Thermohaline instability in the North Atlantic during meltwater events: Stable isotope and ice-rafter detritus records from core SO75-26KL, Portuguese margin

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Abstract. A benthic isotope record has been measured for core SO75-26KL from the upper Portuguese margin (1099 m water depth) to monitor the response of thermohaline overturn in the North Atlantic during Heinrich events. Evaluating benthic δ18O in TS diagrams in conjunction with equilibrium δ18O fractionation implies that advection of Mediterranean outflow water (MOW) to the upper Portuguese margin was significantly reduced during the last glacial (<15% compared to 30% today). The benthic isotope record along core SO75-26KL therefore primarily monitors variability of glacial North Atlantic conveyor circulation. The 14C accelerator mass spectrometry ages of 13.54±0.07 and 20.46±0.12 ka for two ice-rafter detritus (IRD) layers in the upper core section and an interpolated age of 36.1 ka for a third IRD layer deeper in the core are in the range of published 14C ages for Heinrich events H1, H2, and H4. Marked depletion of benthic δ13C by 0.7-1.1% during the Heinrich events suggests reduced thermohaline overturn in the North Atlantic during these events. Close similarity between meltwater patterns (inferred from planktonic δ18O) at Site 609 and ventilation patterns (inferred from benthic δ13C) in core SO75-26KL implies coupling between thermohaline overturn and surface forcing, as is also suggested by ocean circulation models. Benthic δ13C starts to decrease 1.5-2.5 kyr before Heinrich events H1 and H4, fully increased values are reached 1.5-3 kyr after the events, indicating a successive slowdown of thermohaline circulation well before the events and resumption of the conveyor’s full strength well after the events. Benthic δ13C changes in the course of the Heinrich events show subtle maxima and minima suggesting oscillatory behavior of thermohaline circulation, a distinct feature of thermohaline instability in numerical models. Inferred gradual spin-up of thermohaline circulation after H1 and H4 is in contrast to abrupt warming in the North Atlantic region that is indicated by sudden increases in Greenland ice core δ18O and in marine faunal records from the northern North Atlantic. From this we infer that thermohaline circulation can explain only in part the rapid climatic oscillations seen in glacial sections of the Greenland ice core record.

Introduction

Sediments of the northern North Atlantic contain distinctive layers of ice-rafter debris (IRD) which are believed to have resulted from ice sheet instabilities that occurred episodically during the last glacial and triggered sudden surges of icebergs to the North Atlantic [Ruddiman, 1977; Heinrich, 1988; Broecker et al., 1992; Alley and MacAyeal, 1994; Broecker, 1994]. Negative planktonic foraminiferal δ18O excursions in conjunction with increased abundances of polar planktonic foraminiferal species have been used to infer that these events were associated with cooling and freshening of the North Atlantic’s surface waters north of 40°N [Bond et al., 1992, 1993; Keigwin and Lehman, 1994; Bond and Lotti, 1995]. On the basis of a conceptual model that invokes free ice sheet oscillations as a function of atmospheric and geothermal parameters, MacAyeal [1993] predicts that the Heinrich-IRD events were accompanied by a freshwater flux to the North Atlantic of the order of 0.16 Sv over a period of 250-500 years. The negative planktonic foraminiferal δ18O response combined with decreased sea surface temperatures (SSTs) derived from transfer functions has been used to infer that the density of North Atlantic surface waters were lowered during these periods to an extent that vertical overturn spun down or was brought to a complete halt [Maslin, 1995]. This would be consistent with numerical simulations, which have shown that freshwater forcing of much shorter duration and smaller magnitude may cause convective instabilities that weaken or temporarily even terminate deep convection in the North Atlantic, [Paillard and Labeyrie, 1994; Rahmstorf, 1994, 1995; Weaver and Hughes, 1994; Manabe and Stouffer, 1995].

IRD layers have also been documented outside the North Atlantic region of maximum IRD deposition, on the Portuguese and Moroccan continental margins [Kudrass and Thiede, 1970; Kudrass, 1973]. The occurrence of IRD there indicates a flow of icebergs from the north along the Ibero-Moroccan margin during the last glacial and deglacial. Recently, Lebreiro et al. [1996] reported on the occurrence of discrete IRD layers on Tore Seamount off Portugal. Oxygen isotope stratigraphy and mineral composition of these layers revealed that they correspond to Heinrich events 1, 2, 3, and 6.

We use benthic isotope and IRD records from core SO75-26KL from the southern Portuguese margin (Figure 1) to monitor the North Atlantic’s ventilation response to the Heinrich events. Core SO75-26KL today is in the northward advection path of the MOC located at 11.5°N [Broecker, 1994].
The path of Mediterranean Outflow Water (MOW) that flows at water depths between 700 and 1400 m along the western Iberian margin [Zenk and Armi, 1990]. Benthic δ18O in conjunction with TS diagrams implies that the contribution of MOW to middepth water masses at the upper Portuguese margin was significantly reduced during the last glacial so that core SO75-26KL was primarily under the influence of North Atlantic waters. Because the Portuguese margin is far outside the North Atlantic's main IRD belt, IRD fluxes there were considerably lower than in the open North Atlantic, and hemipelagic sedimentation continued during three IRD events found in core SO75-26KL. As a result, abundances of planktonic and benthic foraminifers remained high enough to allow for continuous stable isotope records and detailed 14C-accelerator mass spectrometry (AMS) dating across the IRD layers. The 14C-AMS ages that have been determined for the two upper IRD layers and interpolated ages for the third IRD layer farther down in the core confirm that the IRD layers are coeval with Heinrich layers 1, 2, and 4. Sedimentation rates at the upper Portuguese margin are higher by a factor of 3 compared to the open North Atlantic and allow monitoring of ventilation changes during the Heinrich events at much higher resolution. This provides a unique opportunity to study the response of thermohaline overturn in the North Atlantic to the Heinrich events and associated meltwater events.

