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Surface and deep-water hydrography on Gardar Drift (Iceland Basin) during the last interglacial period

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ABSTRACT

Changes in surface and deep-water hydrography were inferred from variations in stable isotopes and sortable silt mean grain size, respectively, on the southern Gardar Drift in the subpolar North Atlantic. The bathymetric δ^{13}C gradient during the penultimate glaciation was similar to the last glaciation with high-δ^{13}C Glacial North Atlantic Intermediate water above ~2000 m, and low- δ^{13}C water derived from the Southern Ocean below. During Termination II, low-δ^{13}C water was present throughout the water column with minimum values at intermediate depths (~1500–2000 m) and below 3000 m. This pattern continued well into the early part of the Last Interglacial (LIG) period. Sortable silt mean size at 3275 m suggests that deep-water circulation on Gardar Drift was relatively weak during the earliest part of the LIG (128 to 124.5 ka) when planktonic δ^{18}O was at a minimum, reflecting warming and/or reduced salinity. We suggest that low- δ^{13}C water and slow current speed on Gardar Drift during the early part of the LIG was related to increased melt water fluxes to the Nordic Seas during peak boreal summer insolation, which decreased the flux and/or density of overflow to the North Atlantic. The resumption of the typical interglacial pattern of strong, well-ventilated Iceland Scotland Overflow Water was delayed until ~124 ka. These changes may have affected Atlantic Meridional Overturning Circulation.

1. Introduction

The climate of the last interglacial period, corresponding to Marine Isotope Stage (MIS) 5e, was slightly warmer and sea level was several meters higher than today (Overpeck et al., 2006; Jansen et al., 2007; Rohling et al., 2008). Although orbital parameters differed, MIS 5e is commonly used for comparison to the Holocene and for providing a reference of interglacial climate variability in the absence of anthropogenic forcing. A major shortcoming for the study of the LIG is there is yet no complete, undisturbed section of the LIG recovered in Greenland ice cores (North Greenland Ice Core Project Members, 2004). Drilling has now begun in northwestern Greenland with the goal of recovering a complete ice core record of the last interglacial (NEEM — North Greenland Eemian Ice Drilling; http://neem.ku.dk/). An important challenge for paleoceanographers is to identify complementary marine sections with sufficiently high sedimentation rates for comparison with the anticipated NEEM ice core.

Here we report on an expanded section of the last interglacial period recovered on the Gardar Drift south of Iceland (Fig. 1). A unique feature of Integrated Ocean Drilling Program (IODP) Site U1304 is that it possesses an expanded section of the "MIS 5e isotopic plateau" (128 to 116 ka) (Shackleton et al., 2003) and MIS 5d (116 to 105 ka). The sediment deposited in this period includes the intermittent occurrence of laminated diatom mats (LDM) and oozes (LDO) (Fig. 2) (Expedition 303 Scientists, 2006), which were rapidly deposited at the North Subarctic Front (NSAF) (Bodén and Backman, 1996). The high tensile strength of the LDM/LDO suppresses benthic activity and hinders bioturbation in the underlying sediment, thereby offering a rare opportunity to study the last interglacial period and its demise into substage 5d at ultra high resolution (i.e., multi-decadal to centennial time scales). Furthermore, the water depth of Site U1304 at 3076 m extends the depth range of previous reconstructions of bathymetric δ^{13}C from cores taken from North Atlantic sediment drifts (Oppo et al., 1997).

We measured stable isotope ratios of multiple species of foraminifera to reconstruct changes in surface and deep-water hydrography

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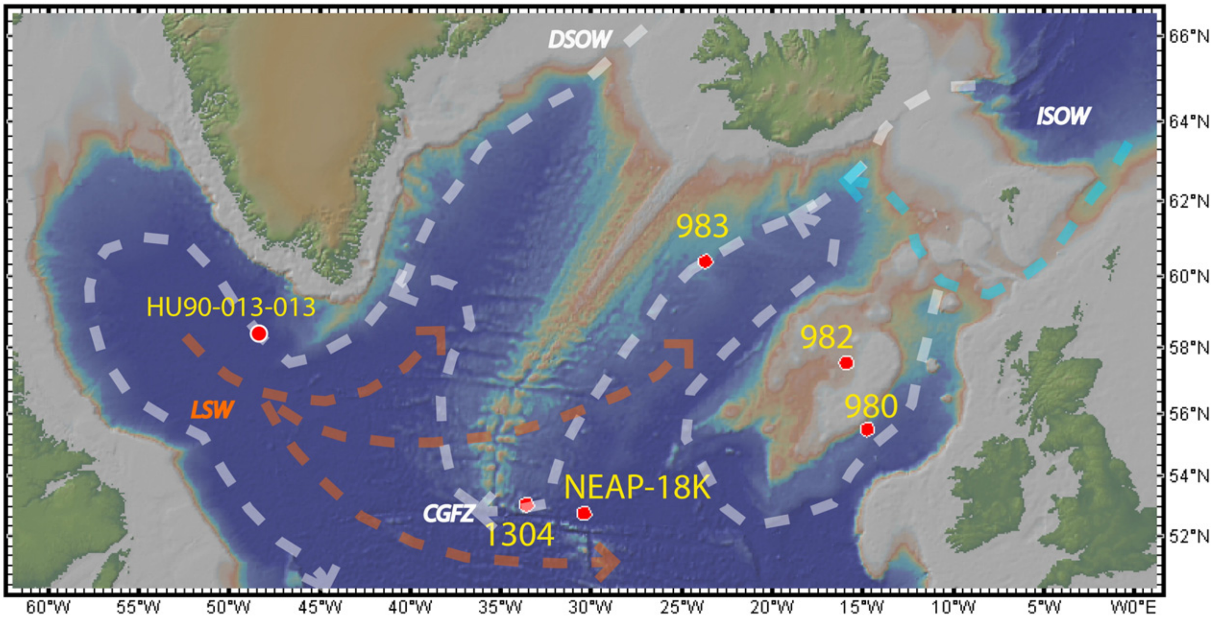

Fig. 1. Location of sites discussed in text relative to surface and deep-water circulation of the subpolar North Atlantic. CGFZ: Charlie Gibbs Fracture Zone; LSW: Labrador Sea Water; ISOW: Iceland Scotland Overflow Water; DSOW: Denmark Strait Overflow Water.

for the period from MIS 6 to substage 5d. Planktonic $\delta^{18}O$ of foraminifera from Gardar Sediment Drift record the nature of surface inflow to the Nordic Seas via the North Atlantic Current, whereas benthic $\delta^{13}C$ monitors the ventilation of deep water. By comparing our planktonic and benthic isotope records with previously published paleocurrent reconstruction of flow speed, changes using sortable silt ($\overline{SS}$, the mean grain size of the 10–63 µm terrigenous sediment subfraction), we examine the timing and linkages between surface water inflow and deep-water outflow over the Iceland–Scotland Ridge during the LIG.

2. Site locations and hydrography

The Gardar Drift is an elongated contourite deposited along the eastern flank of the Reykjanes Ridge. It was formed by the interaction of Iceland Scotland Overflow Water (ISOW), flowing as a Deep Western Boundary Current (DWBC), and local topography (Bianchi and McCave, 2000). The drift deposit is oriented NE–SW and extends for about 1100 km, increasing in water depth from ~1400 m water depth in the NE to >3000 m in the SW near the Charlie Gibbs Fracture Zone (CGFZ) (Fig. 1) (Bianchi and McCave, 2000).

