 2005], and have been interpreted to represent changes in deepwater mass properties driven by climatic changes in the different source regions. Within the hydrographically dynamic North Atlantic Ocean, the benthic $\delta^{18}\text{O}$ signal from Iberian Margin core MD95-2042 closely follows the pattern of surface air temperatures recorded in the Vostok ice core, and was suggested by *Shackleton et al.* [2000] to be forced primarily by global ice volume changes. However, recent work by *Skinner et al.* [2003] have shown that this region is also sensitive to changes in deepwater temperatures, driven in part by the changing influence of relatively warm UNADW and relatively cold Southern Component Waters (SCW) over the core site, and so cannot be uniquely interpreted in terms of whole-ocean $\delta^{18}\text{O}$. A large body of work has shown that the benthic foraminifera $\delta^{18}\text{O}$ signal in the northeast Atlantic Ocean is in fact sensitive to a range of hydrographic influences, which can make disentangling the benthic $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ data particularly difficult [e.g., *Siddall et al.*, 2008].

[26] The effect of sea level changes on the global reservoir of oceanic $\delta^{18}\text{O}$ during Marine Isotope Stage (MIS)-3 is difficult to assess at present, because of the problems of extracting this component from benthic $\delta^{18}\text{O}$ data, of obtaining sediment sequences with the requisite temporal resolution to resolve minor events, and because the mixing time of the oceans may limit the recording of short-duration sea level events in whole-ocean $\delta^{18}\text{O}$. Terrestrial sea level reconstructions during this period are also hindered both by the poor preservation of stratigraphic and geomorphological sea level indicators and by the precision of geochronological techniques applied to date them. Some attempts to estimate the magnitude of sea level fluctuations during the millennial-scale climate events of MIS-3 [e.g., *Lambeck and Chappell*, 2001; *Siddall et al.*, 2003; *Thompson and Goldstein*, 2005; *Arz et al.*, 2007] have suggested that a number of events may have occurred with an amplitude of 20–30 m. A comparison of the available data further suggests that periods of global sea level rise were centered

¹Auxiliary materials are available in the HTML. doi:10.1029/2008PA001595.

on D-O stadial events as seen in the various Greenland ice core records [Siddall *et al.*, 2008], an observation that strongly agrees with the timing of the changes in benthic $\delta^{18}\text{O}$ in MD95-2006, which decreases gradually by $\sim 0.7\text{--}0.9\text{‰}$ during each stadial period. Consequently, by assuming a scaling of $0.01\text{‰}/\text{m}$, we can arrive at a maximum estimate of 0.3‰ $\delta^{18}\text{O}$ change that could be due to changes in global ice volume. It is stressed that this is probably a maximum estimate given that most other benthic $\delta^{18}\text{O}$ records tend to resolve shifts associated with the pronounced Antarctic warm events A1–A4 rather than the relatively minor Greenland stadial events between IS-9 and IS-12 [Siddall *et al.*, 2008].

[27] Establishing a deep water temperature record for MD95-2006 is difficult because of the low abundance of benthic foraminifera available for Mg/Ca measurements. Instead, an approximate estimate of the effect of deepwater temperature changes on the benthic $\delta^{18}\text{O}$ record can be assessed by comparison to Iberian Margin core MD01-2444, where such data is available [Skinner and Elderfield, 2007]. This core is located at 2637 m water depth, but is bathed by the same UNADW water mass as MD95-2006. Assuming that the deepwater temperature signal in this core is mostly driven by changes in the depth of the transitional boundary between UNADW and SCW, which currently lies below the core site, we should expect that the observed temperature changes will be similar to those in MD95-2006, which is also sensitive to this boundary. This approach assumes that deepwater temperature changes over MD95-2006 are also predominantly related to UNADW/SCW boundary shifts, although it is acknowledged that other influences such as the presence of cold deepwater brines may also be important. Future deepwater temperature reconstructions from this region will help to address the importance of this issue. The temperature reconstructions from MD01-2444 are approximately in phase with GRIP $\delta^{18}\text{O}_{\text{ice}}$, with higher values during interstadials. This differs from the MD95-2006 benthic $\delta^{18}\text{O}$ record, which is clearly offset from GRIP $\delta^{18}\text{O}_{\text{ice}}$, with lowest values found at stadial-interstadial transitions. Despite the chronological differences, the range of deepwater temperature estimates ($\sim 1^\circ\text{C}$) in MD01-2444 equates to $\sim 0.25\text{‰}$ change in benthic $\delta^{18}\text{O}$ in MD95-2006 [Kim and O'Neil, 1997; Leng and Marshall, 2004], which cannot account for the full range of values we observe. Even when combined with a conservative 0.3‰ shift in whole-ocean $\delta^{18}\text{O}$ associated with changes in global ice volume, $\sim 0.15\text{‰}$ of the observed amplitude of benthic $\delta^{18}\text{O}$ still has to be accounted for by other influences (twice the estimated analytical error in $\delta^{18}\text{O}$ measurements).