Materials and Methods

A benthic foraminiferal isotope record was measured for core SO75-26KL (37°49.3'N, 09°30.2'W, 1099 m water depth) from the upper Portuguese margin. The last glacial-interglacial transition and three IRD layers which occur along the core were sampled at 2 cm intervals to obtain a higher stratigraphic resolution and check for hydrographic and sedimentologic fine structure during these intervals. The rest of the core was sampled at 5-10 cm intervals. Stable isotope measurements were run on 1 to 21 specimens of Cibicidoides wuellerstorfi or C. pseudoungerianus. Within the IRD layers, isotope measurements were carried out on C. pseudoungerianus only, with a minimum of six specimens per isotope sample. The foraminiferal tests were picked from the size fractions >250 μm. In addition, a planktonic isotope record was measured using 25 specimens of Globigerina bulloides from the 315-400 μm size fraction. The planktonic isotope record was used to enhance stratigraphic control on the core. Prior to isotope analysis the foraminiferal shells were cracked open to release potential sediment fillings. They were then ultrasonically rinsed in methanol and transferred to a CARBO KIEL automated carbonate preparation device that is linked on-line to a FINNIGAN MAT 251 mass spectrometer. Long-term reproducibility was 0.08 %o for δ18O and 0.05 %o for δ13C as calculated from replicate analyses of an internal carbonate standard (Solnhofen limestone, 63-80 μm) that was routinely run at a ten-sample interval. The isotope data are referred to the Pee Dee belemnite (PDB scale).

The 14C ages were determined via accelerator mass spectrometry (AMS) using the 3MV Tandetron system at the Leibniz-Labor of Kiel University [Nadeau et al., 1997]. Dated were 15 monospecific samples of G. bulloides containing 590
Figure 2. (a) Age model developed for the last glacial/interglacial section of core SO75-26KL by correlating the planktonic δ¹⁸O record with the continuously ¹⁴C-accelerated mass spectrometry (AMS) dated planktonic δ¹⁸O record from nearby core SU81-18 [Bard et al., 1989], and further ¹⁴C-AMS data deeper in the core. (b) Benthic δ¹⁸O record. (c) The ¹⁴C-AMS ages obtained across two IRD horizons. Mean ages of 13.34±0.7 and 20.46±1.2 ka for these horizons and an interpolated age of 36.1 ka for the third IRD horizon deeper down the core correlate with the range of radiocarbon ages for Heinrich events H1, H2, and H4 measured at open North Atlantic core sites (see discussion in the text). (d) Age model for core SO75-26KL which implies that sedimentation rates varied between 15 and 55 cm ky⁻¹.

to 1090 tests in the size fraction >250 µm and two shell fragments. All samples were ultrasonically rinsed in methanol. The shell fragments were treated with 10% H₂O₂ and dilute HCl to remove organic material and carbonate dust.

Ice-rafted detritus (IRD) was first counted from the size fraction >250 µm at 10 cm intervals. Three discrete IRD maxima were found at 127-143, 264-278, and 556-584 cm core depth. Detailed IRD countings were then carried out at 2 cm intervals across the IRD maxima to check for IRD variability within the maxima (Figure 2). To facilitate microscopic work, these countings were done on the size fraction >355 µm. Bulk volume of the samples was 45 cm³ on average. IRD grains were counted from the total sample at the above given grain sizes, no split samples were used for the countings. Maximum countings reached 689 grains for IRD layer 3. The mineral composition of the IRD layers was determined by X ray powder diffraction (XRD) scans on IRD samples using a Siemens D 5000 automated diffractometer with incident and diffracted beam monochrometer (CuKa radiation at 25 mA and 40 kV; scanning angle was 20°-50°). The dₘₐᵦ peaks were identified using the reference lists of Brindley and Brown [1984] and Bayliss et al. [1986]. Individual mineral percentages were determined from analog records using Biscaye’s [1965] planimetry factors. Additional XRD scans were run on samples immediately below and above the IRD layers to obtain the mineralogy of the sediments which were deposited prior to and after the IRD events.

Stratigraphy

Age control on the last glacial-interglacial transition (100-130 cm in core SO75-26KL) was obtained by correlating the
planktonic δ¹⁸O record with the planktonic δ¹⁸O record from a nearby core SU81-18 that has been dated by ¹⁴C-AMS in great detail [Bard et al., 1989]. A series of 17 ¹⁴C-AMS ages was measured farther down the core (Table 1). One sample at 174-178 cm yields a ¹⁴C age (reservoir corrected; see Table 1) of 14.32 ka that fits well with the early phase of deglaciation documented by the initial decrease of benthic and planktonic δ¹⁸O at this core depth (Figures 2a and 2b). Two samples bracketing the subtle benthic δ¹⁸O minimum around 400 cm core depth yield ages of 22.95 and 24.87 ka, somewhat younger than the age of 25.4 ka that has been estimated by Martinson et al. [1987] for oxygen isotope event 3.1. A coral fragment from 385 cm depth with an age of 24.26 ka (Table 1) supports the association with 3.1. A much more pronounced minimum occurs in the planktonic δ¹⁸O record at this level, constraining late stage 3 isotope event 3.1 at 401.5 cm core depth.