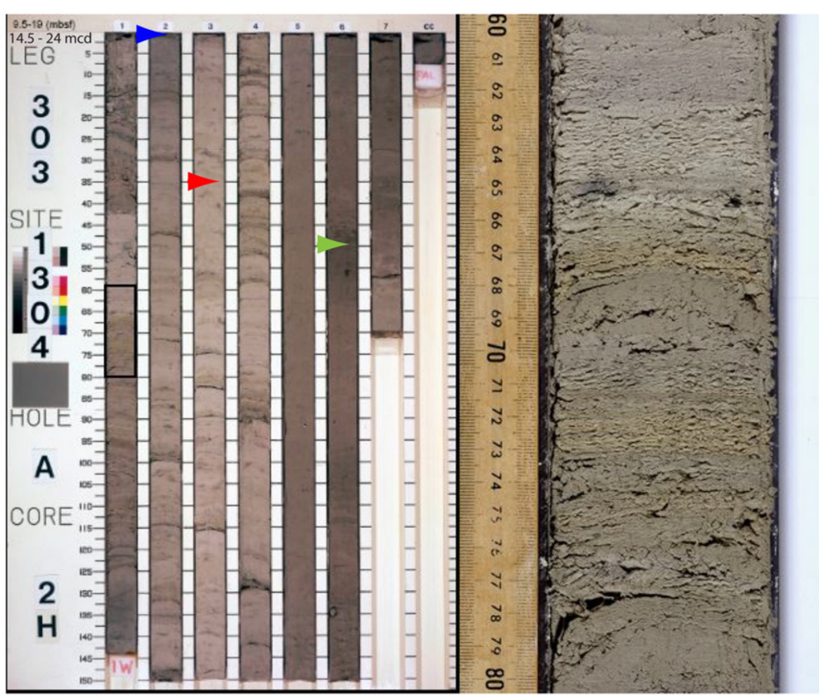

Fig. 2. Left: Core photograph of Core 1304A-2H. The MIS 5e plateau (128 to 116 ka) begins at section 6, 50 cm (green arrow) and ends at section 3, 35 cm (red arrow). Right: close-up for section 1, 60–80 cm indicated by rectangle on core photograph showing an example of diatom layers interbedded with carbonate-rich sediment. After Expedition 303 Shipboard Scientific Party (2006). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

The Iceland–Scotland ridge is the site of one of the overflows of water from the Nordic Seas to the North Atlantic, the other being Denmark Strait (Fig. 1). ISOW forms through convection in the Norwegian Greenland Sea via cooling of warm surface waters transported north by the North Atlantic Current (NAC). ISOW entrains ambient water, mainly Labrador Sea Water (LSW), to form a descending plume that transports approximately 6 Sv (1 Sv $=10^6$ m^3 s^{-1}) of water through the CGFZ (Smethie et al., 2007). Along with Denmark Strait Overflow Water (DSOW) and LSW, CGFZ water contributes an important component of North Atlantic Deep Water (NADW) that constitutes the lower limb of the Atlantic Meridional Overturning Circulation (AMOC) (Quadfasel and Kase, 2007).

Found below ISOW on the southern Gardar Drift is Lower Deep Water (LDW), which is sourced from the Southern Ocean as reflected by its high silica content (McCartney, 1992; Reid, 1994). During the last glacial period, the source area for deep-water formation moved south of the Greenland–Scotland Ridge and NADW was replaced by a shallower water mass known as Glacial North Atlantic Intermediate Water (GNAIW) above ~2000 m water depth (Oppo and Lehman, 1993; Curry and Oppo, 2005). Below 2000 m, deep water in the North Atlantic was dominated by a water mass originating in the Southern Ocean. Previous studies have shown the Gardar Drift preserves a detailed record of variations in ISOW flow during the late Pleistocene (Hall et al., 1998; Bianchi and McCave, 1999; Bianchi and McCave, 2000; McCave and Hall, 2006).

We studied Site U1304 located on the southern edge of the Gardar Drift in 3082 m water depth close to CGFZ (Fig. 1) (Channell et al., 2006). Site U1304 is compared with the records from two other sites from the same drift deposit. Site 983 is positioned on the northern part of the drift in 1983 m water depth. It is bathed by a mixture of ISOW and Labrador Sea Water (LSW) today (Schmitz and McCartney, 1993; Reid, 1994). Kasten core NEAP-18K is located ~210 km to the NW of Site U1304 and in slightly deeper water (3275 m) (Hall et al., 1998) (Fig. 1). Both sites U1304 and NEAP-18K are influenced today mostly by ISOW with a minor component of LDW. During the last glaciation, Site U1304 and NEAP-18K were bathed by LDW sourced from the Southern Ocean, whereas Site 983 was positioned near the boundary of GNAIW and LDW (Oppo and Lehman, 1993).

3. Methods

The composite section of Site U1304 was sampled at a constant 5-cm spacing between 24.52 and 15.47 mcd, corresponding to the MIS 6 through MIS 5d interval. The last interglacial and transition to MIS 5d is contained in Core 1304A-2H-1 (Fig. 2). The mean sedimentation rate for MIS 5e and 5d is 40 cm ka^{-1}, which results in an average sampling frequency of 125 years.

Stable oxygen (δ^{18}O) and carbon (δ^{13}C) isotope ratios were measured on three species of planktonic foraminifera (*Globigerina bulloides*, *Neogloboquadrina incompta* (formerly *Neogloboquadrina pachyderma* dextral), and *Globorotalia inflata*) and the epi-benthic foraminfera *Cibicidoides wuellerstorfi*. All isotope measurements were made using a Finnigan-MAT 252 isotope ratio mass spectrometer coupled with a Kiel III carbonate preparation device, and are reported in standard delta notation relative to Vienna Pee Dee Belemnite (VPDB) using NBS-19 for calibration. The estimated analytical error is better than ±0.1‰ for both δ^{18}O and δ^{13}C.

Sediment lightness was measured at 2-cm intervals shortly after cores were split using a Minolta CM1500 Spectrophotometer (Expedition 303 Scientists, 2006). Weight percent $CaCO_3$ was determined using a UIC (Coulometrics) 5011 CO_2 coulometer coupled with an AutoMateFX carbonate prep system. Magnetic susceptibility was measured on u-channel subcores at 1-cm resolution using a custom-built instrument in the paleomagnetic laboratory at the University of Florida (Thomas et al., 2003).

4. Results

4.1. Oxygen isotopes

Termination II, the glacial to interglacial transition from MIS 6 to 5, is marked by a rapid decrease in δ^{18}O in all species of foraminifera at ~22.7 mcd (Fig. 3). Following this event, benthic δ^{18}O (*C. wuellerstorfi*) reveals a broad plateau of unchanging values for 4.9 m from 22.6 to 17.7 mcd during MIS 5e (Fig. 3). Benthic δ^{18}O values average 2.69 ±0.08‰ ($n=86$) and variation is within analytical error. Greater variability is expressed in planktonic δ^{18}O during the plateau interval. The duration of minimum δ^{18}O during MIS 5e varies among planktonic species. The δ^{18}O minimum is briefest in *G. bulloides*, slightly longer in *G. inflata*, and longest-lived in *N. pachyderma* (dex). A second δ^{18}O minimum occurs at the substage 5e/5d boundary in *G. bulloides* and *G. inflata* only. All planktonic δ^{18}O records increase during MIS 5d towards a distinct maximum at 16 mcd.

4.2. Carbon isotopes

Benthic δ^{13}C ranges from −0.4 to 1.2‰. No significant change occurs in benthic δ^{13}C across Termination II as δ^{13}C remains low during the early part of MIS 5e (Fig. 3). Benthic δ^{13}C increases at 21.3 mcd and remains relatively high until the end of the MIS 5e plateau (116 ka) when δ^{13}C shows two minima near the MIS 5e/5d transition. The lows in benthic δ^{13}C at the beginning of MIS 5e and near the substage 5e/5d boundary coincide with minima in the δ^{18}O of *G. bulloides* and *G. inflata* (Fig. 3). Benthic δ^{13}C increases to maximum values of 1.2‰ at 15.5 mcd in MIS 5d.

4.3. Lithostratigraphy

MIS 6 sediment is composed of gray clay with wt.% $CaCO_3$ averaging ~20% and high, but variable, magnetic susceptibility (Fig. 3). A large peak in ice-rafted detritus (IRD) occurs at 22.5 mcd on Termination II, marked by carbonate content approaching 0%. Following this event (Heinrich Event 11) (Heinrich, 1988), wt.% $CaCO_3$ rises to 39% at 22 mcd and then decreases to 32% at 21.5 mcd. The earliest part of MIS 5e is generally marked by darker sediment with higher magnetic susceptibility and lower carbonate content than the latter part of MIS 5e. Lightness and $CaCO_3$ content increase and magnetic susceptibility decreases from 22 to 20 mcd with another step at 19 mcd, reaching a maximum wt.% $CaCO_3$ value of 64%. Sediment lightness and $CaCO_3$ remain high and magnetic susceptibility low until the end of the MIS 5e plateau (17.7 mcd). During the transition to MIS 5d, $CaCO_3$ content declines and magnetic susceptibility increases, reaching a peak at ~16 mcd corresponding to event C24 (McManus et al., 1994). Diatom layers become common in the section at 20.5 mcd (base of 1308A-2H-4; Fig. 2).

