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Evolution and demise of the Last Interglacial warmth in the subpolar North Atlantic

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Abstract

Detailed faunal, isotopic, and lithic marine records provide new insight into the stability and climate progression of the last interglacial period, Marine Isotope Stage (MIS) 5, which peaked approximately 125,000 years ago. In the eastern subpolar North Atlantic, at the latitude of Ireland, interglacial warmth of the ice volume minimum of substage 5e (MIS 5e) lasted ~10,000 years (10 ka) and its demise occurred in two cooling steps. The first cooling step marked the end of the climatic optimum, which was 2–3 ka long. Minor ice rafting accompanied each cooling step; the second, larger, step encompassing cold events C26 and C25 was previously identified in the northwestern Atlantic. Approximately 4 °C of cooling occurred between peak interglacial warmth and C25, and the region experienced an additional temporary cooling of at least 1–2 °C during C24, a cooling event associated with widespread ice rafting in the North Atlantic. Beginning with C24, MIS 5 was characterized by oscillations of at least 1–2 °C superimposed on a generally cool baseline. The results of this study imply that the marine climatic optimum of the last interglacial was shorter than previously thought. The finding that the eastern subpolar North Atlantic cooled significantly before C24 reconciles terrestrial evidence for progressive climate deterioration at similar and lower latitudes with marine conditions. Our results also demonstrate a close association between modest ice rafting, cooling, and deep ocean circulation even during the peak of MIS 5e and in the earliest stages of ice growth.

1. Introduction

Marine Isotopic Stages (MIS) and substages are identified on the basis of foraminiferal $\delta^{18}O$ variations (Shackleton, 1967). High-resolution marine and terrestrial records from the subpolar North Atlantic and surrounding landmasses have yielded a coherent view of the last interglacial, MIS 5. The ice volume minimum of the last interglacial was restricted to the earliest substage, MIS 5e (Shackleton, 1969). It is well established that climate instability followed the peak warmth of the last interglacial. Initially, six abrupt surface ocean cooling ("C") events (C19–C24) were documented during the younger isotopic substages of MIS 5 (MIS 5a–5d); five of these events were associated evidence of ice rafting throughout the North Atlantic, and each had a counterpart on land (McManus et al., 1994; Kukla et al., 1997). Subsequent work suggested earlier cooling or ice-rafting events within MIS 5 (e.g. Chapman and Shackleton, 1999; Cortijo et al., 1999; McManus et al., 2002; Oppo et al., 1997; Oppo et al., 2001), possibly confined to the far northern and western North Atlantic (e.g., Chapman and Shackleton, 1999; McManus et al., 2002). The Eemian represents the peak European terrestrial warmth associated with a portion of MIS 5; mounting evidence suggests that it was characterized by modest climate fluctuations (e.g. Frogley et al., 1999) and that its end was diachronous, with cooling beginning earlier at higher latitudes (e.g., Kukla et al., 2002; Turner, 2002; Müller and Kukla, 2004; Sánchez-Goñi et al., 2005). Atlantic meridional overturning circulation (MOC), which results in the conversion of surface to deep water, is a significant source of heat to the North Atlantic region, and variations in this system may have brought an end to the warmth of MIS 5e (Adkins et al., 1997; Hall et al., 1998) and contributed to renewed

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warmth during the MIS 5d glacial inception (McManus et al., 2002). The Atlantic MOC also appears to have weakened during the coldest events of MIS 5 (Chapman and Shackleton, 1999; Oppo et al., 2001; Lehman et al., 2002). Here, we present well-resolved records from a high accumulation rate site in the eastern subpolar North Atlantic that provide new and detailed perspectives of the climate progression of the glacial inception from this region, and how it relates to climate variability documented elsewhere.

2. Regional setting

We examined sediments from peak glacial MIS 6 and all of MIS 5 from Ocean Drilling Program (ODP) site 980, located on the Feni Drift in the eastern subpolar North Atlantic (55°29′N, 14°42′W, 2179 m) (Fig. 1). Previous studies document the high sensitivity of eastern subpolar North Atlantic surface waters to regional climate change, with the warm North Atlantic Current passing over the site during interglacial and interstadial times contrasting with incursion of cold polar waters during glacial times and stadial times (e.g., Bond et al., 1997; McManus et al., 1994; Ruddiman and McIntyre, 1976). This region is also a sensitive place to monitor the input of detritus-laden icebergs to the Atlantic, as it lies just north of the main path of icebergs crossing the North Atlantic from eastern North America and also records the input of ice-rafted detritus (IRD) from Greenland and Iceland (e.g. Bond and Lotti, 1995; Bond et al., 1997; Ruddiman, 1977). Today, deep waters at the site derive mainly from North Atlantic Deep Water (NADW). However, lower oxygen concentrations than in newly ventilated waters reflect the presence of relatively poorly ventilated Eastern Basin Deep Water (EBDW) from the south, which consists of recirculated NADW (Curry et al., 1988) and perhaps a small component of southern ocean water (SOW) (Lonsdale and Hollister, 1979). During glacial and deglacial times, the subpolar North Atlantic at the depth of site 980 contained an admixture of waters derived from the north and south (e.g., Oppo and Lehman, 1993; McManus et al., 1999). During smaller episodic reductions in NADW, site 980 may also be sensitive to the incursion of poorly ventilated EBDW (Fig. 1).