An additional age control point was obtained by assigning oxygen isotope event 3.13 of Martinson et al. [1987] to the δ¹⁸O minimum at 651.5 cm in the benthic isotopic record (Table 2, Figure 2b). Beyond about 20 ka the Martinson et al. [1987] timescale is based on tuning to orbital parameters compared with U/Th dates. That is the age of 43.9 ka given by Martinson et al. [1987] for oxygen isotope event 3.13 must be considered a calendar or calibrated age. Laj et al. [1996] suggest a correction of 2000 years at 40 ka (¹⁴C), rapidly decreasing to zero around 47 ka (¹⁴C), to convert the ¹⁴C timescale to calendar years. Thus we use a ¹⁴C age of 42 ka for oxygen isotope event 3.13 to ensure compatibility with the conventional ¹⁴C timescale that we use for core SO75-26KL.

Detailed ¹⁴C-AMS Dating of IRD Layers 1 and 2 (Equivalent to Heinrich Events 1 and 2)

Six ¹⁴C ages each were measured across IRD layers 1 (127-143 cm) and 2 (264-278 cm) (Figure 3). Of the six ¹⁴C ages across IRD 1, the first two show an increase with depth, while the lower four form a plateau in the age-depth function and even decrease with increasing depth by 140 years. This decrease is, however, similar to the standard deviations and therefore statistically not significant. The closely similar ages at different core depths indicate increased sedimentation rates for this part of the core, which contains the heart of the IRD layer 1, between 130 and 142 cm core depth (Figure 3a). The three AMS dates for this interval give a mean age of 13.54±0.7 ka (average depth 136 cm). The apparent age discontinuity between the IRD layer and overlying sediments is expected as a result of bioturbation whereby, after the rapid deposition of the IRD layer, younger tests of G. bulloides are mixed down to and into the upper reaches of the IRD layer, but few older tests are mixed up. This leads to artificially young dates for the upper boundary of the IRD layer [Manighetti et al., 1995; Trauth, 1995]. The same reasoning predicts an artificially high age for the bottom of the IRD layer due to lack of young tests being mixed down once rapid deposition of the IRD layer started. This is not observed as sample KIA 0007 at 147 cm shows the same age as the IRD layer. For our age model we use the average value of 13.54±0.7 ka at 136 cm (samples KIA 0004-0006, Table 1) for IRD layer 1 together with the other three measured ages (Table 2).
Table 2. Chronostratigraphic Control Points for Core SO75-26KL

<table>
<thead>
<tr>
<th>Core Depth, cm</th>
<th>Age, ka&lt;sup&gt;a&lt;/sup&gt;</th>
<th>Age Marker</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.5</td>
<td>3.0&lt;sup&gt;b&lt;/sup&gt;</td>
<td>^14C-AMS</td>
<td>Bard et al. [1989]</td>
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<tr>
<td>81.5</td>
<td>8.8&lt;sup&gt;b&lt;/sup&gt;</td>
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<tr>
<td>121.5</td>
<td>12.7&lt;sup&gt;b&lt;/sup&gt;</td>
<td>^14C-AMS</td>
<td>Bard et al. [1989]</td>
</tr>
<tr>
<td>124.0</td>
<td>12.64</td>
<td>^14C-AMS</td>
<td>KIA 0002&lt;sup&gt;c&lt;/sup&gt;</td>
</tr>
<tr>
<td>130.0</td>
<td>12.91</td>
<td>^14C-AMS</td>
<td>KIA 0003&lt;sup&gt;c&lt;/sup&gt;</td>
</tr>
<tr>
<td>136.0</td>
<td>13.54</td>
<td>^14C-AMS</td>
<td>KIA 0004-0006&lt;sup&gt;d&lt;/sup&gt;</td>
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<tr>
<td>147.0</td>
<td>13.48</td>
<td>^14C-AMS</td>
<td>KIA 0007&lt;sup&gt;c&lt;/sup&gt;</td>
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<tr>
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</tr>
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<td>KIA 0009</td>
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<td>KIA 0017&lt;sup&gt;c&lt;/sup&gt;</td>
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<tr>
<td>651.5</td>
<td>42.00&lt;sup&gt;e&lt;/sup&gt;</td>
<td>3.13&lt;sup&gt;e&lt;/sup&gt;</td>
<td>Martinson et al. [1987]</td>
</tr>
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<sup>a</sup> All ages are corrected for reservoir age of 400 years.
<sup>b</sup> Radiocarbon date from nearby core SU81-18.
<sup>c</sup> KIA number is lab code listed in Table 1.
<sup>d</sup> Oxygen isotope event; age is in ^14C-kiloannums (see text).

For the six AMS ^14C dates across IRD 2 the age differences are likewise small, and because of the larger statistical uncertainties at this age, generally not statistically significant (Table 1, Figure 3b). The three dates from the heart of the IRD layer (265-279 cm, Figure 3b) give an age of 20.46±0.12 ka at 272 cm depth for this episode of rapid accumulation. Again, the two overlying dates are younger, though the age difference to sample KIA 0009 at 20.16±2. ka is statistically insignificant. The average of 20.46±0.12 ka (KIA 0011-0013, Table 1) for IRD layer 2 together with the other three dates is used in the age model (Table 2).