4.4. Chronology and sedimentation rates

We adopt the definition of the CAPE Project Members (2006) who took the LIG to represent the interval of minimum ice volume when benthic δ^{18}O values were low and relatively constant during the MIS 5e plateau. We note, however, that the terrestrial record of the last interglacial (i.e., Eemian in Europe) differs in timing compared to the marine record (Shackleton et al., 2002, 2003).

Shackleton et al. (2002, 2003) developed a radiometric time scale for the LIG that is independent of astronomical calibration. They estimated an age of 128 ka for the start of the MIS 5e plateau, and an age of 116 ka for the end of the plateau (Shackleton et al., 2002, 2003). We adopt these calibration points and use another at the peak of MIS 5d (107 ka) corresponding to C24 (McManus et al., 1994) or Greenland Stadial (GS) 24 (Rousseau et al., 2006). Sedimentation rates at Site U1304 average 40 cm ka^{-1} during the MIS 5e plateau interval.

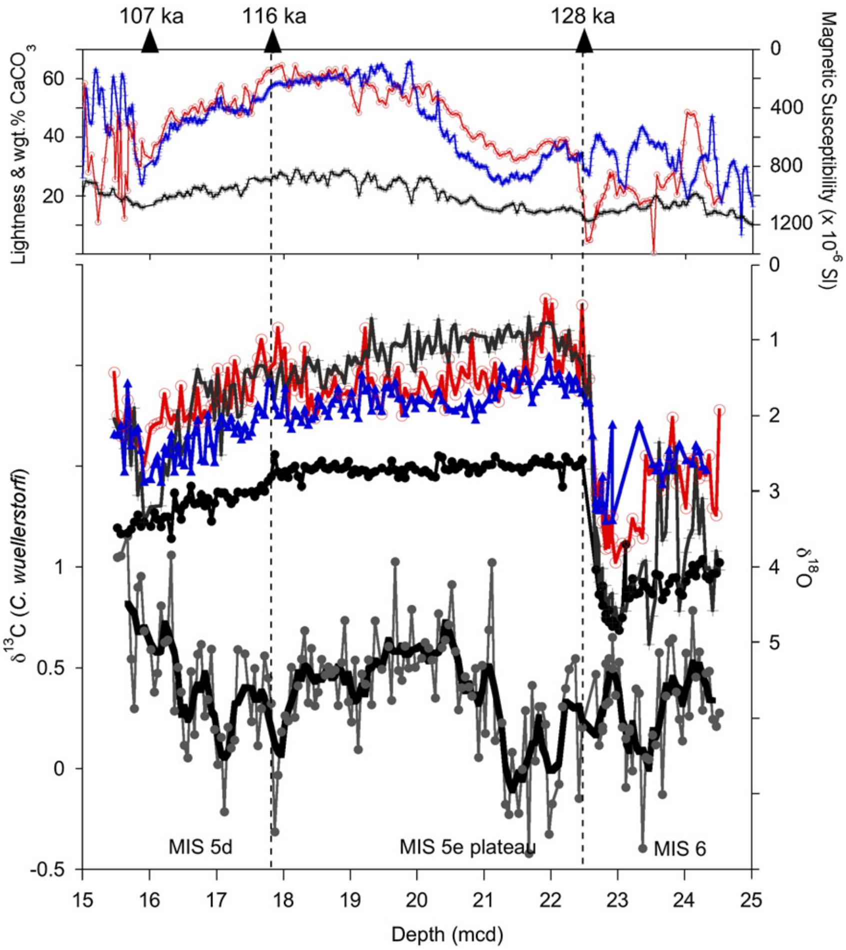

Fig. 3. Site U1304, Lower panel: oxygen isotope results measured on multiple species of planktonic foraminifera (*G. bulloides* — red open circles; *G. inflata* — blue diamonds; *N. pachyderma* (dex) — gray crosses). $\delta^{18}O$ (black circles) and $\delta^{13}C$ (gray circles) of the benthic foraminifer *C.wuellerstorfi*. Bold line represents a 5-point running mean of benthic $\delta^{13}C$. Vertical dashed lines designate a 4.88-m section of near constant benthic $\delta^{18}O$ corresponding to the MIS 5e plateau (~128 to 116 ka). Upper panel: Color lightness (black), weight percent $CaCO_3$ (red), and magnetic susceptibility (blue). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

5. Discussion

5.1. Comparison with other records

We compare the record from Site U1304 with previously published results from two sites on Gardar Drift: ODP Site 983 (Channell et al., 1997) and NEAP-18K (Hall et al., 1998).

Oxygen isotopes of *Cibicidoides* and *G. bulloides* at Site 983 show uniformly low values from 15.7 to 13.1 mcd. Sedimentation rates average 17 cm ka^{-1} during the MIS 5e plateau. Similar to Site U1304, the early part of the LIG is marked by lower carbonate content than the latter part. The water depth of Site 983 (1983 m) is more than a km shallower than Site U1304 (3082 m), yet the benthic $\delta^{13}C$ record shows a similar minimum at Termination II and into the earliest part of the MIS 5e plateau (Fig. 4). Based on a constant sedimentation rate between the start and end of the MIS 5e plateau at Site U1304, we estimate this low- $\delta^{13}C$ water mass persisted for ~3 ka into the LIG until ~124.5 ka. A similar calculation for Site 983 yields a slightly shorter duration lasting until ~126 ka.

The bottom of core NEAP-18K just reached Termination II as Heinrich event 11 occurs at the base of the record (Chapman and Shackleton, 1999). The MIS 5e plateau at NEAP-18K spans 80 cm with sedimentation rate averaging 6.6 cm ka^{-1} (Fig. 5). The $\delta^{18}O$ record of *G. bulloides* at NEAP-18K shows the same pattern as Site U1304, with lowest values at the beginning of the MIS 5e plateau associated with peak summer boreal insolation (Fig. 6A,B).

The benthic $\delta^{13}C$ record at NEAP-18K shows similar trends as those at Site U1304, but NEAP-18K has consistently higher values throughout MIS 5e and 5d (Fig. 6C). This is unlikely to be the result of the two sites being bathed by different water masses because they are only 210 km apart and separated by ~200 m water depth. The Site U1304 benthic $\delta^{13}C$ record shows several minima (with values approaching −0.5‰) that are not apparent at NEAP-18K. The lower sedimentation rate (~6 cm ka^{-1}) and greater mixing by bioturbation at NEAP-18K may filter the short-term $\delta^{13}C$ variations, but this shouldn't result in a constant offset between the two records. Alternatively, low benthic $\delta^{13}C$ values at Site U1304 may reflect a phytodetritus effect related to the increased flux of organic matter to the sediment-water interface. Mackensen et al (1993) proposed that a high flux of organic matter to the sediment-water interface can create a phytodetritus layer with low $\delta^{13}C$ derived from the oxidation of ^{13}C-depleted organic matter. Epifaunal benthic foraminifera, such as *C. wuellerstorfi*, living in this microhabitat might record $\delta^{13}C$ values lower than ambient bottom water. Organic matter associated with increased diatom productivity is not likely the cause for the low benthic $\delta^{13}C$ values in the early LIG, however, because the common occurrence of diatom ooze does not begin until ~20.5 mcd, which is well above the interval of lowest $\delta^{13}C$ values (Fig. 3).

