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Thermohaline instability in the North Atlantic during meltwater events: Stable isotope and ice-rafted 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 $\delta^{18}O$ in TS diagrams in conjunction with equilibrium δ_c 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 ^{14}C-accelerator mass spectrometry ages of 13.54±.07 and 20.46±.12 ka for two ice-rafted 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 ^{14}C ages for Heinrich events H1, H2, and H4. Marked depletion of benthic $\delta^{13}C$ 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 $\delta^{18}O$) at Site 609 and ventilation patterns (inferred from benthic $\delta^{13}C$) in core SO75-26KL implies coupling between thermohaline overturn and surface forcing, as is also suggested by ocean circulation models. Benthic $\delta^{13}C$ 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 $\delta^{13}C$ 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. Inferrred 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 $\delta^{18}O$ 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-rafted 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 $\delta^{18}O$ 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 $\delta^{18}O$ 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

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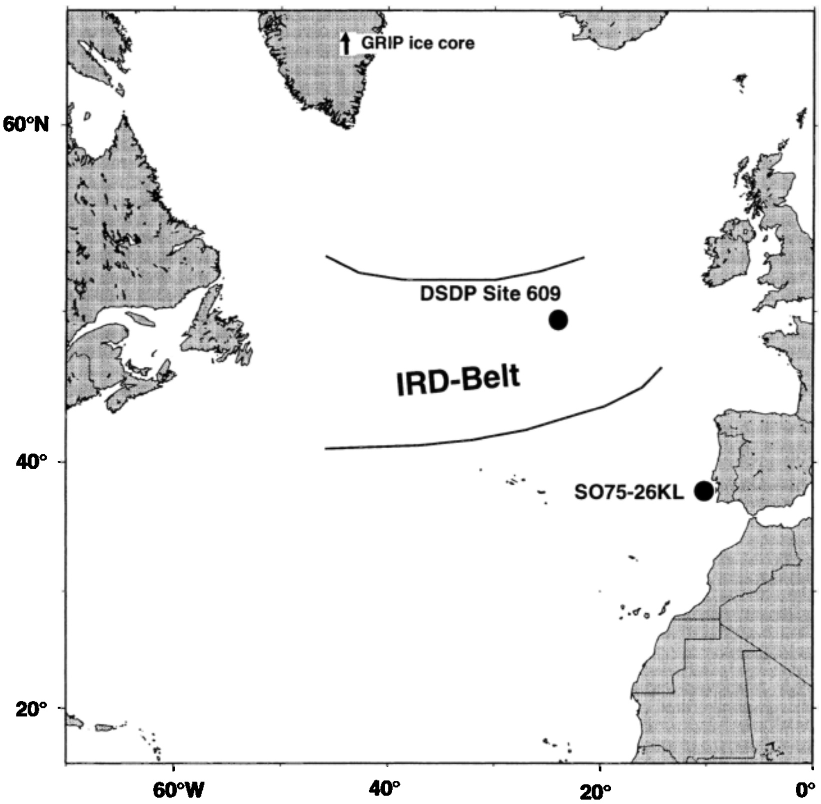

Figure 1. Position of R/V SONNE core SO75-26KL on the upper continental margin off Portugal. Ice-rafted detritus (IRD) and planktonic $\delta^{18}O$ records at DSDP Site 609 [*Bond and Lotti*, 1995] are used as reference for comparison with records from core SO75-26KL.

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 $\delta^{18}O$ 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 ^{14}C-accelerator mass spectrometry (AMS) dating across the IRD layers. The ^{14}C-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 ‰ for $\delta^{18}O$ and 0.05 ‰ for $\delta^{13}C$ 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 ^{14}C 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