The age model was calculated using a polynomial fit through the ^14C-AMS data (Figure 4). According to this model, IRD layers 1 and 2 lasted from 13.1 to 13.4 ka and from 20.2 to 20.6 ka. Mean sedimentation rates are 22 cm ky<sup>-1</sup> at core SO75-26KL. Sedimentation rates increase to 55 cm ky<sup>-1</sup> and 32 cm ky<sup>-1</sup> during IRD layers 1 and 2, respectively, suggesting rapid sedimentation during H1 and H2. Sevenfold to tenfold increases in sediment flux during Heinrich events have been inferred using ^230Th<sub>ex</sub> data from Heinrich layers in the open North Atlantic [Francois and Bacon, 1994; Thomson et al., 1995]. Smaller increases at our core site support the...
contention that sediment flux from icebergs was reduced off Portugal as the site is outside the region of maximum iceberg flow.

Using the polynomially fitted age model, total duration of H1 and H2 is 300 and 400 years, respectively. These estimates depend on how good the age estimates for the top and bottom of the Heinrich layers are. Rapid sedimentation in conjunction with differential bioturbation likely make the top age too young and the bottom age too old. As we have discussed above, our detailed 14C-AMS dating shows some evidence for the first but not for the latter. Still, spacing of the 14C data across both IRD layers is not sufficient to estimate ages of IRD boundaries unambiguously. Using 14C-AMS data and age errors at face value, this would limit IRD 1 to 200-300 years, and IRD 2 to 300-600 years duration. These are still fairly wide ranges, but they fit to similar estimates derived from 230Th data [Francois and Bacon, 1994; Thomson et al., 1995].

Mineralogy of IRD Layers Off Portugal

A distinct property of the Heinrich deposits in the North Atlantic is the elevated concentration of detrital carbonate and the presence of dolomite grains in the IRD layers [Bond et al., 1992; Bond and Lotti, 1995]. This has been related to limestone bedrocks in the Hudson Strait area and in Arctic Canada which were eroded by ice, transported by floating icebergs into the North Atlantic, and finally delivered to the seafloor by melting of the icebergs. Lead, strontium, and neodymium isotope compositions of the IRD particles support this hypothesis in that these data point to the Canadian shield as the primary source for the lithic particles [Gwiazda et al., 1996; Revel et al., 1996]. Ice-rafted particles in IRD layer 1 off Portugal and Morocco consist of a great variety of rock types with 17% detrital carbonate and some striated specimens [Kudrass, 1973]. Our XRD scans show that primary IRD components are quartz, plagioclase, feldspar, and calcite (Figure 5). Since the XRD measurements were performed on bulk samples, most of the carbonate signal is likely of biogenic origin, that is, calcareous foraminiferal tests and nanoplank-
Hughes, 1994; Rahmstorf, 1994, 1995]. From these simulations it was inferred that only small freshwater perturbations on the large-scale freshwater budget but more on local freshwater fluxes that are targeted at the immediate region of convection [Rahmstorf, 1995].

During the last glacial, convection of deep waters shifted from the Norwegian-Greenland Seas into the northern North Atlantic. As a result of the southward shift of convection and the associated reduction in thermic disequilibrium between surface waters and overlying atmosphere, evaporation rates were likely reduced leading to lower salinities and densities of surface waters and thus the ensuing reduction of convection rates in the North Atlantic [Boyle and Keigwin, 1987]. As a result, the core layer of deep water flow shifted in depth from 3000 m today to around 2000 m during the last glacial [Duplessy et al., 1988; Sarnthein et al., 1994]. Only in the immediate vicinity of convection, that is, north of 45°N, did the influence of newly convected deep waters reach water depths similar to todays [Sarnthein et al., 1994].

Iceberg drift and associated meltwater flow during Heinrich events occurred close to sites of convection in the glacial North Atlantic. Paleoceanographic studies at northern North Atlantic sites have suggested that Heinrich events caused a significant drop of surface temperatures and salinities that also resulted in changes of surface circulation [Bond et al., 1993; Maslin, 1995; Maslin et al., 1995; Sarnthein et al., 1995]. Enhanced meltwater flux in the course of the Heinrich events in this area thus likely induced convective instabilities that would have slowed down or even stopped thermohaline overturn. It has not been possible to test this hypothesis in detail because of the lack of continuous benthic isotope records across the IRD layers at type-core sites in northern North Atlantic that have been used to define the Heinrich events. Various studies provide evidence for a close coupling between decreased deep water production and variable meltwater discharge during the last deglaciation [Keigwin et al., 1991; Lehman and Keigwin, 1992]. However, these records do not reach far enough back in time to monitor thermohaline response to earlier Heinrich events, and the high flux of IRD at coring sites from the northern North Atlantic wiped out the presence of benthic foraminifera because of dilution with IRD of the pelagic sediments in which benthic foraminiferal abundances are low.

Thus sediment cores from outside the immediate area of maximum IRD deposition will provide the only chance to document thermohaline patterns during these meltwater events, provided that IRD deposition at these sites was small enough to ensure benthic foraminiferal abundances high enough for

**Figure 5.** Clay mineral composition of IRD layers and sediments immediately above and below in core SO75-26KL. Dolomite is present in small quantities in IRD layers 1 and 3 (equivalent to Heinrich events H1 and H4), calcite is most likely of biogenic origin (foraminifera and nanoplankton).
continuous benthic isotope measurements across the IRD layers. These sediment cores need to be taken from advection pathways that are linked to thermohaline overturn in the North Atlantic. Core SO75-26KL fulfills these requirements because (1) sedimentation rates are high, (2) IRD deposition was sufficiently low to leave enough benthic foraminifera for isotope analysis, and (3) the area is downstream from convection areas in the glacial North Atlantic.