[28] It is possible that the presence of SCW over the study site (linked to reductions in deepwater temperatures in MD01-2444 [Skinner and Elderfield, 2007]) could additionally alter $\delta^{18}\text{O}$ by lowering the CO_3^{2-} concentration of bottom waters over the core site [Spero *et al.*, 1997]. However, a $\sim 70\text{ }\mu\text{mol}/\text{kg}$ change in CO_3^{2-} ion concentration would be required to explain the 0.15‰ residual benthic $\delta^{18}\text{O}$ observed once whole-ocean and temperature effects have been accounted for [Spero *et al.*, 1997], when the present-day contrast between Antarctic Bottom Water

and NADW end-members is only $\sim 40\text{ }\mu\text{mol}/\text{kg}$ [Broecker, 2003b].

[29] The transfer of sea ice derived brines to the deep ocean is one explanation to account for the $\sim 0.1\text{--}0.3\text{‰}$ of the benthic $\delta^{18}\text{O}$ data in MD95-2006 that cannot be accounted for by whole-ocean or temperature changes. The formation of sea ice over freshwater lenses introduced by meltwater and melting icebergs can concentrate salts in surface waters through the process of brine rejection [Craig and Gordon, 1965]. This process inhibits the air-sea exchange of oxygen in surface waters, therefore allowing saline/dense waters to retain their low initial $\delta^{18}\text{O}$, which are transferred through the water column once a salinity threshold is reached and density-driven overturning takes place [e.g., Dokken and Jansen, 1999; van Kreveld *et al.*, 2000; Hillaire-Marcel and De Vernal, 2008]. Furthermore, sea ice inhibits the growth of primary producers, limiting the uptake of inorganic ^{12}C and causing low $\delta^{13}\text{C}$ in the resulting brines. This pattern is opposite to the carbon isotopic signature of UNADW, which tends to have high ($\sim 1\text{‰}$) $\delta^{13}\text{C}$ after being stripped of nutrients during exposure at the ocean surface prior to overturning [e.g., Kroopnick, 1985; Duplessy *et al.*, 1988].

[30] This pattern is seen in the benthic $\delta^{13}\text{C}$ data from MD95-2006. If lower benthic $\delta^{18}\text{O}$ were being driven purely by an increase in “warmer” UNADW over the core site, then a coeval increase in $\delta^{13}\text{C}$ could also be expected. However, the benthic $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ data vary in phase with each other, decreasing to minimum values during stadial interstadial transitions and increasing again through each interstadial period (Figure 4). The highest amplitude stadial-interstadial decrease in benthic $\delta^{18}\text{O}$ ($\sim 0.9\text{‰}$) appears to occur prior to IS-12, and cannot be explained by analytical error once a total range of $\sim 0.55\text{‰}$ related to whole ocean and temperature effects has been removed. However, this change occurs over a gap in the benthic isotope data and consequently may be influenced by data outliers. A large negative benthic $\delta^{18}\text{O}$ excursion has also been identified during this interval in nearby core NA87-22, and attributed to the transfer of brines into the northeast Atlantic during Heinrich 5 [Vidal *et al.*, 1998]. This event may have also influenced MD95-2006, which was potentially also sensitive to more minor brine “events” during the other D-O stadials. The formation of deepwater brines related to periods of expanded high-latitude sea ice has previously been suggested to explain negative benthic $\delta^{18}\text{O}$ excursions during D-O events in the Nordic and Irminger Seas [Dokken and Jansen, 1999; van Kreveld *et al.*, 2000] and during Heinrich Events in the North Atlantic Ocean [Vidal *et al.*, 1998]. Low benthic $\delta^{18}\text{O}$ signatures below 2000 m in the northeast Atlantic during the last deglaciation suggest that Nordic Sea brine waters may have replaced deepwater formed from open-ocean convection [Meland *et al.*, 2008]. An increase in the influence of open-ocean convection at these depths during the Bolling-Allerod is analogous to the changes observed during interstadials in MD95-2006, suggesting that the mechanisms we identify were not unique to the MIS-3 D-O events.