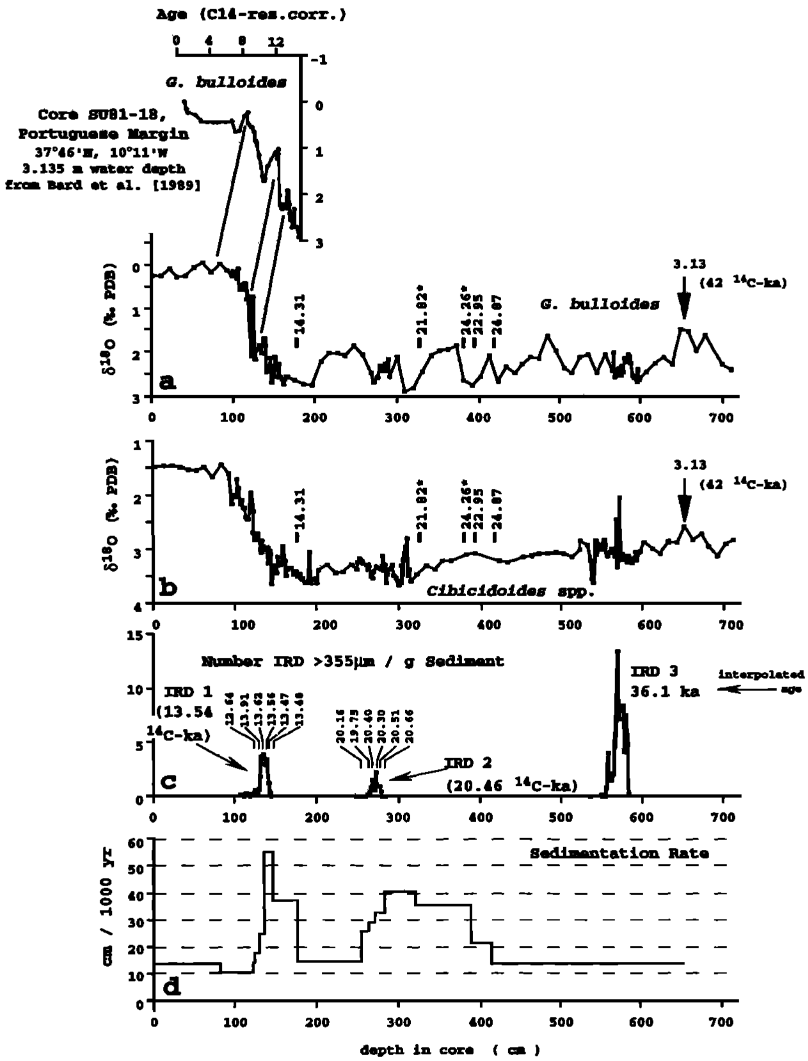

Figure 2. (a) Age model developed for the last glacial/interglacial section of core SO75-26KL by correlating the planktonic $\delta^{18}O$ record with the continuously ^{14}C-accelerated mass spectrometry (AMS) dated planktonic $\delta^{18}O$ record from nearby core SU81-18 [*Bard et al.*, 1989], and further ^{14}C-AMS data deeper in the core. (b) Benthic $\delta^{18}O$ record. (c) The ^{14}C-AMS ages obtained across two IRD horizons. Mean ages of 13.54±.07 and 20.46±.12 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 kyr^{-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% H_2O_2 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 cm^3 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 d_{hkl} 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 $\delta^{18}O$ record with the planktonic $\delta^{18}O$ record from a nearby core SU81-18 that has been dated by ^{14}C-AMS in great detail [*Bard et al.*, 1989]. A series of 17 ^{14}C-AMS ages was measured farther down the core (Table 1). One sample at 174-178 cm yields a ^{14}C 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 $\delta^{18}O$ at this core depth (Figures 2a and 2b). Two samples bracketing the subtle benthic $\delta^{18}O$ 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 $\delta^{18}O$ 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 $\delta^{18}O$ 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 (^{14}C), rapidly decreasing to zero around 47 ka (^{14}C), to convert the ^{14}C timescale to calendar years. Thus we use a ^{14}C age of 42 ka for oxygen isotope event 3.13 to ensure compatibility with the conventional ^{14}C timescale that we use for core SO75-26KL.

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

Six ^{14}C ages each were measured across IRD layers 1 (127-143 cm) and 2 (264-278 cm) (Figure 3). Of the six ^{14}C 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±.07 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±.07 ka at 136 cm (samples KIA 0004-0006, Table 1) for IRD layer 1 together with the other three measured ages (Table 2).

Table 1. ^{14}C-AMS Ages for Core SO75-26KL

Lab Code	Depth Interval, cm	Mean Depth, cm	Species	Individuals	^{14}C Age[a]	Error +	Error -	^{14}C Age[b]
KIA 0002	122-126	124	*G. bulloides*	1,090	13,040	140	130	12,640
KIA 0003	128-132	130	*G. bulloides*	840	13,310	100	100	12,910
KIA 0004	132-136	134	*G. bulloides*	680	14,020	110	110	13,620
KIA 0005	136-140	138	*G. bulloides*	800	13,960	160	160	13,560
KIA 0006	140-144	142	*G. bulloides*	680	13,870	90	90	13,470
KIA 0007	144-150	147	*G. bulloides*	640	13,880	230	230	13,480
KIA 0008	174-178	176	*G. bulloides*	690	14,710	130	130	14,310
KIA 0009	251-257	254	*G. bulloides*	590	20,560	220	210	20,160
KIA 0010	261-265	263	*G. bulloides*	1,080	20,150	220	220	19,750
KIA 0011	267-269	268	*G. bulloides*	890	20,800	240	230	20,400
KIA 0012	271-275	273	*G. bulloides*	900	20,700	380	360	20,300
KIA 0013	275-277	276	*G. bulloides*	930	21,450	390	370	21,050
KIA 0014	279-287	283	*G. bulloides*	760	21,060	230	230	20,660
KIA 0015	320-322	321	gastropod shell		22,220	370	360	21,820
KIA 0016	385	385	deep- water coral		24,660	340	330	24,260
KIA 0017	390-394	392	*G. bulloides*	725	23,350	520	490	22,950
KIA 0018	413.5-416.5	415	*G. bulloides*	930	25,270	660	610	24,870

[a] Uncorrected.
[b] Corrected for reservoir age of 400 years.

Table 2. Chronostratigraphic Control Points for Core SO75-26KL

Core Depth, cm	Age, ka[a]	Age Marker	Reference
1.5	3.0[b]	^{14}C-AMS	*Bard et al.* [1989]
81.5	8.8[b]	^{14}C-AMS	*Bard et al.* [1989]
121.5	12.7[b]	^{14}C-AMS	*Bard et al.* [1989]
124.0	12.64	^{14}C-AMS	KIA 0002[c]
130.0	12.91	^{14}C-AMS	KIA 0003[c]
136.0	13.54	^{14}C-AMS	KIA 0004-0006[c,d]
147.0	13.48	^{14}C-AMS	KIA 0007[c]
176.0	14.31	^{14}C-AMS	KIA 0008[c]
254.0	20.16	^{14}C-AMS	KIA 0009
263.0	19.75	^{14}C-AMS	KIA 0010[c]
272.0	20.46	^{14}C-AMS	KIA 0011-0013[c,d]
283.0	20.66	^{14}C-AMS	KIA 0014[c]
321.0	21.82	^{14}C-AMS	KIA 0015[c]
392.0	22.95	^{14}C-AMS	KIA 0017[c]
415.0	24.87	^{14}C-AMS	KIA 0018[c]
651.5	42.00[e]	3.13[e]	*Martinson et al.* [1987]

[a] All ages are corrected for reservoir age of 400 years.
[b] Radiocarbon date from nearby core SU81-18.
[c] KIA number is lab code listed in Table 1.
[e] Oxygen isotope event; age is in ^{14}C-kiloannums (see text).