Core SO75-26KL as a Monitor of Thermohaline Patterns in the Northern North Atlantic

At a water depth of 1099 m, core SO75-26KL today lies in the advection path of Mediterranean Outflow Water (MOW) that enters the North Atlantic with temperature-salinity (TS) values of 13ºC/38.4 [Zenk, 1975] (Figure 6a). The density of this water is around 37.4 (σ2=density on 200 dbar surface), that is, considerably higher than the density of 36.7 of North Atlantic Deep Water (NADW). Rapid mixing with less saline North Atlantic Central Water (NACW, TS=13º/35.6 [Zenk, 1975]) and Labrador Sea Water (LSW, a component of upper NADW; TS=3º/34.85 [Talley and McCartney, 1982], both flowing at the depth level of MOW in the North Atlantic, reduces the density of MOW so that it flows northward along the upper Portuguese margin at water depths of 700-1400 m (upper and lower core layers of MOW are at 750 and 1250 m [Zenk and Armi, 1990]). TS values from hydrocasts near core SO75-26KL are 11ºC/36.3 [Zahn, 1993]. Using end-member TS values of 13º/38.4 for MOW entering the North Atlantic, 13º/35.6 for NACW, and 3º/34.9 for LSW, one arrives at a mixing ratio of 0.34:0.50:0.16 for MOW:NACW:LSW to generate in situ TS values of 11º/36.3 at the site of core SO75-26KL (Figure 6a). That is, the middepth water mass at the upper Portuguese margin today consists of approximately 30% MOW and 70% middepth waters from the open North Atlantic, in good agreement with estimates derived from the "cascade box model" of Zenk [1975].

Following the concept of using TS diagrams with equilibrium δ18O (δs) fractionation [Zahn and Mix, 1991] we determine the potential contribution of MOW to the site of core SO75-26KL during the last glacial maximum (LGM) (Figure 6). TS-δs diagrams provide a valuable tool for the interpretation of foraminiferal δ18O values in that they add the constraint of vertical density stratification that helps to narrow down possible TS scenarios as inferred from foraminiferal δ18O to physically plausible solutions [Zahn and Mix, 1991; Labeyrie et al., 1992; Sarnthein et al., 1995; Weinelt et al., 1996]. The modern δs fractionation lines shown in Figure 6a have been computed using the paleotemperature equation of Shackleton [1974] and a δsw:salinity slope of 0.5 (δsw is the oxygen isotope composition of seawater) which results from a North Atlantic freshwater end-member around -18‰ in δsw (SMOW). This slope represents the δsw:salinity relationship for modern North Atlantic deep and middepth waters. We have fixed NADW at a benthic δs of +3.6‰ PDB and TS values of 2ºC/34.9 that are commonly measured at coring sites in the North Atlantic [e.g., Labeyrie et al., 1992]. For the TS values of 11ºC/36.3 that have been measured in hydrocasts near the site of SO75-26KL [Zahn, 1993] we obtain a δs value of +2.1‰ PDB. This value is consistent with late Holocene δ18O of C. wuellerstorfi of +1.48 ‰ PDB in core SO75-26KL (Figures 2 and 4) once the values have been corrected to the Uvigerina scale by adding 0.64‰.

For a first evaluation of water mass distribution at the LGM, the δs fractionation lines have been computed using the same slope as for the modern δs lines (Figure 6b). Glacial maximum lower NADW (LNADWLGM) was fixed to TS values of 0ºC/35.8 and a δs value of +5.3‰ PDB (Figure 6b) [Labeyrie et al., 1992; Sarnthein et al., 1995; Jung, 1996]. From glacial-interglacial variations of planktonic foraminiferal δ18O and from planktonic foraminiferal census counts along sediment cores from the Mediterranean it was concluded that salinities of glacial maximum Mediterranean waters were higher by 1.2-2.7 and temperatures were lower by 4º-6ºC [Thiede, 1978; Thunell, 1979; Thunell and Williams, 1989]. Using mean values of 5ºC cooling and a salinity increase of 2 at the LGM, we have set MOWLGM as it enters the North Atlantic to TS values of 8º/40.3 (Figure 6b).

To determine the potential contribution of MOWLGM to ambient water masses at core SO75-26KL we need to determine TS values for middepth waters in the glacial maximum North Atlantic. One end-member water mass is Glacial North Atlantic Intermediate Water (GNAIW) [Duplessy et al., 1988] or Upper North Atlantic Deep Water (UNADWLG M) [Sarnthein et al., 1994; Jung, 1996]. The reconstructions by Duplessy et al. [1992], Labeyrie et al. [1992], and Sarnthein et al. [1994, 1995] infer that salinities of UNADWLG M were similar to those of LNADWLGM, 35.8. Using a benthic δ18O of +5.0 PDB that is documented in sediment cores from the northern North Atlantic at water depths of 1100-1500 m [Jung, 1996] and a salinity of 35.8 yields a paleotemperature of +1ºC for UNADWLG M (Figure 6b). The TS values for UNADWLG M thus are 1º/35.8.

Few paleodata are available to trace the glacial-interglacial evolution of NACW as the second end-member for mixing with MOWLGM. Slowey and Curry [1995] infer a cooling of 2ºC at middepths (1-2 km) around the Bahamas, pointing to similar cooling of surface waters in the North Atlantic subtropical gyre, the potential source region for NACW. Summer sea surface temperature (SST) in the glacial maximum North Atlantic ranges from 5º to +3ºC at the northern margin of the subtropical gyre [Climate Long Range Investigation, Mapping, Prediction (CLIMAP), 1981; Sarnthein et al., 1995; Weinelt et al., 1996]. Winter temperatures were lower, approaching 0ºC in the northern North Atlantic [Sarnthein et al., 1995]. Thus glacial NACW temperatures could have been anywhere between present-day values (13ºC) and close to upper glacial deep water temperatures. Temperatures warmer than todays appear unlikely given that upper North Atlantic water masses were colder during the last glacial [Slowey and Curry, 1995].