3. Material and methods

We approximately tripled the resolution for MIS 5 of the previously published (McManus et al., 1999, 2002; Oppo et al., 2001) benthic and planktonic isotope stratigraphies from site 980 and increased the resolution of the IRD records by seven-fold. We also tripled the resolution of the record of the relative abundance of the polar planktonic foraminifera, sinistral-coiling *Neogloboquadrina pachyderma* (*N. pachyderma* (s.)), and, for each faunal sample, generated quantitative warm and cold season sea

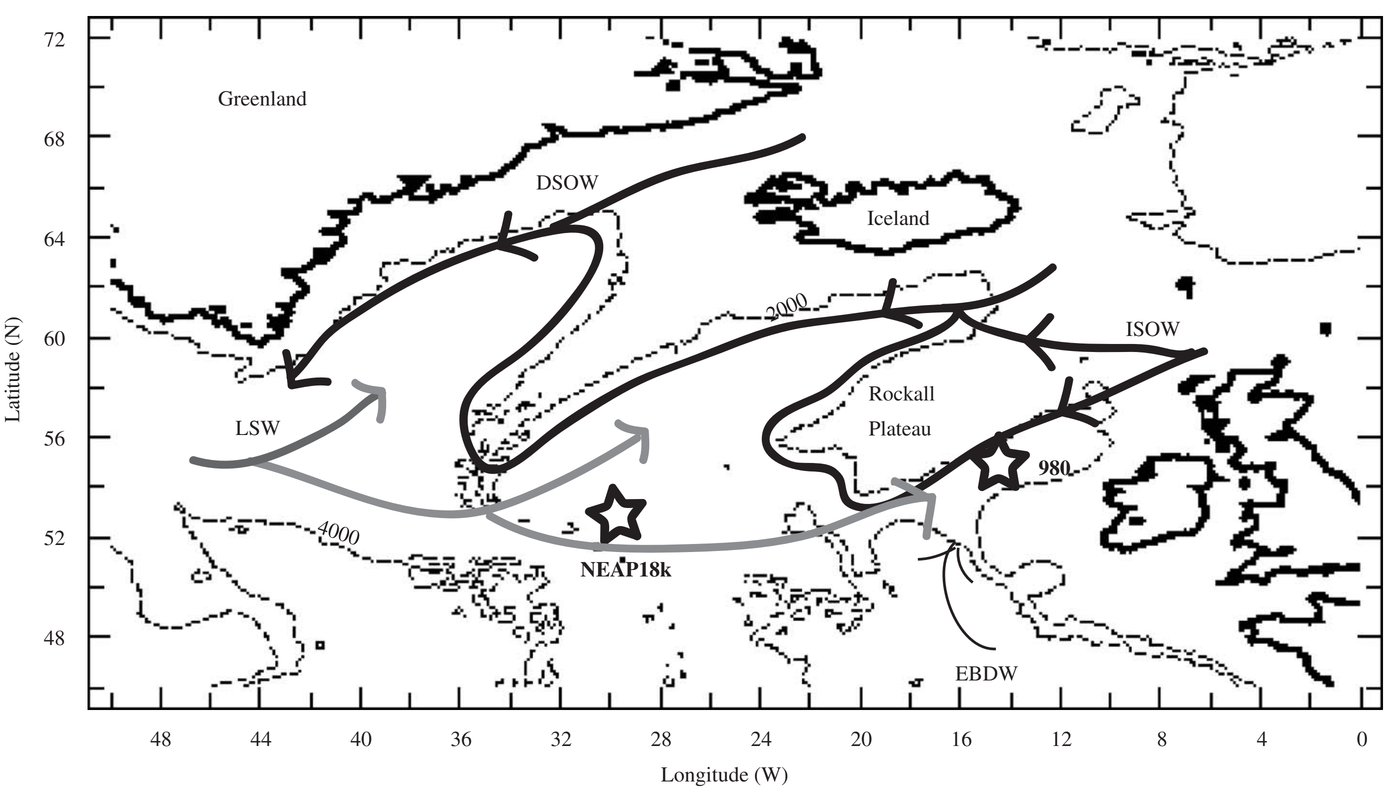

Fig. 1. Location of ODP Site 980 and NEAP18K. Deepwater at Site 980 is largely Labrador Sea Water (LSW) with a small component of Iceland–Scotland Overflow Water (ISOW). Eastern Basin Deep Water (EBDW) is also present in the southern Rockall Trough. The depth of NEAP18K is within ISOW and below LSW. The 2000- and 4000-m contours are shown as dotted lines.

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surface temperature (SST) estimates from planktonic foraminiferal assemblage changes using the Modern Analog Technique (MAT) (Prell, 1985) and the Brown University core-top database (Prell et al., 1999). Benthic $\delta^{13}C$ values, a proxy for deep ocean circulation, allow us to assess potential links between surface climate and the MOC.