For the six AMS ^{14}C 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 ^{14}C-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 kyr^{-1} at core SO75-26KL. Sedimentation rates increase to 55 cm kyr^{-1} and 32 cm kyr^{-1} 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 $^{230}Th_{ex}$ 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

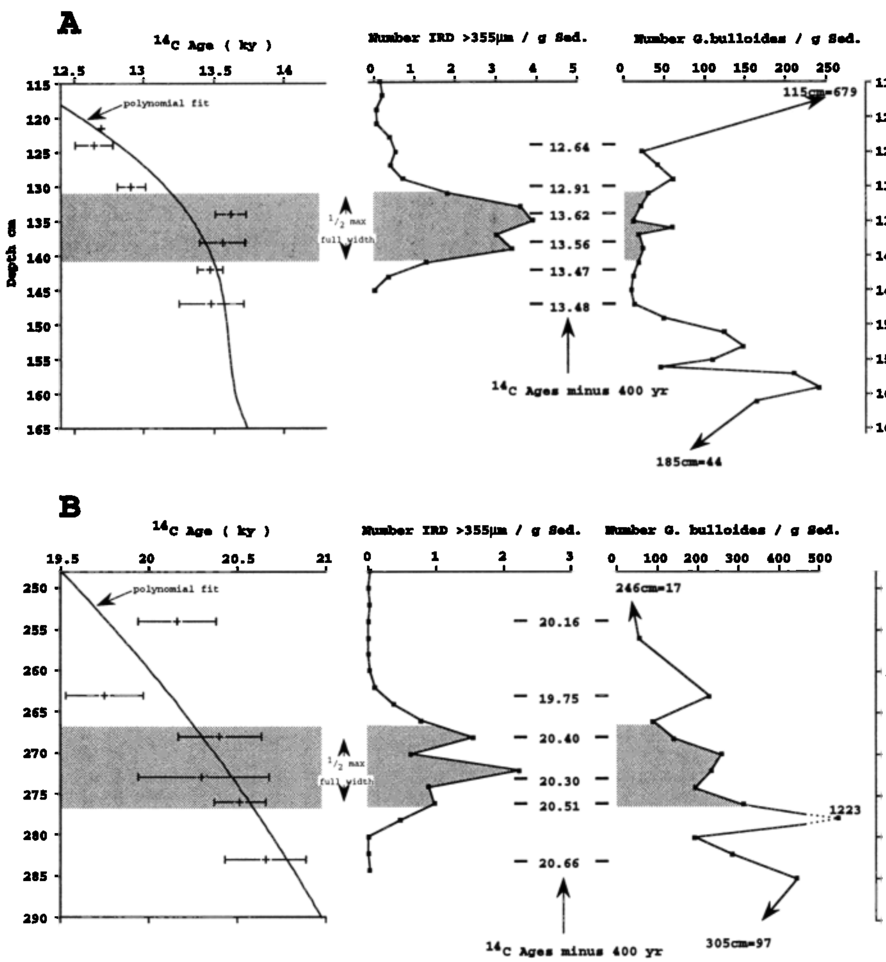

Figure 3. Age-depth function across (a) IRD layer 1 and (b) IRD layer 2. Within the range of error, ^{14}C-AMS ages do not change from below the IRD layers into the IRD maxima and imply rapid sedimentation during both events. The right-hand panel shows the abundance distribution for *G. bulloides* that was used for ^{14}C-AMS dating. See text for discussion.