[31] Increases in the volume of deepwater brines introduced to the subpolar North Atlantic Ocean during stadial

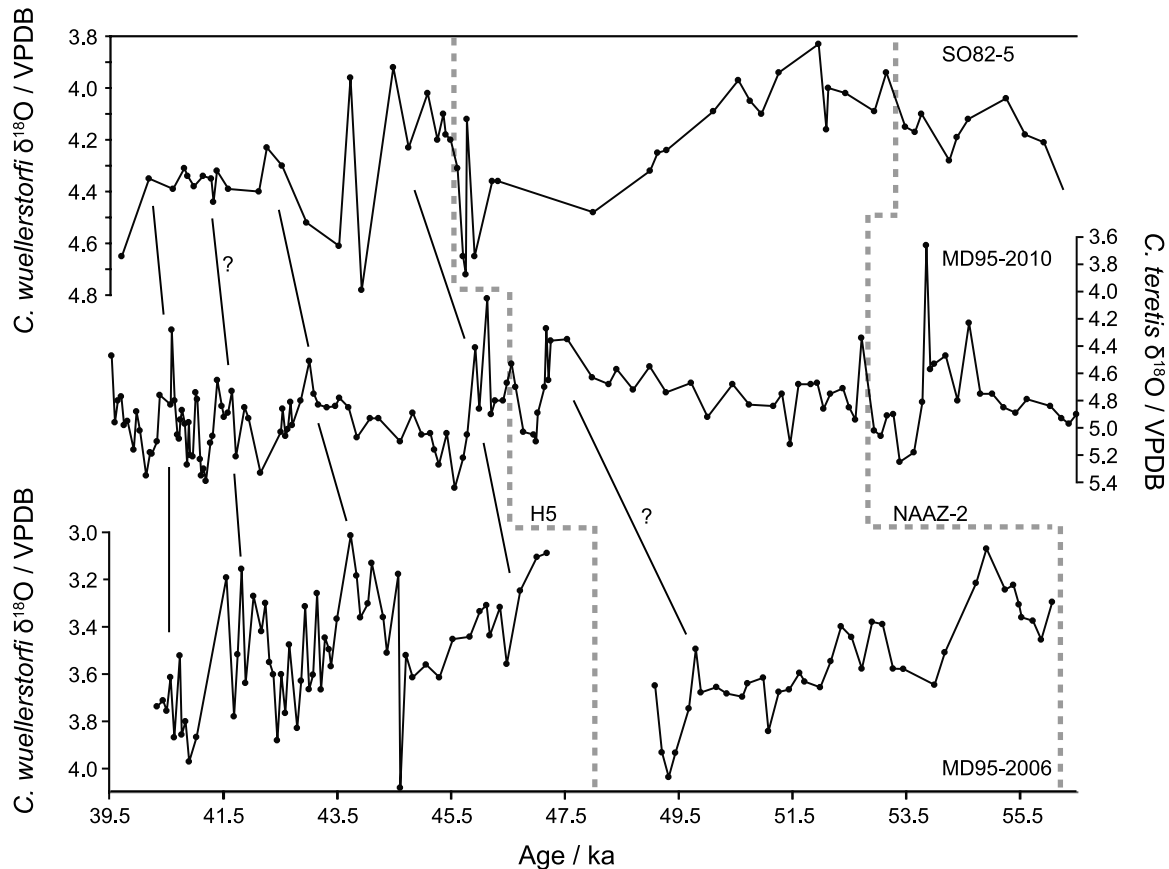


Figure 5. Comparison of benthic $\delta^{18}\text{O}$ records for the period 39.5–56.5 ka from MD95-2006 (this study), MD95-2010 [Dokken and Jansen, 1999] and SO82-5 [van Kreveld et al., 2000]. Each record is plotted on its original age model, and proposed correlations (black lines) are guided by the positions of North Atlantic Ash Zone 2 (NAAZ-2) and IRD peaks associated with Heinrich 5 (thick dashed lines) in each core.