Sediment samples were dry-sieved at 150 μm and then split to give approximately 300 planktonic foraminifera for census counts. Forty-one species of planktonic foraminifera were identified, if present. The actual number of planktonic foraminifera counted in a sample ranged from 259 to 611 specimens. For the MAT SST estimates, we used the average temperature of the best five analogs from the North Atlantic Ocean. Lithic fragments in the larger than 150-μm fraction were also counted and converted to number/gram data. Benthic isotope measurements were made on 1–3 specimens of *Cibicidoides wuellerstorfi* from the greater than 150-μm fraction. Planktonic isotope measurements were obtained on samples of *N. pachyderma* (dextral), each containing 7–9 specimens from the 212–250-μm fraction. All stable isotope measurements were made at the Woods Hole Oceanographic Institution on a Finnigan MAT252 and calibrated to Vienna Pee Dee Belemnite (VPDB) following standard procedures (e.g., Ostermann and Curry, 2000).

4. Results

All data are displayed in Fig. 2, and will be archived at the NOAA World Data Center for Paleoclimatology (http://www.ncdc.noaa.gov/paleo/) upon publication.

4.1. Chronology and accumulation rates

Glacial MIS 6 and interglacial MIS 5, including its substages, are readily identified on the basis of benthic $\delta^{18}O$ at site 980 (Figs. 2(a) and (b)), enabling the correlation of the record to a standard orbital chronology (e.g., Martinson et al., 1987) (inset 2a). The resulting age model suggests accumulation rates ranged from >20 cm/ka during the deglaciation and peak of the last interglacial, MIS 5e, to ~7 cm/ka during the younger substages, MIS 5d through MIS 5a. Our detailed records come from the ice volume maximum of MIS 6, identified by highest benthic $\delta^{18}O$ values, and from the whole of MIS 5. High accumulation rates during the MIS 6/5e deglacial transition, MIS 5e and the MIS 5e/5d ice growth transition, coupled with close (1–4 cm) sampling, afford the opportunity to examine, in detail, the onset and demise of peak warmth with multiple measures of surface conditions. The resolution of the new records ranges from <100 years to ~400 years.

4.2. IRD and N. pachyderma (s.) relative abundance

The IRD record displays the well-known sequence of events from this region (Fig. 2(j)). During late MIS 6, the abundance of IRD grains varied between a few grains and several hundred grains. The abundance of IRD grains increased by an order of magnitude at the end of MIS 6, marking an event of massive iceberg discharge, Heinrich event 11 (H11) (Bond et al., 1992; Broecker et al., 1992; Heinrich, 1988; McManus et al., 1994, 1998). Following H11, IRD decreased to pre-H11 levels before nearly disappearing during the peak of MIS 5e. The subsequent chronology of IRD events is consistent with previous records from the region (McManus et al., 1994). The abundance of IRD remained low until C24, which occurred late in glacial MIS 5d. IRD rose above the interglacial baseline four more times during MIS 5. Also consistent with previous studies, the interval from the beginning of MIS 5e until C24 was characterized by near-zero relative abundances of the polar planktonic foraminifera *N. pachyderma* (s.) (Fig. 2(i)), hinting at a lengthy warm and stable interval (McManus et al., 1994; 2002). The subsequent cold events C24–C19 were registered as increases in the relative abundance of *N. pachyderma* (s.).

4.3. SSTs

The summer SST estimates reveal gradual and abrupt changes, even before C24 (Fig. 2(i)), when an increase in the abundance of *N. pachyderma* (s.) suggests an abrupt cooling (Fig. 2(g). Low dissimilarity coefficients (⩽0.1) (Fig. 2(e)) and low standard deviation of the SST estimates (Fig. 2(f)) derived from the five best analogs (~1 °C) suggest reliable SST estimates for most of MIS 6 and substage 5e. High dissimilarity coefficients (>0.2) and large standard deviation of the SST estimates (>2 °C) during a small section of the MIS 5e/5d transition (near 15 mcd) indicate poor modern analogs and low confidence in the estimates for these samples. Modest dissimilarity coefficients (0.1–0.2) during the remainder of MIS 5 suggest that the magnitude of the SST estimates should be interpreted with caution, but as each SST event is accompanied by an increase in planktonic $\delta^{18}O$ values (Fig. 2(c)), there is little doubt of cooling.

4.4. Benthic $\delta^{13}C$ variations

High-amplitude benthic $\delta^{13}C$ variations (~0.5‰) occurred throughout MIS 6 and 5 (Fig. 2(d)), suggesting variable deep ocean circulation. Intervals of relatively high $\delta^{13}C$ values are believed to represent times when a higher proportion of deep waters newly formed in the North Atlantic bathed the site (e.g., Curry et al., 1988; Duplessy et al., 1988). Recent work using an independent radionuclide tracer of the MOC confirms a correspondence of suborbital $\delta^{13}C$ variations with variations in the strength of the MOC, at least during the last deglaciation (McManus et al., 2004). Nearly every benthic $\delta^{13}C$ variation recorded at site 980 within MIS 5 has an equivalent in a lower accumulation rate core from the Iceland Basin (Oppo and