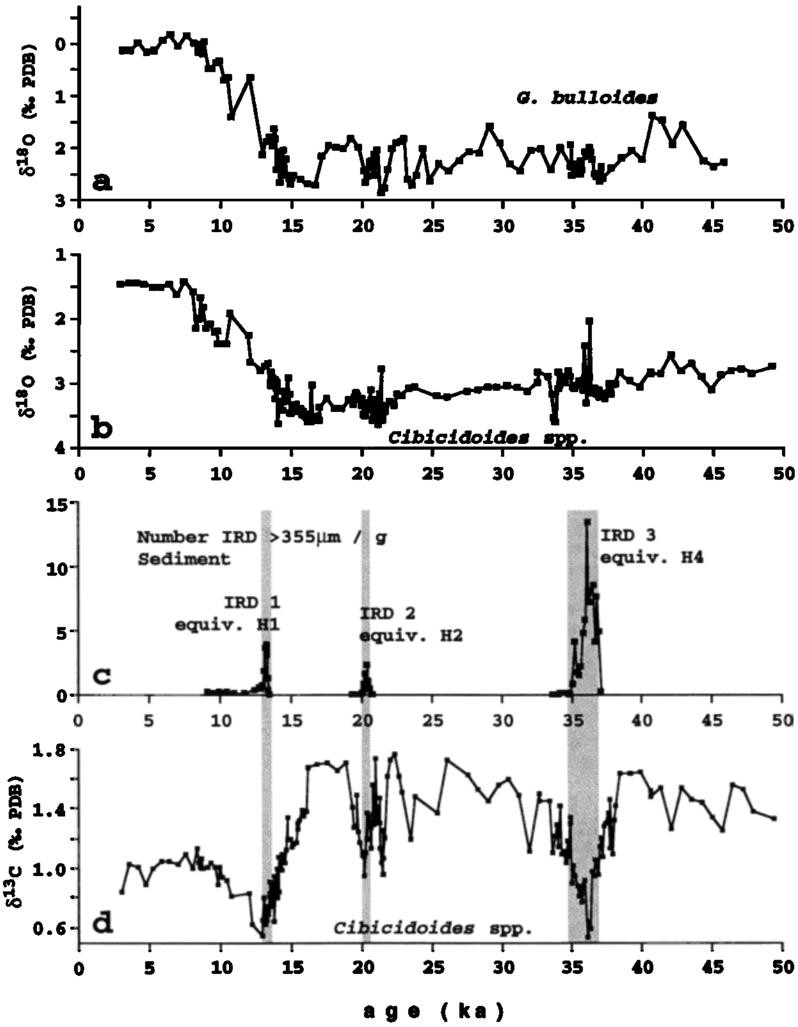

Figure 4. (a) Planktonic and (b) benthic $\delta^{18}O$ records and (c) IRD abundances on age scales, and (d) benthic $\delta^{13}C$ record. The $\delta^{13}C$ record displays increased values during the last glacial. Negative anomalies correlate with Heinrich events 1, 2, and 4. We use this correlation to infer reduced thermohaline overturn in the subpolar North Atlantic during these events.

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 ^{14}C-AMS dating shows some evidence for the first but not for the latter. Still, spacing of the ^{14}C data across both IRD layers is not sufficient to estimate ages of IRD boundaries unambiguously. Using ^{14}C-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 $^{230}Th_{ex}$ data [*Francois and Bacon*, 1994; *Thomson et al.*, 1995]. They are at the low side of estimates (250-1250 years) that have been derived from theoretical considerations of iceberg calving and sediment flux [*MacAyeal*, 1993; *Dowdeswell 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 where 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 nannoplank-

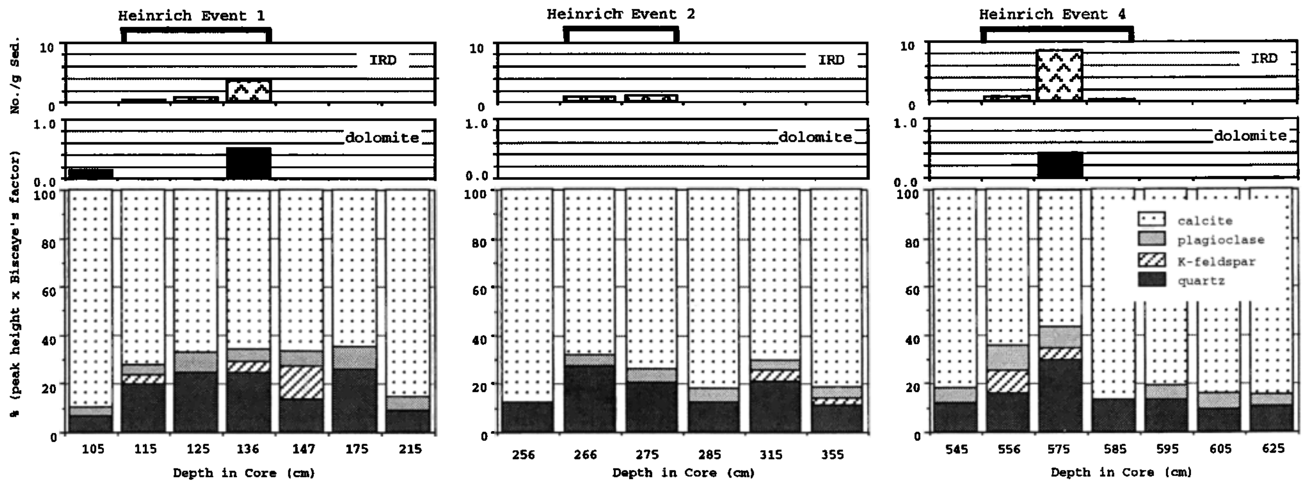

Figure 5. Clay mineral composition of IRD layers and sediments immediately above and below in core SO75-26KL. Dolomite is present in small quantities in IRD layers 1 and 3 (equivalent to Heinrich events H1 and H4), calcite is most likely of biogenic origin (foraminifera and nannoplankton).

ton. Dolomite has been detected in minor quantities in samples from IRD layers 1 and 3 (equivalent to H1 and H4; Figure 5). After treatment with acetic acid the carbonate peaks disappeared from the diffractograms, but the intensity peak of dolomite was still observed, confirming the presence of dolomite in both IRD layers.

Dolomite-bearing carbonate bedrock is widespread in the Laurentide domain, but small outcrops also exist in Ireland and Great Britain and along the western shores of Norway and Svalbard [*Bond et al.*, 1992, Figure 1]. Thus the presence of dolomite cannot uniquely document a Laurentide versus Scandinavian source for the icebergs. However, our XRD data demonstrate that the mineralogy of the IRD deposits off Portugal is not different from those of the open North Atlantic Heinrich layers. In addition to the apparent close chronological correlation with the Heinrich layers in the North Atlantic, we use this as further evidence that the IRD layers in core SO75-26KL are derived from the same iceberg surges and ice sources that caused the North Atlantic Heinrich events.