The benthic $\delta^{13}C$ records at Sites 983 and U1304 show nearly identical patterns across Termination II, favoring the interpretation that low benthic $\delta^{13}C$ reflects variations in deep-water ventilation. Furthermore, the similarity of the Site U1304 benthic $\delta^{13}C$ and the $\overline{SS}$

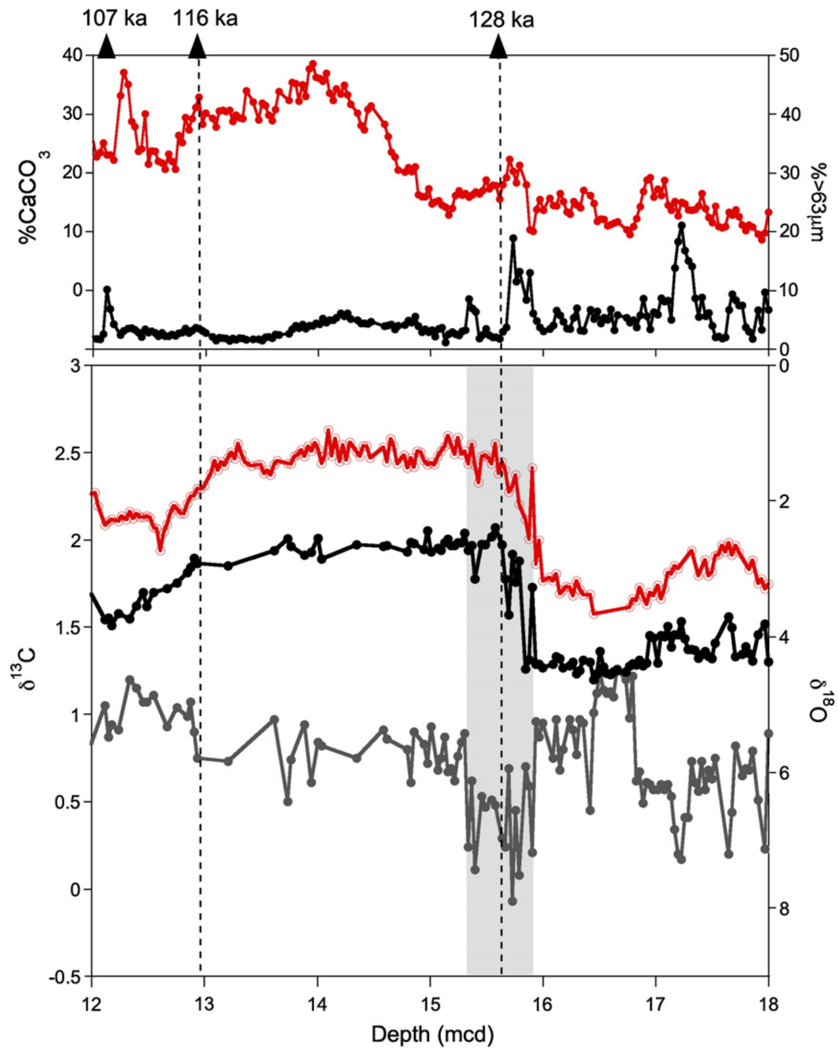

Fig. 4. Site 983, Bottom: Benthic $\delta^{13}C$ of *C. wuellerstorfi* (gray circles) and $\delta^{18}O$ of *C.wuellerstorfi* (black closed circles) and *G. bulloides* (red open circles). Top: weight % of the >63µm fraction of the sediment and weight % $CaCO_3$. Vertical dashed lines designate a 2.6-m section of the MIS 5e plateau (~128 to 116 ka). Gray shading indicates intervals of low benthic $\delta^{13}C$. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

near-bottom flow speed variations at NEAP-18K suggest both are recording circulation changes (Fig. 6D) (Hall et al., 1998). Each of the minima in benthic $\delta^{13}C$ at Site U1304 coincides with a decrease in $\overline{SS}$ indicating reduced vigor of deep water on the southern Gardar Drift. Unlike nutrient related hydrographic proxies such as benthic $\delta^{13}C$, $\overline{SS}$ is a physical proxy of relative current speed that is unaffected by biological processes. At NEAP-18K, $\overline{SS}$ suggests that flow speeds are lowest during the earliest part of the MIS 5e plateau and at the transition from MIS 5e to 5d (Figs. 5 and 6D). The $\overline{SS}$ record was compared to a very similar record in the N. American Basin on Bermuda Rise at 4462 m by Hall et al. (1998), They noted that the T-II/early 5e slow speeds corresponded to the nutrient-rich (high Cd/Ca) water of southern origin identified by Adkins et al. (1997). Thus the coordinated behavior of water masses and their flow speed was likely both ocean-wide and controlled from the Southern Ocean source.

5.2. Bathymetric $\delta^{13}C$ gradients

Following the lead of Oppo et al. (1997), we constructed bathymetric $\delta^{13}C$ gradients for several time slices including MIS 6 (at maximum benthic $\delta^{18}O$ value), Termination II, and early and late MIS5e (Fig. 7). Oppo et al. (1997) used sites spanning a depth range from 1451 to 2658 m for their reconstruction based mainly on cores from the Bjorn Drift. To these data we add IODP Site U1304 (3076 m) and 983 (1983 m) from Gardar Drift and Site 982 (1145 m) from the Rockhall Plateau.

The carbon isotopic gradient during MIS 6 was very similar to the last glacial period with high values above 2000 m and low values below (Fig. 7). This pattern reflects the presence of GNAIW at intermediate depths and southern-sourced water below ~2000 m (Oppo and Lehman, 1993). Carbon isotopic values decreased throughout the water column from MIS 6 to Termination II with the most pronounced change occurring at intermediate depth. The bathymetic $\delta^{13}C$ gradient displays an S-shaped pattern with a distinct minimum at intermediate depth (~1500–1700 m), higher values between ~2000 and 2700 m, and another low value below 3000 m. The pattern of the bathymetric $\delta^{13}C$ gradient remains unchanged into the earliest part of MIS 5e (Fig. 7). During later MIS 5e, the benthic $\delta^{13}C$ gradient assumes a Holocene-like pattern with high $\delta^{13}C$ throughout the upper 2500 m of the water column.

Many other studies have reported a nutrient-rich water mass in the North Atlantic during Termination II (Zahn et al., 1987; Oppo and Lehman, 1993; Adkins et al., 1997; Oppo et al., 1997; Lototskaya and Ganssen, 1999; Oppo et al., 2001; Oppo et al, 2006; Skinner and Shackleton, 2006). In some cores, the low carbon isotopic values of Termination II extend into the early part of the LIG resulting in what

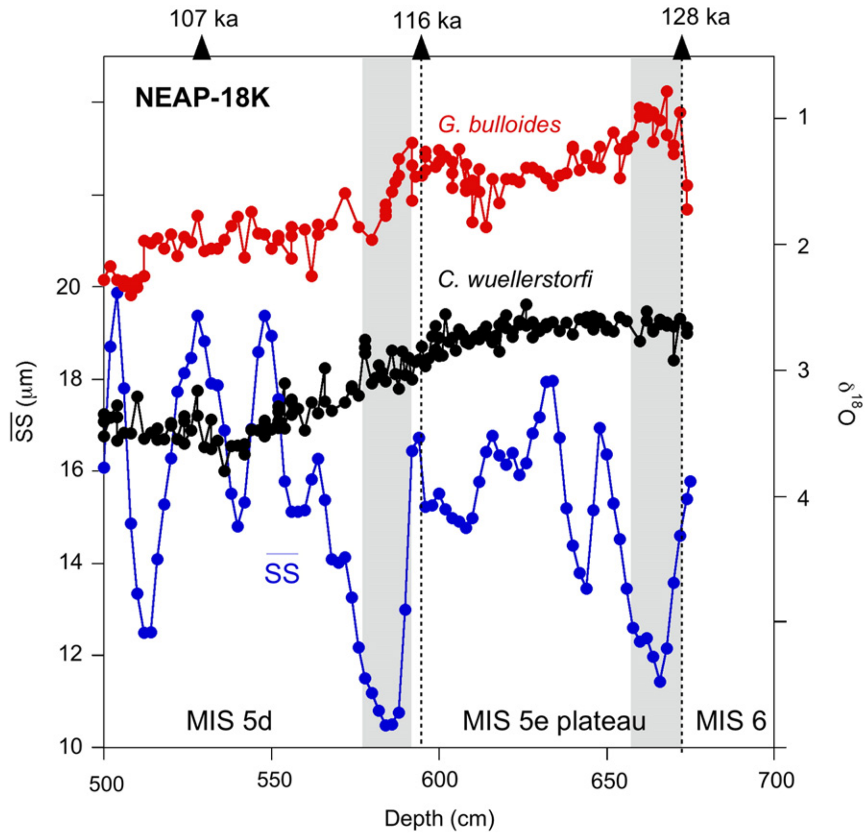

Fig. 5. Sortable silt mean size ($\overline{SS}$) (blue circles), benthic $\delta^{18}O$ of *C. wuellerstorfi* (black circles) and planktonic $\delta^{18}O$ of *G. bulloides* (red circles) for core NEAP-18K (Hall et al., 1998; Chapman and Shackleton, 1999). Vertical dashed lines designate an 80-cm section of the MIS 5e plateau (~128 to 116 ka). Shaded gray intervals denote minima in mean sortable silt denoting "slow events" in ISOW. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Skinner and Shackleton (2006) describe as a drawn out appearance to the deglaciation where benthic $\delta^{13}C$ lags behind the sharper increases in temperature and $\delta^{18}O$. At intermediate Site 982 on Rockall Plateau at 1145 m water depth, benthic carbon isotope minima are a common feature of Pleistocene deglaciations, including Termination II where the $\delta^{13}C$ depletion extends into MIS 5e (Venz et al., 1999).