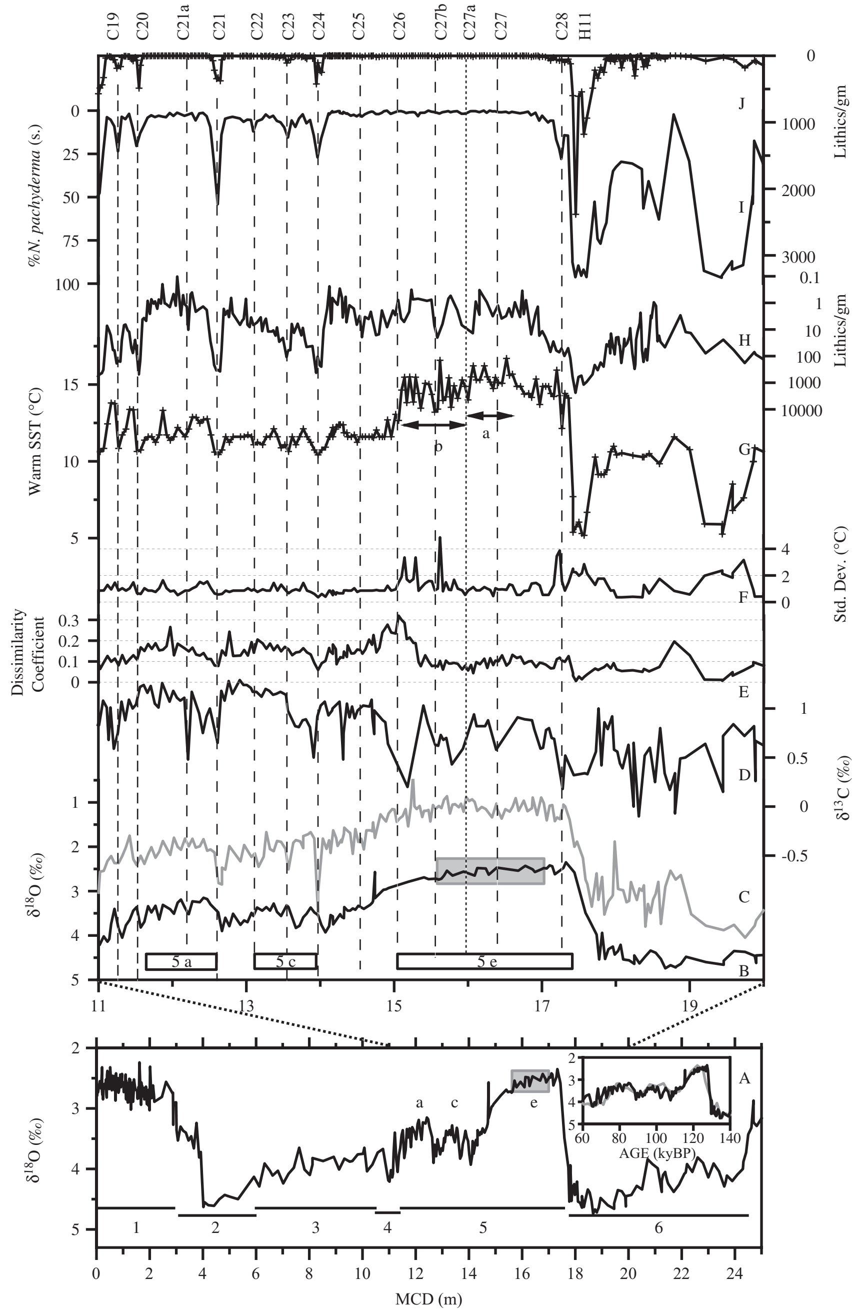

Fig. 2. Site 980 data plotted against composite depth, including new and published data (McManus et al., 1999, 2002; Oppo et al., 2001, 2003). (a) Benthic δ^{18}O data. Glacial (even) and interglacial (odd) marine isotope stages are labeled. Interstadial substages of MIS 5 are also labeled. The gray shading denotes the benthic δ^{18}O plateau (Shackleton et al., 2003). As described in Section 5.1, the presence of IRD just before the δ^{18}O plateau suggests that the equally low δ^{18}O values do not represent the sea level high stand. Inset shows the benthic δ^{18}O (black) data on an orbital time scale compared to the benthic orbital chronology of Martinson et al. (1987) (gray); (b) benthic δ^{18}O data for MIS 6 and 5 only; (c) planktonic δ^{18}O data; (d) benthic δ^{13}C data; (e) dissimilarity coefficients for SST estimates; (f) standard deviation of SST estimates from top five analogs; (g) SST estimates derived from modern analog technique; (h) lithic abundance data, plotted on a logarithmic scale; (i) relative abundance of *N. pachyderma* (s.); and (j) lithic abundance data, plotted on a linear scale.
Intervals marked "a" and "b" on Fig. 2(g) as discussed in Section 5.1. Vertical dashed lines denote previously identified IRD and/or cold events, as well as two additional events that are likely to be of a regional nature (C27b and C21a). Dotted lines denote additional events whose spatial extent is unclear. H11 refers to Heinrich event 11.

Lehman, 1995), lending confidence to the interpretation of benthic δ^{13}C variations as reflecting regional deep ocean circulation, rather than being driven by local processes (e.g., Mackensen et al., 1993).

5. Sea surface evolution

In addition to confirming previous investigations, the detailed records provide a new perspective on the onset, duration and demise of interglacial warmth and stability in this region (Fig. 2). For example, smaller IRD events, each resolved by multiple data points, are revealed late in the benthic δ^{18}O "plateau" (Shackleton et al., 2003) and on the MIS 5e/5d ice growth transition when the record is plotted on a logarithmic scale (Fig. 2(h)). Each of these IRD peaks is characterized by an order of magnitude less IRD than the later MIS 5 cold events (e.g., C19–C24) and two orders of magnitude less IRD than H11. Furthermore, the SST estimates (Fig. 2(g)) reveal more structure than variations in the relative abundance of *N. pachyderma* (s.) alone (Fig. 2(i)).