Heinrich Events and North Atlantic Thermohaline Instabilities

Numerical modeling has shown that the North Atlantic's thermohaline circulation constitutes a delicate balance between atmospheric forcing, northward heat transfer, and freshwater forcing [*Paillard and Labeyrie*, 1994; *Weaver and Hughes*, 1994; *Rahmstorf*, 1994, 1995]. From these simulations it was inferred that only small freshwater perturbations are sufficient to trigger significant decreases and ultimately a complete shutdown of convection. Convective instability in the North Atlantic thus appears to be dependent not so much 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 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 13°C/38.4 [*Zenk*, 1975] (Figure 6a). The density of this water is around 37.4 (σ_2=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=13°/35.6 [*Zenk*, 1975]) and Labrador Sea Water (LSW, a component of upper NADW; TS=3°/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 11°C/36.3 [*Zahn*, 1993]. Using end-member TS values of 13°/38.4 for MOW entering the North Atlantic, 13°/35.6 for NACW, and 3°/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 11°/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 $\delta^{18}O$ (δ_c) 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-δ_c diagrams provide a valuable tool for the interpretation of foraminiferal $\delta^{18}O$ values in that they add the constraint of vertical density stratification that helps to narrow down possible TS scenarios as inferred from foraminiferal $\delta^{18}O$ to physically plausible solutions [*Zahn and Mix*, 1991; *Labeyrie et al.*, 1992; *Sarnthein et al.*, 1995; *Weinelt et al.*, 1996]. The modern δ_c fractionation lines shown in Figure 6a have been computed using the paleotemperature equation of *Shackleton* [1974] and a δ_w:salinity slope of 0.5 (δ_w is the oxygen isotope composition of seawater) which results from a North Atlantic freshwater end-member around -18‰ in δ_w (SMOW). This slope represents the δ_w:salinity relationship for modern North Atlantic deep and middepth waters. We have fixed NADW at a benthic δ_c of +3.6‰ PDB and TS values of 2°C/34.9 that are commonly measured at coring sites in the North Atlantic [e.g., *Labeyrie et al.*, 1992]. For the TS values of 11°C/36.3 that have been measured in hydrocasts near the site of SO75-26KL [*Zahn*, 1993] we obtain a δ_c value of +2.1‰ PDB. This value is consistent with late Holocene $\delta^{18}O$ 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 δ_c fractionation lines have been computed using the same slope as for the modern δ_c lines (Figure 6b). Glacial maximum lower NADW ($LNADW_{LGM}$) was fixed to TS values of 0°C/35.8 and a δ_c value of +5.3‰ PDB (Figure 6b) [*Labeyrie et al.*, 1992; *Sarnthein et al.*, 1995; *Jung*, 1996]. From glacial-interglacial variations of planktonic foraminiferal $\delta^{18}O$ 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 4°-6°C [*Thiede*, 1978; *Thunell*, 1979; *Thunell and Williams*, 1989]. Using mean values of 5°C cooling and a salinity increase of 2 at the LGM, we have set MOW_{LGM} as it enters the North Atlantic to TS values of 8°/40.3 (Figure 6b).

To determine the potential contribution of MOW_{LGM} 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 ($UNADW_{LGM}$) [*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 $UNADW_{LGM}$ were similar to those of $LNADW_{LGM}$, 35.8. Using a benthic $\delta^{18}O$ 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 +1°C for $UNADW_{LGM}$ (Figure 6b). The TS values for $UNADW_{LGM}$ thus are 1°/35.8.

Few paleodata are available to trace the glacial-interglacial evolution of NACW as the second end-member for mixing with MOW_{LGM}. *Slowey and Curry* [1995] infer a cooling of 2°C 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 5° to >13°C 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 0°C in the northern North Atlantic [*Sarnthein et al.*, 1995]. Thus glacial NACW temperatures could have been anywhere between present-day values (13°C) 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 $NACW_{LGM}$ that formed in the area of the positive salinity anomaly at 51°-54°N inferred by *Duplessy et al.* [1991], with TS values of 6°/36, and (2) a warm $NACW_{LGM}$ that formed along the northern margin of the subtropical gyre (around 40°N), with TS values of 13°/37.5. In the first scenario, NACW would have been considerably colder than today (ΔT = -7°C) and salinity increased by 0.4, whereas in the second