We discuss the following two scenarios (Figure 6b) using the temperature and salinity estimates given by Duplessy et al. [1991] and Sarnthein et al. [1995]. (1) a cold subpolar NACWLGM that formed in the area of the positive salinity anomaly at 51º-54ºN inferred by Duplessy et al. [1991], with TS values of 6º/36, and (2) a warm NACWLGM that formed along the northern margin of the subtropical gyre (around 40ºN), with TS values of 13º/37.5. In the first scenario, NACW would have been considerably colder than today (ΔT = -7ºC) and salinity increased by 0.4, whereas in the second
Figure 6. (a) TS diagram showing water mass end-members that contribute today to the site of core SO75-26KL. Density lines are for the 200 dbar surface (approximately 2000 m water depth). Isolines of δ18O equilibrium fractionation are computed using the paleotemperature equation of Shackleton [1974] and a δ18O-salinity slope of 0.5 for the modern North Atlantic; freshwater δ18O is -18‰ (SMOW). NACW (triangle) is North Atlantic Central Water, MOW (solid dot) is Mediterranean Outflow Water, NADW (rectangle) is North Atlantic Deep Water, and LSW (diamond) is Labrador Sea Water. The cross shows TS values at the site of core SO75-26KL. Dotted lines give mixing triangle between water mass end-members. Today, a mixing ratio of 34% MOW, 50% NACW, and 16% LSW contributes to the hydrography at the site of core SO75-26KL. See text for discussion. (b) Same as in Figure 6a, except for the last glacial maximum (LGM). TS values for mixing end-members have been changed to estimated LGM values (see text). UNADW (diamond) is Upper North Atlantic Deep Water. Crosses indicate TS values at a benthic δ18O value of +4.0 Pee Dee belemnite (PDB) for highest possible MOW contribution if MOW mixes with cold (cross 1) or warm (cross 2) NACW. Mixing ratios for crosses 1 and 2 are indicated. The δ18O fractionation is computed using the same parameterization as in Figure 6a and correcting for a mean glacial δ18O increase of 1.2‰. (c) Same as in Figure 6b, except that the δ18O-salinity slope is set to 0.8, and freshwater δ18O is -30‰ (SMOW). This configuration accommodates glacially lowered precipitation temperatures and the contribution of meltwaters. See text for discussion.
scenario, NACW temperature is the same as today but salinity is considerably increased by 1.9.

To estimate the maximum possible contribution of MOW to middepth waters at the site of core SO75-26KL, the following boundary conditions apply: equilibrium δ13C of +4‰ PDB has to be maintained to be consistent with the observed benthic δ18O value (corrected to the Uvigerina scale) in core SO75-26KL, and density of the ambient water mass at the core site must not exceed that of underlying UNADV_LGM. Maximum density for this middepth water mass is defined in the TS-δc diagram as the intercept between the +4‰ δc isoline and the 37.6 (σ2) isopycnal of UNADV_LGM at a TS value of 7.2°/37.1 (Figure 6b). This point also defines the maximum contribution of MOW as it is closest along the +4‰ δc isoline to the TS coordinate of glacial MOW (Figure 6b).

As is shown in Figure 6b, the cold and warm NACW scenarios both indicate a maximum possible contribution of 10% to the middepth water mass at the Portuguese margin. In the case of the cold NACW, maximum MOW contribution is entirely defined by mixing between MOW and NACW; TS values of the mixing product are 6°/36.4. In the case of a warm NACW, MOW mixes with a water mass that consists of roughly equal parts of UNADV and NACW. TS values are 7.2°/37.1, as defined by the intercept between the +4‰ δc isoline and the 37.6 (σ2) isopycnal of UNADV_LGM (Figure 6b).

Both scenarios imply that the contribution of MOW to middepth waters at the upper Portuguese margin was only 10% compared to 30% today. These numbers change slightly if we assume that MOW_LGM contributed less than today to the hydrography of the shallow North Atlantic, significant benthic δ13C Anomalies in Core SO75-26KL: Benthic δ13C Anomalies in Core SO75-26KL: Thermochemical Instability During Heinrich Events

during Heinrich Events

Benthic δ13C in core SO75-26KL is increased during the last glacial by 0.7‰ compared to Holocene values (Figure 4d). Increased benthic δ13C has been observed in glacial sediments at middepth sites from various ocean basins and has been used to infer an enhanced partitioning of carbon between the upper and deep ocean [Boyle, 1986; Zahn et al., 1987; Oppo and Fairbanks, 1987; Boyle, 1988; Duplessy et al., 1988; Oppo and Fairbanks, 1990; Mix et al., 1991; Zahn et al., 1991; de Menocal et al., 1992; Yu et al., 1996]. It seems today more plausible to assume that the subpolar North Atlantic remained an important source of middepth and deep ventilation, also during the last glacial. Therefore the influence of MOW on the upper water masses of the North Atlantic must have been reduced at the LGM, as is also inferred from the TS-δc evaluations discussed above. The reduction was probably less severe during early stage 2 and stage 3 when sea level was higher, but from the above considerations it must be concluded that the glacial sections of the benthic isotope record from core SO75-26KL primarily document the variability of North Atlantic middepth ventilation and thus monitor thermohaline overturn in the glacial North Atlantic.
was considerably reduced, so that the contribution of MOW to
the increased benthic δ13C levels was minor and the data
primarily document ventilation from a northern North
Atlantic source.