5.1. MIS 6 and the deglaciation

At site 980, the period of highest benthic δ^{18}O values, or the ice volume maximum of MIS 6, was marked by at least one SST oscillation having an amplitude of approximately 5 °C. Late glacial MIS 6 SSTs ranged between 5 and 10 °C colder than the peak warmth of MIS 5e, similar to that estimated from lower resolution records of planktonic foraminiferal assemblage changes (Ruddiman and McIntyre, 1984; Kandiano et al., 2004). As inferred previously from *N. pachyderma* (s.) relative abundance changes (Oppo et al., 2001), deglacial SST dropped markedly at site 980 in association with H11 and remained low until ice rafting declined to pre-H11 values. The SST record allows us to place an estimate of at least 5 °C on the magnitude of the deglacial (H11) cooling. The high standard deviation (~2 °C) of the H11 SST estimates results from the high proportion of *N. pachyderma* (s.) in these samples. Core-top assemblages suggest that the correspondence between *N. pachyderma* (s.) abundances and surface temperature weakens as percentages approach 100% and hence may correspond to a wide range of surface temperatures (Kohfeld et al., 1996; Prell et al., 1999).

The SST record further indicates that surface waters warmed abruptly by ~8 °C at the end of H11. A brief temperature oscillation suggested by SST estimates and by an increase in the relative abundance of *N. pachyderma* (s.) may be equivalent to a late deglacial oscillation, C28, inferred from planktonic δ^{18}O oscillations in two subtropical northwest Atlantic cores (Oppo et al., 2001). Surface waters warmed by an additional 2 °C before the maximum warmth of MIS 5e was achieved. This final-warming period was accompanied by considerable, but declining, input of IRD. Relatively, low planktonic δ^{18}O values preceding the interval of peak warmth most likely resulted from the continued delivery of low-δ^{18}O melt water to the site rather than similar warmth or sea level. Low benthic δ^{18}O values at this time were also likely due to the incorporation of melt water into the deep ocean (e.g., Lehman et al., 1993).

5.2. Substage 5e and glacial inception

The SST record indicates that the climatic optimum, denoted by an arrow labeled "a" (Fig. 2(g), was short lived. Faunal SSTs and a small increase in planktonic δ^{18}O values allow for a small cooling (C27) within the climate optimum of MIS 5e, consistent with previous suggestions based on marine (e.g., Cortijo et al., 1994; Seidenkrantz et al., 1995; Oppo et al., 2001; Knudsen et al., 2002; Fronval and Jansen, 1997) and terrestrial (e.g., Frogley et al., 1999) evidence. At site 980, this portion of the climatic optimum was characterized by trace amounts of IRD.

Faunal SST estimates suggest a ~1 °C cooling terminated the climatic optimum. For ease of discussion, we label the event terminating the climatic optimum "C27a", recognizing, however, that the spatial extent of this cooling is not yet clear. Although within the error of SST estimates, a student *t*-test indicates a high probability ($p > 0.0001$) that the mean SSTs during peak warmth (interval "a") were warmer (by ~1 °C) than during the following interval "b". A second *t*-test that did not include the samples with high dissimilarity coefficients (>0.15) confirmed that the means of the two intervals are statistically different ($p > 0.0005$). During C27a, IRD abundances reached 10 lithcs/gm for the first time in MIS 5e. In the middle of interval b, IRD abundances rose again to 10 lithics/gm ("C27b"), and SST estimates allow for the possibility of an associated brief cooling (Fig. 2(g)). Comparison of the placement of the labeled C27b event relative to the detailed benthic δ^{18}O chronology suggests that it was coeval with an unnamed event recorded as a significant positive excursion in the planktonic δ^{18}O record from sediment core NEAP18 K (Chapman and Shackleton, 1999), located in the northwestern central Atlantic (Fig. 1). Hence, event C27b likely represents a regional cooling episode.

The surface records suggests a second, larger, cooling step of ~3 °C, occurring at the beginning of a sequence of episodes of minor ice rafting, each one containing slightly higher IRD concentrations than the previous one. The precise placement of the cooling step is ambiguous because of the high dissimilarity coefficients immediately above and below the apparent cooling. However, dissimilarity coefficients are low before and after the apparent cooling step, lending confidence to the general interpretation of cooling. Furthermore, the cooling step occurred after the end of the MIS 5e benthic δ^{18}O plateau consistent with previous placement of cooling event C26 elsewhere (Chapman and Shackleton, 1999; Oppo et al., 2001; Lehman et al., 2002).