At Site U1304, there was relatively little change in benthic $\delta^{13}C$ across Termination II from MIS 6 to early MIS 5e (Fig. 3). It is likely that Site U1304 continued to be bathed by southern sourced water through Termination II and early MIS 5e. The $\overline{SS}$ record at NEAP-18K indicates relatively weak flow below ~3000 m on the southern Gardar Drift during the early part of the MIS5e plateau (Figs. 5 and 6). Slightly higher $\delta^{13}C$ values occur between ~2000 and 2700 m in the depth range occupied today by ISOW on Gardar and Bjorn Drifts (Bianchi and McCave, 2000), suggesting that ISOW may have formed during Termination II and early MIS 5e but did not penetrate to depths of 3000 m.

GNAIW, which formed in the subpolar North Atlantic south of Iceland during MIS 6, ceased being produced during Termination II and the intermediate-depth North Atlantic was bathed by a low-$\delta^{13}C$ water mass. Previously suggested sources for this low-$\delta^{13}C$ water mass include nutrient-rich water sourced from the Southern Ocean or low-$\delta^{13}C$ water from the Nordic Seas (Oppo et al., 1997; Raymo et al., 2004).

Intermediate depths in the North Atlantic between ~800 and ~2000 m are influenced today by three principal water masses: LSW, MOW and Antarctic Intermediate Water (AAIW) (Kawase and Sarmiento, 1986). LSW is the dominant water mass between ~1000 and ~1600 m, but it was not formed during the LIG (Hillaire-Marcel et al., 2001). Similarly, MOW, which lies between ~800 and ~1000 m, was reduced during Termination II as marked by a prominent benthic $\delta^{13}C$ minimum in cores off the Strait of Gibraltar (Zahn et al., 1987). MOW was also likely reduced during early MIS 5e as this was the time of deposition of sapropel S5 indicating anoxic to euxinic deep water conditions in the Mediterranean (Rohling et al., 2006). If GNAIW, LSW, and MOW were greatly reduced or curtailed during Termination II and early 5e, then intermediate waters from the Southern Ocean could have penetrated farther into the North Atlantic. AAIW occupies levels shallower than 800 m today in the North Atlantic and is difficult to trace north of 25°. Other studies have suggested that AAIW was volumetrically more important during the past, penetrating to as high as 60° in the North Atlantic during the last termination (Rickaby and Elderfield, 2005).

If the low $\delta^{13}C$-water at intermediate depth in the North Atlantic has a Southern Ocean origin, then it may be related to carbon isotopic minima found elsewhere in the oceans. For example, Spero and Lea (2002) described the occurrence of a carbon isotope minimum event at Terminations I and II in deep-dwelling planktonic foraminifera in the equatorial Pacific. They suggested its origin was low-$\delta^{13}C$ water from the Southern Ocean that was transported to the base of the tropical thermocline by AAIW. In this case, the mid-depth low in $\delta^{13}C$ at Termination II in the North Atlantic may partly reflect the deglacial escape of old carbon and nutrients that were stored in bottom waters during the penultimate glaciation.

Alternatively, the low-$\delta^{13}C$ water in the intermediate-depth North Atlantic may have originated from the Nordic Seas (Bauch et al., 2000). Raymo et al. (2004) showed that the high-$\delta^{13}C$ water that forms in the Nordic Seas today was unusual in the past, and only occurred during extreme interglacials of the late Pleistocene. Our bathymetric $\delta^{13}C$ profile for Termination II and early MIS 5e is similar to their glacial-interglacial profiles between 0.6 and 1.8 Ma, which show a strong mid-depth $\delta^{13}C$ minimum. They concluded that low-$\delta^{13}C$, mid-depth water originated from the Nordic Seas, perhaps from water formed by brine

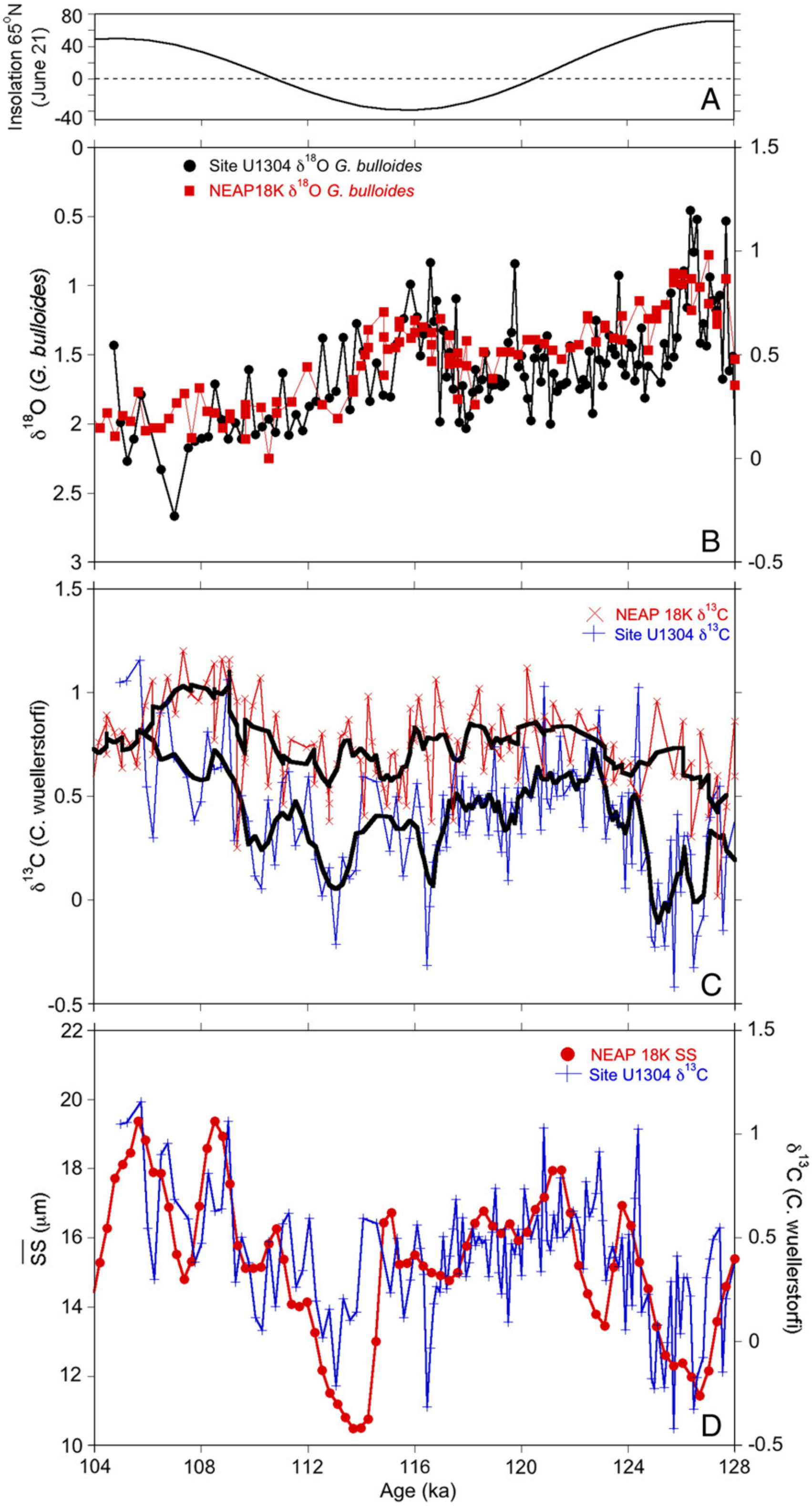

Fig. 6. A. Insolation anomaly relative to today at 65°N at boreal summer solstice. B. δ^{18}O of *G. bulloides* at Site U1304 (black) and NEAP-18K (red) (Chapman and Shackleton, 1999); C. Benthic δ^{13}C at Sites U1304 (blue line with bold 5-poing running mean) and NEAP-18K (red line with bold 5-point mean); D. Benthic δ^{13}C at Site U1304 (blue) and sortable silt (red) at NEAP-18K (Hall et al., 1998). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

rejection under sea ice (Dokken and Jansen, 1999; Bauch et al., 2000). Low benthic foraminiferal δ^{13}C has been noted in cores from the Norwegian Sea during the penultimate deglaciation and into the beginning of the LIG (Duplessy and Shackleton, 1985; Oppo and Lehman, 1995; Fronval and Jansen, 1997; Bauch et al., 2000).