Decreases in benthic δ13C of 0.7-1.1‰ that parallel the oc-
currence of IRD layers in core SO75-26KL document strongly
reduced water mass ventilation at the upper Portuguese margin
during Heinrich events H1, H2, and H4 (Figures 4d and 7a). A
fourth δ13C anomaly occurs at an interpolated age of 23.4-
25.4 ka (Figure 4d) which is not accompanied by an IRD layer
in core SO75-26KL. This δ13C minimum could represent the
convective slowdown that was associated with Heinrich event
H3 (27 ka [Bond et al., 1993]). We have no independent age
control on this anomaly, so the difference in age to the H3
event would be due to insufficient stratigraphic control. H3
has been considered exceptional in that its IRD and trace ele-
ment composition points to a more northerly (Scandinavian?)
origin and because it contains little or no carbonate minerals
and is not well represented in the North Atlantic’s Heinrich
belt [Grousset et al., 1993]. An IRD layer coeval with H3 was
found in core D11975P on Tore Seamount (39°N, 12.5°W),
about 92.6 km to the northwest of our core site [Lebreiro et
al., 1996]. That is, the icebergs reached Tore Seamount farther
offshore Portugal, but they did not reach nearshore site SO75-
26KL either because iceberg drift during H3 was reduced com-
pared to the other Heinrich events or because a strong boundary
current along the Portuguese margin prevented the
icebergs from reaching the site of core SO75-26KL.

The pattern of reduced ventilation during the Heinrich
events that is implied by the negative benthic δ13C anomalies
in core SO75-26KL during IRD deposition confirms considera-
tions of planktonic δ18O and SST anomalies that have been
used to infer significant density decreases of surface waters in
the northern North Atlantic during Heinrich events which
should have resulted in reductions of water mass convection
[Maslín et al., 1995]. Similar evidence has been found for
meltwater pulses during the last deglaciation [e.g., Keigwin et
al., 1991], and the evidence found in core SO75-26KL sup-
ports the hypothesis of a close linkage between surface ocean
conditions in the North Atlantic and deep ventilation [Paillard
and Labeyrie, 1994; Weaver and Hughes, 1994; Rahmstorf,
1994, 1995]. To evaluate the relation between changing sur-
face forcing in the course of Heinrich events and reduced ther-
mohaline overturn in more detail, we compare the IRD and
benthic δ13C records from core SO75-26KL with the IRD and
planktonic δ18O records from Site 609 and the Greenland Ice
Core Project (GRIP) ice core δ18O record (Figure 7).

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Core Project (GRIP) ice core δ18O record (Figure 7).

The records from core SO75-26KL and Site 609 do not ex-
actly match each other because of differences in chronology
that are likely related to a stronger influence of bioturbation at
Site 609 where sedimentation rates are much lower and IRD
concentrations are higher than at core SO75-26KL, thus en-
hancing the effects of bioturbation, for examples, on the top
and bottom ages of the IRD layer at Site 609 [Manighetti et
al., 1995; Trauth, 1995]. Also, we have not converted the
14C timescales of the marine records into a calendar year timescale
because this conversion is still uncertain. Comparison of the
marine records with the GRIP ice core δ18O record which has a
calendar year chronology thus remains tentative.

Despite the uncertainties in chronology, we find remark-
able similarities between the marine records (Figure 7a).
Absolute benthic δ13C minima in core SO75-26KL during H4,
before and near the end of H2, and immediately after H1, as
well as the secondary δ13C minima during the late stage of H4
and during H1, are mirrored in planktonic δ18O minima at Site
609 that occur in the same stratigraphic positions relative to
the IRD layers. The structural correspondence thus suggests a
close correlation between maximum meltwater flux and de-
creased thermohaline convection during the Heinrich events.
A conspicuous feature in the benthic δ13C record is the gradual
decline of values that starts 1.5-2.5 kyr before the abrupt on-
set of IRD deposition during H1 and H4. After both events,
benthic δ13C gradually increases over 1.2-3 kyr. Assuming a
close correlation between the rate of thermohaline circulation
and benthic δ13C levels, the gradual changes in benthic δ13C
mirror gradual changes of thermohaline overturn in the North
Atlantic that started well before the Heinrich events and lasted
much longer than the events.

Successive reduction of oceanic heat flux to the North
Atlantic as the conveyor circulation gradually weakened would
have led North Atlantic climate to colder conditions. A trend
toward colder conditions prior to H4 is documented in increas-
ingly more negative δ18O maxima of interstadials 11, 10, and
9 in the Greenland ice core δ18O record (Figure 7). The benthic
δ13C decrease prior to H4 thus is evidence that this cooling
was associated with decreasing thermohaline overturn.

Warming in the North Atlantic region, on the other hand, oc-
curred abruptly at the end of H4, as is indicated by sudden in-
creases in ice core δ18O and in marine faunal records from the
northern North Atlantic [Dansgaard et al., 1993; Bond et al.,
1993] (Figure 7). This is in contrast to the gradual increase in
benthic δ13C that would suggest a gradual strengthening of
thermohaline circulation and associated heat flux to the North
Atlantic.