At site 980, IRD continued to increase after the C26 cooling step, whereas SST estimates suggest little additional cooling. A similar pattern is evident in the northwestern Atlantic in sediment core NEAP18 K (Chapman and Shackleton, 1999), where almost no cooling occurred between C26 and an IRD peak, assumed coeval with C25, an event originally identified in the Nordic Seas as an increase in IRD (e.g., Fronval and Jansen, 1997). Consistent with previous work, the C25 IRD event at 980 occurred near the middle of the MIS 5e/5d transition, during a time when there was a pause in the benthic δ^{18}O increase and a minor increase in the percentage of *N. pachyderma* (s.) (e.g., Chapman and Shackleton, 1999;

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Oppo et al., 2001; McManus et al., 2002). By contrast, C25 was associated with greater cooling in regions to the north and/or west of site 980 and NEAP18 K (e.g., Oppo et al., 1997; Cortijo et al., 1999; McManus et al., 2002), consistent with the incursion of polar waters from the northwest during these events (Bond et al., 2001).

5.3. Substages 5d–5a

Ice rafting decreased and surface temperatures may have warmed briefly following C25, but the region was generally cool for the remainder of MIS 5 at site 980. Beginning with C24, the rest of MIS 5 was characterized by frequent SST oscillations, as already inferred from abundance variations of *N. pachyderma* (s.) in a nearby core (McManus et al., 1994). At site 980, the faunal estimates suggest that the amplitude of the SST variations ranged from 1–2 °C. Both the planktonic δ^{18}O record and faunal SST estimates suggest that the greatest cooling occurred during events C24 and C21, with the planktonic δ^{18}O record suggesting a somewhat larger cooling (~4 °C) during C24. A previously unidentified minor IRD event (C21a) having ~10 lithics/gm occurred between C21 and C20. The SST record suggests that this event marked the end of the relatively warm portion of interstadial substage 5a. This event is likely coeval with a marked excursion in the planktonic δ^{18}O records from the subtropical western North Atlantic (Oppo et al., 2001; Huesser and Oppo, 2003), suggesting that the cooling was of regional extent.

MIS 5 cold events C24 and C21, occurring near the beginning of substages 5c and 5a, respectively, were associated with declining or low benthic δ^{18}O values, consistent with their possible association with decreasing ice volume (Chapman and Shackleton, 1999). However, the relatively large amplitude of other benthic δ^{18}O excursions (~0.5‰) is unlikely to have resulted from sea level decreases alone, as suborbital sea-level oscillations within MIS 5 were likely to be on the order 10 m or so (Thompson and Goldstein, 2005), significantly smaller than the 50 m or more implied by the benthic δ^{18}O. Instead, a portion of the δ^{18}O variations was caused by changes in the δ^{18}O of the deep water, possibly as a results of bottom water temperature changes (Shackleton, 1987; Cutler et al., 2003), water mass changes (e.g., Skinner and Shackleton, 2004), or of the contribution of a small amount of low-δ^{18}O melt water, either directly incorporated into sinking surface waters (Lehman et al., 1993) or transferred to deep waters during brine rejection (Vidal et al., 1998).

6. Deep ocean circulation

Each benthic δ^{13}C minimum during glacial MIS 6 was associated with an increase in ice rafting, consistent with previous work (e.g., Labeyrie et al., 1999). As already noted (Oppo et al., 2001), benthic δ^{13}C values at site 980 were low during and following H11, suggesting a reduced MOC for the duration of the deglaciation. The more pronounced cold events within MIS 5, those with significant ice rafting—C24, C23, C21, C20, and C19—were also associated with declining or low benthic δ^{13}C, indicating declining or reduced deep ocean ventilation. Event C21a may also have been associated with a reduced MOC. As during some MIS 3 (Oppo and Lehman, 1995) and last deglacial events (McManus et al., 2004), the reduced ventilation often outlasted the peak of the IRD event. For example, between C24 and C23, the sustained low δ^{13}C values suggest that the MOC did not fully recover between events.

Benthic δ^{13}C events of similar magnitude also occurred earlier, within MIS 5e, consistent with sedimentological evidence of persistent MOC variability (Bianchi et al., 2001). Low benthic δ^{13}C values were associated with labeled C27 and C27a events. Benthic δ^{13}C values remained low following C27a until the end of C27b. Benthic δ^{13}C values decreased again during C26, but recovered before the peak of the IRD event (C25). Thus, nearly every IRD peak within MIS 5, whether associated with less than 10 grains/gm or more than 100 grains/gm was accompanied by reduced benthic δ^{13}C values, implying a weakened MOC.

7. Climatic implications

The site 980 records confirm the variability of North Atlantic surface climate and deep ocean circulation throughout interglacial MIS 5. IRD increased before C25 in the Greenland and Iceland Seas (Fronval and Jansen, 1997). The first increase in ice rafting in the Greenland Sea within MIS 5e may be equivalent to C27. Two additional episodic IRD increases occurred in the Nordic Seas before C25, possibly corresponding to events C27a and C27b, both characterized by minor IRD at site 980. These early IRD increases are believed to indicate times when tidewater glaciers first reached the margins of the Nordic Seas (Fronval and Jansen, 1997).