5.3. Surface-deep connection

Planktonic δ^{18}O of foraminifera from Gardar Drift records temperature and salinity conditions of subpolar surface waters, which flow into the Nordic Seas via the North Atlantic Current. Minimum δ^{18}O of

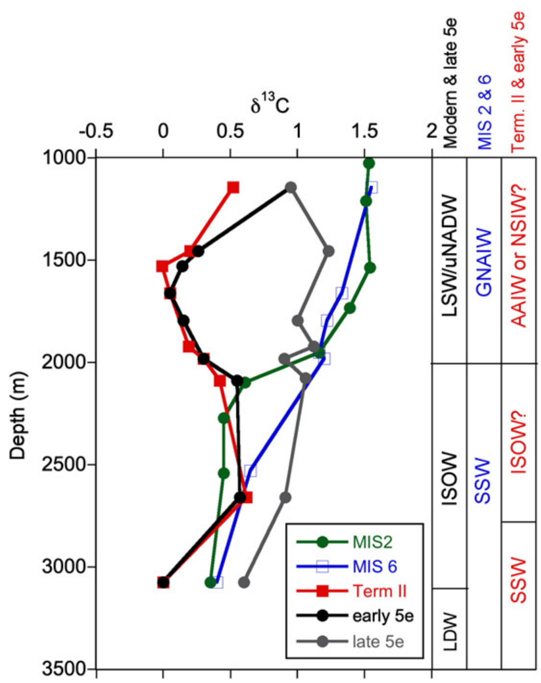

Fig. 7. Bathymetric δ^{13}C gradients. LSW = Labrador Sea Water; uNADW = upper North Atlantic Deep Water; ISOW = Iceland Scotland Overflow Water; LDW = Lower Deep Water; GNIW = Glacial North Atlantic Intermediate Water; SSW = Southern Source Water; AAIW = Antarctic Intermediate Water; NSIW = Norwegian Sea Intermediate Water. Data from Oppo et al. (1997) and this study.

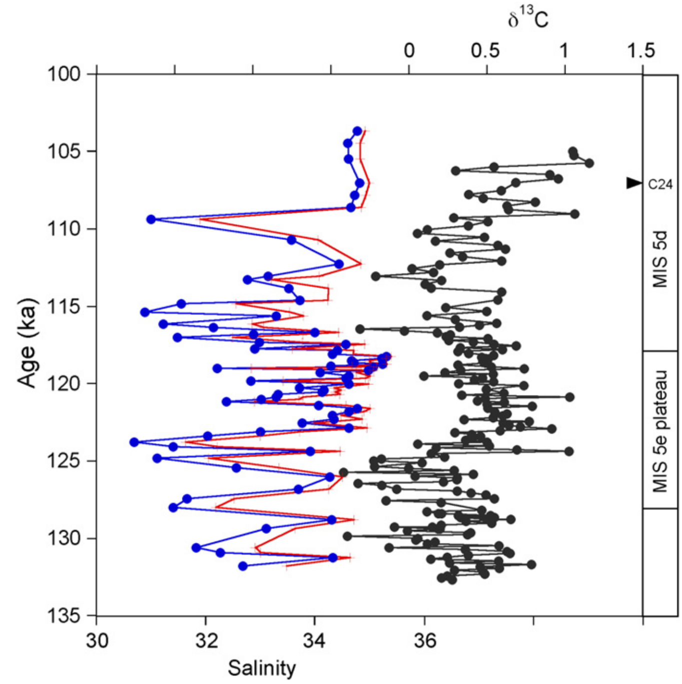

Fig. 8. Summer (blue) and winter (red) salinity estimates by dinocysts at core sample HU-90-013-013 off southern Greenland (de Vernal and Hillaire-Marcel, 2008) compared with benthic δ^{13}C at Site U1304 (black).

G. bulloides and *G. inflata* from ~128 to 126 ka indicates warming and/or reduced salinity of surface waters associated with peak boreal summer insolation (Fig. 6A,B). This occurs approximately at the same time as the low benthic δ^{13}C at Sites U1304 and 983 and low $\overline{SS}$ values at NEAP-18K. The peak summer insolation during early MIS 5e may have resulted in increased melting of continental ice, thereby increasing fresh water fluxes and decreasing surface water salinity in source areas for deep-water formation in the Nordic Seas. The Greenland ice sheet is particularly sensitive to warm, early summer conditions and likely had a negative mass budget during the earliest LIG (Otto-Bliesner et al., 2006).

We suggest that low-δ^{13}C water and slow current speed on Gardar Drift at 3–3.3 km during the early part of the LIG was the result of increased melt water fluxes to the Nordic Seas during peak boreal summer insolation, which decreased the strength and/or density of overflow to the North Atlantic. This left the slowly circulating SSW in place until a strengthened/denser ISOW was able to replace it later in 5e. Support for this interpretation comes from a comparison of the benthic δ^{13}C of Site U1304 and salinity estimated from dinocysts off southern Greenland (Fig. 8) (de Vernal and Hillaire-Marcel, 2008). The early part of the LIG was marked by generally lower and more variable surface salinity conditions, resulting from high rates of melt water flux from the retreating Greenland ice sheet. In the Norwegian Sea, the earliest part of the LIG was marked by significant freshwater and IRD inputs (Van Nieuwenhove et al., 2008; Bauch and Erlenkeuser, 2008). This period corresponds to the low benthic δ^{13}C at Site U1304 and slow current speed at NEAP-18K.

Modeling studies suggest that orbital parameters during the last interglacial favored a high-index state of the North Atlantic Oscillation (NAO) (Felis et al., 2004). Less vigorous ISOW due to larger admixture of fresher LSW has been associated with positive phases of the NAO during the late Holocene (Boessenkool et al., 2007). If peak boreal summer insolation during the early LIG favored a persistent positive NAO, then this state may have contributed to weak ISOW at 3000 m inferred from the deep Gardar Drift.

Higher and more stable salinity conditions prevailed during the latter part of MIS 5e beginning ~124 ka (Fig. 8) (Hillaire-Marcel et al., 2001; de Vernal and Hillaire-Marcel, 2008). At the same time, benthic δ^{13}C at Site U1304 and $\overline{SS}$ at NEAP-18K increased and remained high throughout the remainder of MIS 5e until 117 ka, indicating penetration of ISOW to depths of >3.3 km. Near the end of the MIS 5e plateau at 117 ka, salinity off south Greenland decreased and became more variable at the same time as benthic δ^{13}C and $\overline{SS}$ decreased with the lowest values occurring at 113 ka near the transition from MIS 5e to 5d. These results suggest a weakening of ISOW on the deep Gardar Drift in response to lowered surface salinity in the Labrador, Irminger and Nordic Seas.

Other studies have also reported a brief, but pronounced reduction in NADW at the end of MIS 5e and a greater proportion of SSW in the deep North Atlantic (Adkins et al., 1997; Lehman et al., 2002). In contrast, McManus et al. (2002) inferred enhanced ventilation at intermediate Site 980 and increased heat transport to the subpolar North Atlantic with glacial initiation (McManus et al., 2002). Our results are consistent with both of these observations because whereas benthic δ^{13}C increased at intermediate depth (Fig. 4), benthic δ^{13}C and $\overline{SS}$ decreased below 3000 m on Gardar Drift (Fig. 6D). These data can be explained by a shoaling of the boundary between intermediate and deep waters on the Gardar Drift such that the contribution and vigor of ISOW was reduced in deep-water sites but an ISOW influence persisted at intermediate depth.