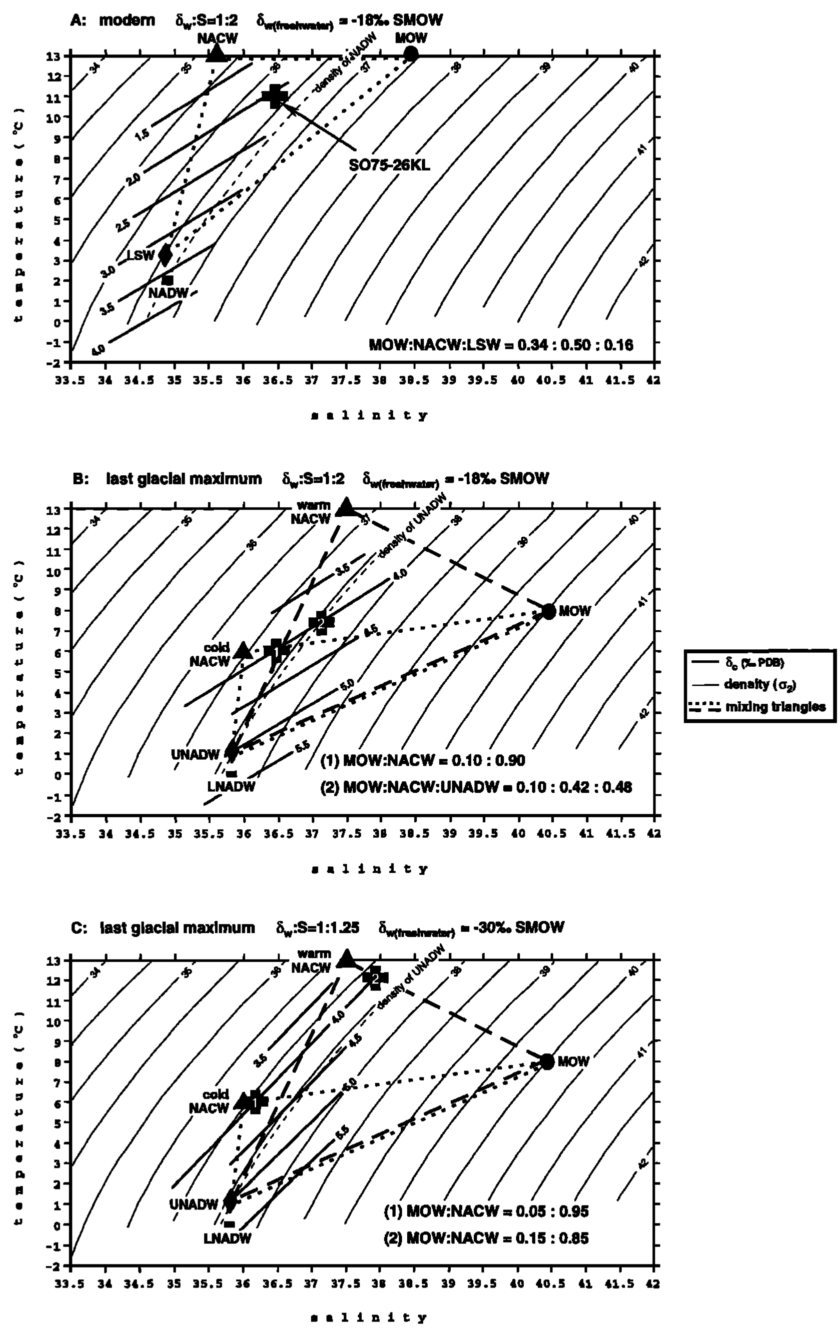

Figure 6. (a) TS diagram showing water mass end-members that contribute today to the site of core SO75-26KL. Density lines are for the 200 dbar surface (approximately 2000 m water depth). Isolines of δ_c equilibrium fractionation are computed using the paleotemperature equation of *Shackleton* [1974] and a δ_w:salinity slope of 0.5 for the modern North Atlantic; freshwater δ_w 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 $\delta^{18}O$ 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 δ_c fractionation is computed using the same parameterization as in Figure 6a and correcting for a mean glacial δ_w increase of 1.2‰. (c) Same as in Figure 6b, except that the d_w:salinity slope is set to 0.8, and freshwater δ_w 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 δ_c of +4‰ PDB has to be maintained to be consistent with the observed benthic $\delta^{18}O$ 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 $UNADW_{LGM}$. Maximum density for this middepth water mass is defined in the TS-δ_c diagram as the intercept between the +4‰ δ_c isoline and the 37.6 (σ_2) isopycnal of $UNADW_{LGM}$ 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 ‰ δ_c 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 UNADW and NACW. TS values are 7.2°/37.1, as defined by the intercept between the +4 ‰ δ_c isoline and the 37.6 (σ_2) isopycnal of $UNADW_{LGM}$ (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 use a more negative freshwater δ_w value of -30‰ (SMOW) to account for lower glacial precipitation temperatures in the North Atlantic region and a stronger contribution of glacial meltwater (Figure 6c). Using a δ_w value of -30‰ (SMOW), the slope of the δ_c lines in the TS field is steeper and the +4‰ δ_c fractionation line does not intersect the density isoline of UNADW. Thus the maximum possible MOW contribution is defined by the intercepts of the +4‰ δ_c fractionation line with the mixing lines between MOW and warm or cold NACW. Maximum MOW contributions are 5% if it mixes with cold NACW and 15% if it mixes with warm NACW (Figure 6c). That is, the contribution of glacial MOW to the middepth North Atlantic would have been reduced by 50% to more than 80% of its present contribution.

Sea level lowering of as much as 120 m during the last glacial would reduce the geometry of the Strait of Gibraltar thus reducing the through flow of MOW to the North Atlantic [*Bryden and Stommel*, 1984]. Therefore it appears plausible to assume that MOW_{LGM} contributed less than today to the hydrography at the upper Portuguese margin. This contention is further supported by physical considerations that TS values of 8°/40.3 estimated for MOW_{LGM} yield a density of 40 (σ_2) which is considerably higher than that of 37.6 for $UNADW_{LGM}$. The density contrast between MOW and underlying deep waters is thus increased from 0.7 today (Figure 6a) to 2.4 at the LGM (Figures 6b and 6c). If MOW still contributed to the hydrography of the shallow North Atlantic, significant mixing with less saline North Atlantic waters was required to lower MOW density and to allow it to flow at shallow depths. The combination of enhanced density, which requires intensive mixing with less saline Atlantic waters to increase its buoyancy, and reduced volume of MOW therefore supports the conclusion that MOW played a less significant role in the glacial maximum North Atlantic.