Regardless of the precise correlation between events recorded at site 980 and in the Nordic Seas, our results suggest that early MIS 5e instabilities recognized at higher latitudes (e.g., Fronval and Jansen, 1997; Knudsen et al., 2002; Seidenkrantz et al., 1995) were also registered in the eastern subpolar North Atlantic as SST decreases and episodic increases in ice rafting. This result is consistent with the finding in a nearby site (V29-191; 54° 16′N, 16° 47′W) of millennial cycles within MIS 5e of the percentage of lithic grains derived from northerly sources, and of possibly correlative millennial cycles in detrital carbonate in a core from the northwest Atlantic (Bond et al., 2001). Increases in the abundance of these tracers during MIS 5e and the Holocene were interpreted to indicate times when cold waters from north of Iceland and from Labrador penetrated farther south and eastward, respectively, possibly due to solar-forced alterations of the atmospheric circulation. That these events were associated with the increased influence of polar waters is consistent with the

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observation that the C25 cooling was more significant in the northwestern Atlantic than in the eastern North Atlantic, as discussed in Section 5.2.

The site 980 SST and IRD records bear a close resemblance to the record of air temperatures over Greenland, as recorded by δ^{18}O in the North Greenland Ice Core Project (NGRIP) ice core (North Greenland Ice Core Project Members, 2004) (Fig. 3). Most noteworthy is the early cooling leading into C25 that is seen in both records. The early cooling leading into C25 has been previously designated C26 (Chapman and Shackleton, 1999) and a comparison to the site 980 SST record suggests that the NGRIP record captured most, if not all, of that cooling (Fig. 3).

According to our age model, the warmest SSTs (interval a) lasted ~2.5 ka, approximately comparable to the duration of the climatic optimum in western Europe (Sánchez-Goñi et al., 2005; Tzedakis, 2006). Interglacial warmth of MIS 5e lasted ~9–10 ka at site 980, and ended with the C26 cooling step. The stepwise cooling may indicate a threshold response to gradual changes in insolation forcing (e.g. Tzedakis, 2006). Oceanic cooling in the northeastern subpolar Atlantic may have contributed to the progressive climate deterioration documented in western European pollen sequences from similar latitudes (e.g. Müller and Kukla, 2004). Cooling contemporaneous with C26 was inferred from increases in planktonic δ^{18}O in the central subpolar North Atlantic (Chapman and Shackleton, 1999), west of Iberia (Sánchez-Goñi et al., 2005) and from planktonic δ^{18}O (Oppo et al., 2001) and alkenone (Lehman et al., 2002) records from the western subtropical North Atlantic, suggesting its widespread occurrence, and parallel cooling in southern Europe was inferred from terrestrial pollen sequences in marine cores (Sánchez-Goñi et al., 2005). Likewise, a cooling during C25 occurred in many regions of the North Atlantic although at site 980 and NEAP18K (Fig. 1), it was marked by increased IRD rather than significant cooling.

Variations in the North Atlantic MOC, which transports heat from low to high latitudes, and is a major source of heat to the North Atlantic region, have been implicated as a driver or amplifier of millennial scale climate change. It is, therefore, noteworthy that the amplitude of benthic δ^{13}C variability, a proxy for changes in North Atlantic deep overturning, was similar during MIS 6, the deglaciation, substage 5e, and the remainder of MIS 5. Arguably, the δ^{13}C excursions that occurred in association with H11 and the cooling before C25 cooling were larger than the δ^{13}C excursions associated with weaker cooling, allowing a weak correspondence with the amplitude of cooling.

Examination of a benthic δ^{13}C record from NEAP18K (Fig. 1), a deeper (3275 m) subpolar North Atlantic core

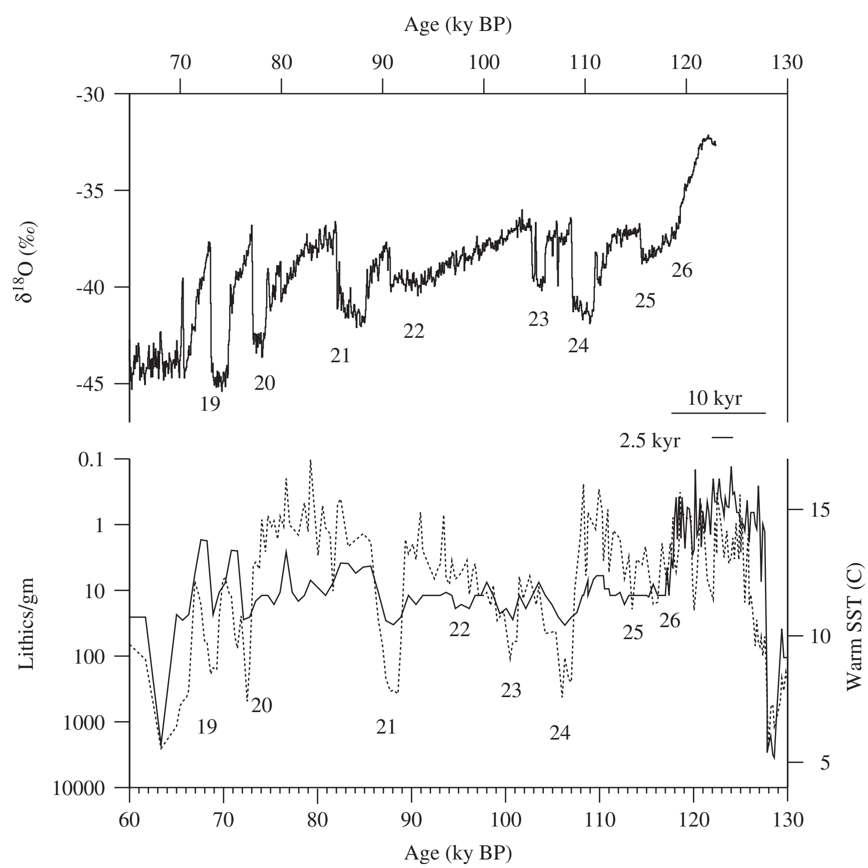

Fig. 3. Summer SST estimates (solid) and IRD data (dashed) showing the same sequence of events as NGRIP δ^{18}O (North Greenland Ice Core Project Members, 2004). Age model for site 980 as in Fig. 2(a).