Both chemical and physical ventilation tracers suggest a deepening and strengthening of ISOW on the deep Gardar Drift beginning at ~109 ka (Hall et al., 1998). At this time, salinity off south Greenland increased and remained high and steady (Fig. 8). Even though climate continued to deteriorate during MIS 5d, both chemical and physical ventilation tracers suggest a deepening and strengthening of ISOW on the deep Gardar Drift (Hall et al., 1998). For example, there is little δ^{13}C difference between Sites 983 and U1304 with both sites displaying high values of ~1‰, similar to the modern hydrography (Figs. 3, 4).

6. Summary and significance

We found evidence for low-$\delta^{13}C$ water throughout much of the water column during Termination II and the earliest part of the LIG on Gardar Drift. Minimum values of benthic $\delta^{13}C$ occurred at intermediate depths (~1500–2000 m) and below 3000 m. Sortable silt mean size at NEAP-18K (3275 m) was low during the early part of MIS 5e indicating relatively slow circulation from ~128 to 124.5 ka. The low-$\delta^{13}C$ water and slow current speed coincided with minimum planktonic $\delta^{18}O$ on Gardar Drift and low salinity surface conditions off southern Greenland (de Vernal and Hillaire-Marcel, 2008). We suggest that low- $\delta^{13}C$ water and slow current speed on deep Gardar Drift was related to increased meltwater fluxes to the Labrador and Nordic Seas during peak boreal summer insolation, which decreased the strength and/or density of overflow to the North Atlantic. The resumption of strong, well-ventilated ISOW on Gardar Drift was delayed until ~124 ka. The lagged response of ISOW on Termination II may have had an effect on AMOC, but because ISOW provides only one component of NADW, additional information is needed on the response of other contributors such as DSOW and LSW.

Most computer simulations of climate forced by enhanced greenhouse gas concentrations predict that AMOC will weaken, perhaps by as much as 30% by the end of this century, as the Nordic Seas become fresher and warmer (Manabe and Stouffer, 1995; Rahmstorf and Ganopolski, 1999; Rahmstorf, 2003; Delworth and Dixon, 2006; Dickson et al., 2002). Although future greenhouse gas forcing will be different than the insolation forcing of the LIG, our findings indicate that circulation on the Gardar Drift was weaker during the earliest part of the LIG when climate was warmer than present and the Greenland ice sheet was retreating (Otto-Bliesner et al., 2006).

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References

Adkins, J.F., Boyle, E.A., Keigwin, L.D., Cortijo, E., 1997. Variability of the North Atlantic thermohaline circulation during the last interglacial period. Nature 390, 795–798.

Bauch, H.A., Erlenkeuser, H., 2008. A "critical" climatic evaluation of last interglacial (MIS 5e) records from the Norwegian Sea. Polar Res. 27, 135–151.

Bauch, H.A., Erlenkeuser, H., Jung, S.J.A., Thiede, J., 2000. Surface and deep water changes in the subpolar North Atlantic during Termination II and the last interglaciation. Paleoceanography 15, 76–84.

Bianchi, G.G., McCave, I.N., 1999. Holocene periodicity in North Atlantic climate and deep-ocean flows south of Iceland. Nature 397, 515–517.

Bianchi, G.G., McCave, I.N., 2000. Hydrography and sedimentation under the deep western boundary current on Björn and Gardar Drifts, Iceland Basin. Mar. Geol. 165, 137–169.

Bodén, P., Backman, J., 1996. A laminated sediment sequence from the northern North Atlantic Ocean and its climatic record. Geology 24, 507–510.

Boessenkool, K.P., Hall, I.R., Elderfield, H., Yashayaev, I., 2007. North Atlantic climate and deep-ocean flow speed changes during the last 230 years. Geophys. Res. Lett. 34, L13614.

CAPE Project Members, 2006. Last Interglacial Arctic warmth confirms polar amplification of climate change. Quat. Sci. Rev. 25, 1383–1400.

Channell, J.E.T., Hodell, D.A., Lehman, B., 1997. Relative geomagnetic paleointensity and $\delta^{18}O$ at ODP Site 983 (Gardar Drift, North Atlantic) since 350 ka. Earth Planet. Sci. Lett. 153, 103–118.

Channell, J.E.T., Sato, T., Kanamatsu, T., Stein, R., Malone, M.J., and the Expedition 303/306 Scientists, 2006. Site U1304 summary. Proceedings of the Integrated Ocean Drilling Program, 303/306. doi:10.2204/iodp.proc.303306.104.2006.

Chapman, M.R., Shackleton, N.J., 1999. Global ice-volume fluctuations, North Atlantic ice-rafting events and deep-ocean circulation changes between 130 and 70 ka. Geology 27, 795–798.

Curry, W.B., Oppo, D.W., 2005. Glacial water mass geometry and the distribution of $\delta^{13}C$ of CO_2 in the Western Atlantic Ocean. Paleoceanography 20, PA1017. doi:10.1029/2004PA001021.

de Vernal, A., Hillaire-Marcel, C., 2008. Natural variability of Greenland climate, vegetation, and ice volume during the past million years. Science 320, 1622–1625.

Delworth, T.L., Dixon, K.W., 2006. Have anthropogenic aerosols delayed a greenhouse gas-induced weakening of the North Atlantic thermohaline circulation? Geophys. Res. Lett. 33, LO2606. doi:10.1029/2005Glo24980.

Dickson, R.R., Yashayaev, I., Meincke, J., Turrell, W., Dye, S., Holfort, J., 2002. Rapid freshening of the deep North Atlantic over the past four decades. Nature 416, 832–837.

Dokken, T., Jansen, E., 1999. Rapid changes in the mode of deep and intermediate water formation linked with atmospheric and ice sheet variations during the last glacial. Nature 401, 458–461.

Duplessy, J.C., Shackleton, N.J., 1985. Response of global deep-water circulation to the earth's climatic change 135,000–107,000 years ago. Nature 316, 500–507.

Expedition 303 Scientists, and the Expedition 303/306 Scientists, 2006. Site U1304. In: Channell, J.E.T., Kanamatsu, T., Sato, T., Stein, R., Alvarez Zarikian, C.A., Malone, M.J. (Eds.), Proc. IODP, 303/306, College Station TX (Integrated Ocean Drilling Program Management International, Inc.). doi:10.2204/iodp.proc.303306.104.2006.

Felis, T., Lohmann, G., Kuhnert, H., Lorenz, S.J., Scholz, D., Patzold, J., Al-Rousan, S.A., Al-Moghrabi, S.M., 2004. Increased seasonality in Middle East temperatures during the last interglacial period. Nature 429, 164–168.

Fronval, T., Jansen, E., 1997. Eemian and early Weichselian (140–60 ka) paleoceanography and paleoclimate in the Nordic seas with comparisons to Holocene conditions. Paleoceanography 12, 443–462.

Hall, I.R., McCave, I.N., Chapman, M.R., Shackleton, N.J., 1998. Coherent deep flow variation in the Iceland and American basins during the last interglacial. Earth Planet Sci. Lett. 164, 15–21.

Heinrich, H., 1988. Origin and consequences of cyclic ice rafting in the northeast Atlantic Ocean during the past 130,000 years. Quat. Res. 29, 142–152.

Hillaire-Marcel, C., de Vernal, A., Bilodeau, G., Weaver, A.J., 2001. Absence of deep-water formation in the Labrador Sea during the last interglacial period. Nature 410, 1073–1077.

Jansen, E., Overpeck, J., Briffa, K.R., Duplessy, J.-C., Joos, F., Masson-Delmotte, V., Olago, D., Otto-Bliesner, B., Peltier, W.R., Rahmstorf, S., Ramesh, R., Raynaud, D., Rind, D., Solomina, O., Villalba, R., Zhang, D., 2007. Palaeoclimate. In: Solomon, S., Qin, D., Manning, M., Chen, Z., Marquis, M., Averyt, K.B., Tignor, K.M., Miller, H.L. (Eds.), Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change. Cambridge University Press, Cambridge, pp. 433–497.