periods would require a large expansion of sea ice in the eastern Nordic Seas and northeast Atlantic Ocean at the same time that the influence of deep waters formed by open-ocean convection were reduced. There are few reliable sea ice reconstructions for MIS 3 in the Nordic Seas, but estimates of sea ice extents for the LGM relative to present based on dinocyst [e.g., De Vernal et al., 2006] and foraminifera [Pflaumann et al., 2003] assemblages can be tentatively used as qualitative analogs for stadial and interstadial conditions. These reconstructions indicate that the expansion of seasonal sea ice cover during the LGM may have been substantial, reaching a limit close to the latitude of MD95-2006. Recent studies have also highlighted the possibility that large increases in seasonality may have been responsible for the large mean annual temperature declines recorded in Greenland ice cores during stadial periods [Denton et al., 2005; Broecker, 2007]. Extremely cold winters would generate a large expansion of winter sea ice in the Nordic Seas, which would inhibit the transfer of surface ocean heat to the atmosphere [Li et al., 2005]. Large increases in winter sea ice would help to generate the

deepwater brines which are interpreted from the MD95-2006 benthic isotope record.

[32] A comparison of the benthic $\delta^{18}\text{O}$ data from cores MD95-2006, MD95-2010 and SO82-5 [Dokken and Jansen, 1999; van Kreveld et al., 2000] shown in Figure 5 suggests that the same brine events recorded during D-O stadials in the Nordic Seas may also be recorded in the deep northeast Atlantic Ocean after ~48 ka B.P. (GRIP ss09sea timescale). While it is possible that a component of each benthic isotopic excursion in MD95-2006 may be related to the influence of locally derived brines from the Hebridean Shelf, the correlation of these events to those in MD95-2010 and SO82-5 suggests that the data presented here is most likely indicative of regional-scale processes linked to the Nordic Sea overflow routes into the North Atlantic Ocean. The data therefore supports the arguments of Dokken and Jansen [1999] and van Kreveld et al. [2000] that the southward export of deepwater brines from the Nordic Seas during periods of expanded sea ice during D-O stadials may have helped to generate D-O warming by forming a weak MOC. This weak circulation state may have in turn helped to destabilize the build up of warm waters in the subtropics

[e.g., Schmidt et al., 2006] whose northward transfer within the NAC may be represented by the initial shallow layer warming seen in the planktonic $\delta^{18}\text{O}$ data of MD95-2006.

6. Conclusion

[33] Surface water hydrography over the site of giant piston core MD95-2006 displays high-amplitude variability that can be linked to surface temperature changes documented in the GRIP $\delta^{18}\text{O}_{\text{ice}}$ record between 39.5 and 57.5 ka B.P. Rapid surface water warming occurs at the start of each interstadial in the upper part of the water column, and extends into deeper water as each interstadial develops. Benthic $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ isotopes vary over millennial timescales, but exhibit gradual increases and decreases that are different in timing and pattern from surface water data. The observed $\sim 0.7\text{‰}$ shifts in benthic $\delta^{18}\text{O}$ cannot be explained by assuming a maximum $\delta^{18}\text{O}$ influence of 0.3‰ related to global ice volume changes and 0.25‰ related to bottom water temperature changes, and call for additional regional influences on the $\delta^{18}\text{O}$ signal. With reference to previous studies [Dokken and Jansen, 1999; van Kreveld et al., 2000; Meland et al., 2008], it is suggested that this residual signal is derived from the influence of low-salinity deepwater brines that formed

during periods of expansive winter sea ice at high northern latitudes during D-O stadials. This interpretation is supported both by the observation of lower benthic $\delta^{13}\text{C}$ coincident with periods of lower benthic $\delta^{18}\text{O}$, and by preliminary correlations to suggested “brine events” in cores NA87-22 [Vidal et al., 1998], MD95-2010 [Dokken and Jansen, 1999], and SO85-2 [van Kreveld et al., 2000]. The coincidence of minimum benthic $\delta^{13}\text{C}/\delta^{18}\text{O}$ with periods of rapid surface water warming over the site of MD95-2006 supports the suggested role of deepwater brines in helping to trigger the abrupt resumption of the warm mode of the MOC by helping to destabilize the build up of warm, saline surface waters in the subtropical North Atlantic [Schmidt et al., 2004, 2006].

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