A similar pattern of benthic δ13C change is observed
during H1 (Figure 7), but this event occurred during a period of
large-scale disintegration of northern hemisphere ice sheets in
the course of the last glacial-interglacial transition. The early
decrease of benthic δ13C correlates with the initiation of ice
sheet collapse that culminated in a sequence of meltwater
pulses and led to a reduction if not complete halt of thermohala-
nine overturn in the North Atlantic [e.g., Keigwin et al., 1991;
Lehman and Keigwin, 1992; Sarnthein et al., 1992]. Absolute
minima in benthic δ13C are recorded during and immediately
after H1. The second minimum likely corresponds to meltwater
pulse 1a (MWP 1a) of Fairbanks [1989]. The gradual in-
crease in benthic δ13C after MWP 1a documents the transi-
tion to the Holocene mode of circulation as meltwater
flux decreased and surface circulation assumed its interglacial
state. Thus environmental boundary conditions were different
for H1, but the response of deep circulation was similar during
H1 and H4 in that the response was more gradual than would
have been predicted from the abrupt onset of IRD deposition
and the abrupt onset of warming seen in the Greenland ice core
record at the end of H1 and H4 (Figure 7).
Benthic $\delta^{13}C$ changes during H2 are more rapid (Figure 7). The $\delta^{13}C$ decline prior to and during H2 is less severe, that is, 0.8‰ compared to 1.0-1.1‰ during H1 and H4. It thus seems that the reduction in thermohaline overturn was less intense during H2, because glacial meltwater flux was either less or not continuously targeted at the site of convection because of variable surface circulation. Apparently, the North Atlantic's conveyor circulation was less inert during H2, allowing thermohaline overturn to respond rapidly to changes in surface forcing.
Whether the inferred changes in thermohaline overturn were caused by gradual increases and decreases of glacial meltwater flux as the Laurentide ice sheet grew, collapsed, and later stabilized, or by changes in surface ocean and/or atmospheric circulation remains speculative. The changes in benthic δ¹³C that were associated with the Heinrich events were not monotonous but show subtle maxima and minima (Figure 7). This suggests oscillatory behavior of thermohaline circulation which is also indicated by numerical models that link convective instabilities to changes in surface ocean forcing [Weaver and Hughes, 1994; Rahmstorf, 1994, 1995]. These models also predict convective discontinuities during which thermohaline overturn abruptly jumps to minimum rates as freshwater forcing exceeds critical threshold values. The abrupt depletion of benthic δ¹³C that is documented in core SO75-26KL at the culmination of IRD deposition during H1 and H4 and during MWP 1a may be evidence for such discontinuous behavior. The early onset of benthic δ¹³C decrease clearly suggests that thermohaline circulation started to spin down well before H4 and that it may have conditioned the North Atlantic region for further ice sheet growth by way of reduced oceanic heat transfer.

The abrupt warmings at the end of the Heinrich events that are documented in marine records from the northern North Atlantic [Bond et al., 1993] and in the Greenland ice core record [Dansgaard et al., 1993] could not have been caused by abrupt increases in oceanic heat transfer to the northern Atlantic. The more gradual increase in benthic δ¹³C suggests that thermohaline circulation wound up slowly and resumed its full strength as much as 3 kyrs after the events. Along with Bond and Lotti [1995] we must therefore conclude that the high-frequency oscillations of North Atlantic climate seen in the Greenland ice core record were at least in part caused by changes outside the ocean system, for example, in atmospheric circulation.

Conclusions

Benthic δ¹³C in core SO75-26KL from the upper Portuguese margin (1099 m water depth) is increased by 0.7‰ during the last glacial, documenting enhanced ventilation of middepth waters that is also seen at other shallow core sites from the North Atlantic. Evaluation of benthic δ¹⁸O in TS diagrams with computed δEC fractionation lines implies that MOW contributed no more than 15% to the ambient water mass at the site of core SO75-26KL, compared to 30% today. Benthic δ¹³C in core SO75-26KL therefore primarily monitors variability of North Atlantic middepth water masses and thus traces thermohaline overturn in the North Atlantic.

The core contains three distinct IRD layers 30-60 cm thick. Mean ¹⁴C-AMS ages for IRD 1 and 2 of 13.5±4.07 and 20.46±12 ka and an interpolated age of 36.1 ka for IRD 3 confirm that the IRD layers are coeval with Heinrich layers H1, H2, and H4. Benthic δ¹³C displays marked minima during these periods, indicating reductions in thermohaline circulation. Enhanced meltwater flux that was associated with the Heinrich events apparently caused convective instabilities in the northern North Atlantic. The apparent coupling between maximum meltwater flux during the Heinrich events as indicated by minimum planktonic δ¹⁸O at Site 609 and minima in thermohaline circulation as indicated by minimum benthic δ¹³C in core SO75-26KL suggests that a thermohaline link existed between the subpolar North Atlantic and middepth waters at the upper Portuguese margin.

Initial decreases in benthic δ¹³C started about 1.5-2.5 kyr before H1 and H4, suggesting that the North Atlantic's conveyor circulation started to slow down well before the Heinrich events. Coeval reduction in nearward oceanic heat transfer likely conditioned North Atlantic climate to enhanced ice sheet growth that ultimately triggered the H4 event. The early spin-down of thermohaline circulation before H1 correlates with the initiation of large-scale ice sheet collapse in the course of the last glacial-interglacial transition. The Greenland ice core record and marine records from the northern North Atlantic suggest abrupt warming after the Heinrich events. The gradual increases of benthic δ¹³C in core SO75-26KL imply that thermohaline circulation resumed its full strength between 1.2 and 3 kyr after the Heinrich events, disqualifying abrupt increases in oceanic heat transfer to the northern North Atlantic as a driving force for these climatic jumps. From this we conclude that changes in atmospheric circulation must have contributed to the climatic oscillations seen in the Greenland record.

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