Kawase, M., Sarmiento, 1986. Circulation and nutrients in middepth Atlantic waters. J. Geophys. Res. 91, 9748–9770.

Lehman, S.J., Sachs, J.P., Crotwell, A.M., Keigwin, L.D., Boyle, E.A., 2002. Relation of subtropical Atlantic temperature, high-latitude ice rafting, deep water formation, and European climate 130,000–60,000 years ago. Quat. Sci. Rev. 21, 1917–1924.

Lototskaya, A., Ganssen, G.M., 1999. The structure of Termination II (penultimate deglaciation and Eemian) in the North Atlantic. Quat. Sci. Rev. 18, 1641–1654.

Mackensen, A., Hubberten, H.-W., Bickert, T., Fischer, G., Fütterer, D.K., 1993. The $\delta^{13}C$ in benthic foraminiferal tests of *Fontbotia wuellerstorfi* (Schwager) relative to the δ13C of dissolved inorganic carbon in southern ocean deep water: implications for glacial ocean circulation models. Paleoceanography 8 (5), 587–610.

Manabe, S., Stouffer, R.J., 1995. Simulation of abrupt climate change induced by freshwater input to the North Atlantic Ocean. Nature 378, 165–167.

McCartney, M.S., 1992. Recirculating components of the deep boundary current of the northern North Atlantic. Prog. Oceanogr. 29, 283–383.

McCave, I.N., Hall, I.R., 2006. Size sorting in marine muds: processes, pitfalls, and prospects for paleoflow-speed proxies. Geochem. Geophys. Geosyst. 7, Q10N05. doi:10.1029/2006GC001284.

McManus, J.F., Bond, G.C., Broecker, W.S., Johnsen, S., Labeyrie, L., Higgins, S., 1994. High-resolution climate records from the North Atlantic during the last interglacial. Nature 371, 326–329.

McManus, J.F., Oppo, D.W., Keigwin, L.D., Cullen, J.L., Bond, G.C., 2002. Thermohaline circulation and prolonged interglacial warmth in the North Atlantic. Quatern. Res. 58, 17–21.

North Greenland Ice Core Project Members, 2004. High-resolution record of Northern Hemisphere climate extending into the last interglacial period. Nature 431, 147–151.

Oppo, D.W., Lehman, S.J., 1993. Mid-depth circulation of the subpolar North Atlantic during the Last Glacial Maximum. Science 259, 1148–1152.

Oppo, D.W., Lehman, S.J., 1995. Suborbital timescale variability of North Atlantic deep water during the past 200,000 years. Paleoceanography 10, 901–910.

Oppo, D.W., Horowitz, M., Lehman, S.J., 1997. Marine core evidence for reduced deep water production during Termination II followed by a relatively stable Substage 5e (Eemian). Paleoceanography 12 (1), 51–63.

Oppo, D.W., Keigwin, L.D., McManus, J.F., 2001. Persistent suborbital climate variability in marine isotope stage 5 and Termination II. Paleoceanography 16 (3), 280–292 doi:10.1029/2000PA000527.

Oppo, D.W., McManus, J.F., Cullen, J.L., 2006. Evolution and demise of the Last Interglacial warmth in the subpolar North Atlantic. Quat. Sci. Rev. 25, 3268–3277. doi:10.1016/j.quascirev.2006.07.006.

Otto-Bliesner, B.L., Marshall, S.J., Overpeck, J.T., Miller, G.H., Hu, A., CAPE Last Interglacial Project members, 2006. Simulating Arctic climate warmth and icefield retreat in the last interglaciation. Science 311, 1751–1753.

Overpeck, J.T., Otto-Bliesner, B.L., Miller, G.H., Muhs, D.R., Alley, R.B., Kiehl, J.T., 2006. Paleoclimatic evidence for future ice-sheet instability and rapid sea-level rise. Science 311, 1747–1750. doi:10.1126/science.1115159.

Quadfasel, D., Kase, R., 2007. Present-day manifestation of the Nordic Seas overflows. In: Schmittner, A., Chiang, J.C.H., Hemming, S. (Eds.), Ocean Circulation: Mechanisms and Impacts, Geophysical Monograph Series, 173. AGU, Washington, DC, pp. 75–89.

Rahmstorf, S., 2003. Thermohaline circulation: the current climate. Nature 421, 699.

Rahmstorf, S., Ganopolski, A., 1999. Long term global warming scenarios computed with an efficient climate model. Climate Change 43, 353–367.
Raymo, M.E., Oppo, D.W., Flower, B.P., Hodell, D.A., McManus, J.F., Venz, K.A., Kleiven, K.F., McIntyre, K., 2004. Stability of North Atlantic water masses in face of pronounced climate variability during the Pleistocene. Paleoceanography 19, PA2008. doi:10.1029/2003PA000921.
Reid, J.L., 1994. On the total geostrophic circulation of the North Atlantic Ocean: flow patterns, tracers and transport. Prog. Oceanogr. 33, 1–92.
Rickaby, R.E.M., Elderfield, H., 2005. Evidence from the high-latitude North Atlantic for variations in Antarctic Intermediate water flow during the last deglaciation. Geochem. Geophys. Geosyst. 6, Q05001. doi:10.1029/2004GC000858.
Rohling, E.J., Hopmans, E.C., Sinninghe Damsté, J.S., 2006. Water column dynamics during the last interglacial anoxic event in the Mediterranean (sapropel S5). Paleoceanography 21, PA2018. doi:10.1029/2005PA001237.
Rohling, E.J., Grant, K., Hemleben, Ch., Siddall, M., Hoogakker, B.A.A., Bolshaw, M., Kucera, M., 2008. High rates of sea-level rise during the last interglacial period. Nature Geosci. 1, 38–42.
Rousseau, D.-D., Kukla, G., McManus, J., 2006. What is what in the ice and ocean? Quatern. Sci. Rev. 25, 2025–2030.
Schmitz, W.J., McCartney, M.S., 1993. On the North Atlantic circulation. Rev. Geophys. 31, 29–49.
Shackleton, N.J., Chapman, M., Sánchez-Goñi, M.F., Pailler, D., Lancelot, Y., 2002. The classic marine isotope substage 5e. Quat. Res. 58, 14–16.
Shackleton, N.J., Sánchez-Goñi, M.F., Pailler, D., Lancelot, Y., 2003. Marine isotope substage 5e and the Eemian Interglacial. Global Planet. Change 36, 151–155.
Skinner, L.C., Shackleton, N.J., 2006. Deconstructing Terminations I and II: revisiting the glacioeustatic paradigm based on deepwater temperature estimates. Quatern. Sci. Rev. 25, 3312–3321.
Smethie, W.M., LeBel, D.A., Fine, R.A., Rhein, M., Kieke, D., 2007. Strength and variability of the deep limb of the North Atlantic Meridional Overturning circulation from chloroflurocarbon inventories. In: Schmittner, A., Chiang, J.C.H., Hemming, S. (Eds.), Ocean Circulation: Mechanisms and Impacts, Geophysical Monograph Series, 173. AGU, Washington, DC, pp. 119–130.
Spero, H.J., Lea, D.W., 2002. The cause of carbon isotope minimum events on glacial terminations. Science 296, 522–525.
Thomas, R.G., Guyodo, Y., Channell, J.E.T., 2003. U channel track for susceptibility measurements. Geochem. Geophys. Geosyst. 4, 1050. doi:10.1029/2002GC000454.
Van Nieuwenhove, N., Bauch, H.A., Matthiessen, J., 2008. Last interglacial surface water conditions in the eastern Nordic Seas inferred from dinocyst and foraminiferal assemblages. Mar. Micropaleontol. 66, 247–263. doi:10.1016/j.marmicro.2007.10.004.
Venz, K.A., Hodell, D.A., Stanton, C., Warnke, D.A., 1999. A 1.0 Ma record of Glacial North Atlantic intermediate water variability from ODP Site 982 in the northeast Atlantic. Paleoceanography 14, 42–52.
Zahn, R., Sarnthein, M., Erlenkeuser, H., 1987. Benthic isotope evidence for changes of the Mediterranean outflow during the late Quaternary. Paleoceanography 2, 543–559.