On the basis of elevated benthic $\delta^{13}C$ levels during the last glacial at core sites immediately west and south of the Gulf of Cadiz, *Zahn et al.* [1987] postulated a stronger influence of MOW on the North Atlantic's hydrography during the LGM. This hypothesis was subsequently supported by *Oppo and Fairbanks* [1987] using similar isotope evidence from Caribbean sediment cores. Even though it was recognized that the advection of MOW must have been reduced at the LGM because of lower sea level it was hypothesized that MOW left a stronger imprint on the North Atlantic than it does today because convection rates, and water mass renewal were thought to be significantly reduced in the glacial-maximum northern North Atlantic [*Zahn et al.*, 1987].

With new data from the northern North Atlantic and the Nordic Seas [*Duplessy et al.*, 1991; *Labeyrie et al.*, 1992; *Oppo and Lehman*, 1993; *Sarnthein et al.*, 1994, 1995; *Weinelt et al.*, 1996; *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-δ_c 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.

Benthic $\delta^{13}C$ Anomalies in Core SO75-26KL: Thermohaline Instabilities in the North Atlantic During Heinrich Events

Benthic $\delta^{13}C$ in core SO75-26KL is increased during the last glacial by 0.7‰ compared to Holocene values (Figure 4d). Increased benthic $\delta^{13}C$ 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; *Oppo and Lehman*, 1993, 1995]. For the North Atlantic it was speculated that a shift of convection cells from the Nordic seas to the northern North Atlantic was accompanied by cooling of surface waters and decreased evaporation rates that resulted in decreased salinities and enhanced buoyancy of the surface waters north of the glacial polar front [*Boyle and Keigwin*, 1987; *Oppo and Lehmann*, 1993]. As a result, surface densities in the northern North Atlantic were just high enough to allow for convection to middepths, around 1000-2000 m water depth.

Increased benthic $\delta^{13}C$ levels at core SO75-26KL from 1099 m water depth fit into this picture in that they document enhanced ventilation of glacial middepth levels at the upper Portuguese margin. As we have discussed above, advection of MOW to the site of core SO75-26KL during the last glacial was considerably reduced, so that the contribution of MOW to the increased benthic $\delta^{13}C$ levels was minor and the data primarily document ventilation from a northern North Atlantic source.

Decreases in benthic $\delta^{13}C$ of 0.7-1.1‰ that parallel the occurrence 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 $\delta^{13}C$ 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 $\delta^{13}C$ 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 element 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 compared 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 $\delta^{13}C$ anomalies in core SO75-26KL during IRD deposition confirms considerations of planktonic $\delta^{18}O$ 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 [*Maslin 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 supports 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 surface forcing in the course of Heinrich events and reduced thermohaline overturn in more detail, we compare the IRD and benthic $\delta^{13}C$ records from core SO75-26KL with the IRD and planktonic $\delta^{18}O$ records from Site 609 and the Greenland Ice Core Project (GRIP) ice core $\delta^{18}O$ record (Figure 7). Planktonic $\delta^{18}O$ at Site 609 and the GRIP ice core $\delta^{18}O$ record are used as monitors for meltwater flux to the northern North Atlantic and the evolution of North Atlantic climate [*Bond and Lotti*, 1995; *Dansgaard et al.*, 1993].

The records from core SO75-26KL and Site 609 do not exactly 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 enhancing 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 ^{14}C 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 $\delta^{18}O$ record which has a calendar year chronology thus remains tentative.

Despite the uncertainties in chronology, we find remarkable similarities between the marine records (Figure 7a). Absolute benthic $\delta^{13}C$ minima in core SO75-26KL during H4, before and near the end of H2, and immediately after H1, as well as the secondary $\delta^{13}C$ minima during the late stage of H4 and during H1, are mirrored in planktonic $\delta^{18}O$ 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 decreased thermohaline convection during the Heinrich events. A conspicious feature in the benthic $\delta^{13}C$ record is the gradual decrease of values that starts 1.5-2.5 kyr before the abrupt onset of IRD deposition during H1 and H4. After both events, benthic $\delta^{13}C$ gradually increases over 1.2-3 kyr. Assuming a close correlation between the rate of thermohaline circulation and benthic $\delta^{13}C$ levels, the gradual changes in benthic $\delta^{13}C$ 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 increasingly more negative $\delta^{18}O$ maxima of interstadials 11, 10, and 9 in the Greenland ice core $\delta^{18}O$ record (Figure 7). The benthic $\delta^{13}C$ 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, occurred abruptly at the end of H4, as is indicated by sudden increases in ice core $\delta^{18}O$ 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 $\delta^{13}C$ that would suggest a gradual strengthening of thermohaline circulation and associated heat flux to the North Atlantic.