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(e.g. Chapman and Shackleton, 1999), shows pronounced δ^{13}C minima during C21 and spanning the interval from C24 to C23, in agreement with the site 980 record. However, unlike site 980, C26 was the only other surface event having a clear association with low δ^{13}C values. By contrast, nearly every ice-rafting event at site 980 was associated with a decrease in benthic δ^{13}C values. The difference between the δ^{13}C records of the two sites may reflect the different sensitivities of the two sites to changes in deepwater ventilation. Even modest reductions in the MOC may be enough to permit poorly ventilated, low-δ^{13}C water EBDW to enter the Rockall Trough (Fig. 1), and this may be what happened during some of the weaker cool events. By contrast, NEAP18K is closer to the path of deep waters formed in the Nordic Seas, and low-δ^{13}C SOW only reaches the site during the more significant reductions of the MOC.

A likely explanation for the persistent connection between MOC reductions and increased IRD at site 980 is that decreases in the MOC contribute to cooling and increased ice rafting (McManus et al., 1999). The mechanism of minor ice-rafting events during the peak of MIS 5e and early on the benthic δ^{18}O transition is likely related to changes in surface circulation (Bond et al., 2001). Larger ice-rafting events beginning with C24 probably involved ice sheet instability (e.g. Broecker et al., 1992; Chapman and Shackleton, 1999). The reduced surface density resulting from the large input of ice and freshwater during large ice rafting events may have further reduced the MOC as registered in deeper North Atlantic cores, providing a positive feedback to amplify North Atlantic surface climate change during these large events (Bond and Lotti, 1995).

The observation that benthic δ^{13}C remained low between C24 and C23 at site 980 and in other cores (Chapman and Shackleton, 1999) adds to evidence from the deep ocean that the MOC occasionally remains weak as surface climate improves. Even during the last deglaciation, when there appeared to be a strong relationship between North Atlantic climate and variations in the MOC, weak ventilation seemed to outlast the IRD events and potentially the associated cooling (McManus et al., 2004). One possible explanation is that vigorous intermediate water formation, which transports considerable heat northward (Talley, 1999), caused the modest warming between C24 and C23. Such an increase in the rate of intermediate water formation may not have been registered as δ^{13}C increases at the depths of site 980 or NEAP18K. Likewise, an early recovery of intermediate water may account for apparent early warming relative to MOC recovery during some MIS 3 events (Oppo and Lehman, 1995).

8. Conclusions

Detailed records for late MIS 6 and the whole of MIS 5 from ODP site 980, a high accumulation-rate site on the Feni Drift, document the history of surface conditions in the eastern subpolar North Atlantic and their relationship to regional and deepwater changes. The records capture the well-known sequence of events from the region, including a cooling during Heinrich event 11, an abrupt warming into MIS 5, and cold events C24 through C19, occurring in MIS 5d–5a.

The new records reveal that interglacial warmth lasted ~10 ka at site 980, with peak warmth lasting as little as ~2–3 ka. The data hint at a small cooling even within the interval of climatic optimum, possibly associated with a small reduction in the MOC. A small (~1 °C) cooling step marked the end of the climatic optimum. Two small episodic increases in ice rafting occurred after the end climatic optimum, but surface waters remained relatively warm until a second more significant (~3 °C) cooling step that occurred after the end of the benthic δ^{18}O plateau, suggesting that ice growth had already begun. This cooling step is roughly comparable to C26, first identified as an IRD event in the Nordic Seas (Fronval and Jansen, 1997). The IRD peak of C25 soon followed. Unlike the northwestern subpolar and subtropical North Atlantic (e.g. Oppo et al., 2001), C25 was not marked by significant additional cooling at site 980 and in the central subpolar North Atlantic.

The finding of significant cooling (~4 °C) between the climatic optimum of MIS 5e and C25 is consistent with evidence from the NGRIP ice core (North Greenland Ice Core Project Members, 2004) and northwesterly regions of the subpolar North Atlantic, also showing significant cooling prior to C24. Based on a correlation to the site 980 SST record, it is possible that the large cooling recorded by the NGRIP record, leading into C25, is actually the C26 event recorded in many marine records. The finding that the eastern subpolar North Atlantic cooled significantly before C24 is also consistent with pollen evidence of progressive climate deterioration in western European following a brief (~2 ka) climatic optimum (Müller and Kukla, 2004; Sánchez-Goñi et al., 2005; Tzedakis, 2006).

The site 980 benthic δ^{13}C record exhibits variability during MIS 6, the deglaciation, and the whole of MIS 5, suggesting that variations in the strength of the MOC occurred throughout the study interval. All low δ^{13}C excursions were associated with cooling and an increase in ice-rafted grains, consistent with the hypothesis that reductions in northward heat transport by the MOC affected surface climate.

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