A similar pattern of benthic $\delta^{13}C$ 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 $\delta^{13}C$ 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 thermohaline overturn in the North Atlantic [e.g., *Keigwin et al.*, 1991; *Lehman and Keigwin*, 1992; *Sarnthein et al.*, 1992]. Absolute minima in benthic $\delta^{13}C$ are recorded during and immediately after H1. The second minimum likely corresponds to meltwater pulse 1a (MWP 1a) of *Fairbanks* [1989]. The gradual increase in benthic $\delta^{13}C$ after MWP 1a documents the transition 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).

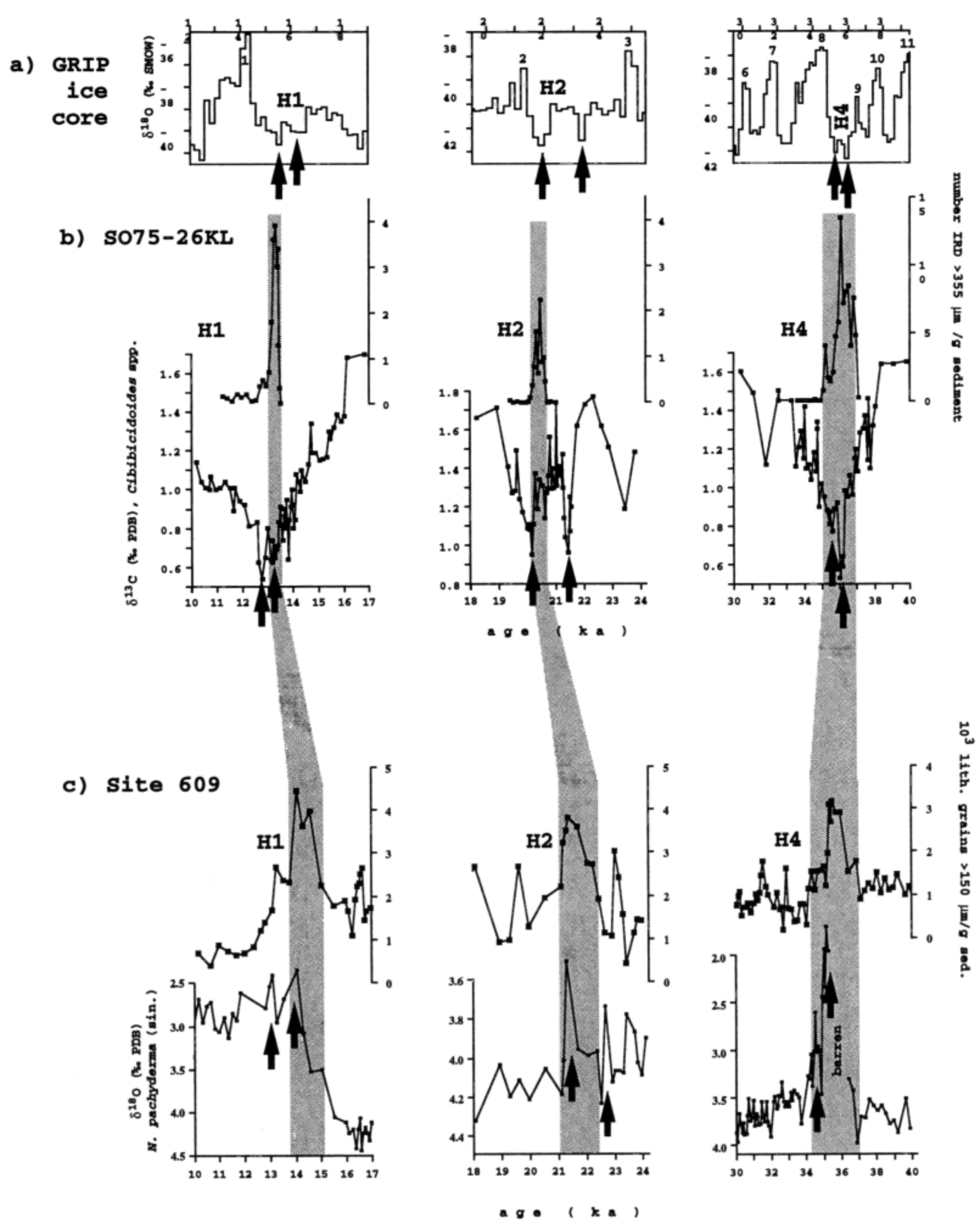

Figure 7. (a) Greenland Ice Core Project (GRIP) $\delta^{18}O$ around Heinrich events 1 (left), 2 (middle), and 4 (right). (b) Benthic $\delta^{13}C$ and IRD patterns in core SO75-26KL during these events. The close correlation of minimum benthic $\delta^{13}C$ and maximum IRD abundances is used to infer that thermohaline convection in the subpolar North Atlantic was reduced during peak iceberg and meltwater flow. (c) Planktonic $\delta^{18}O$ and IRD patterns at Site 609. Arrows mark episodes with minimum planktonic $\delta^{18}O$ at Site 609, minimum benthic $\delta^{13}C$ at core SO75-26KL, and minimum $\delta^{18}O$ in the GRIP ice core. The similarity of the foraminiferal isotope and IRD patterns at both sites and GRIP $\delta^{18}O$ suggests a close coupling of surface forcing and thermohaline overturn during peak meltwater flow. Early decreases and late resumption of fully increased values around H1 and H4 indicate that thermohaline reduction started well before the events and lasted much longer than the events. Inferred gradual spin-up of thermohaline circulation over periods of 1-3 kyr after all three Heinrich events is in direct contrast to sudden warmings seen in the GRIP data.

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 $\delta^{13}C$ 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 $\delta^{13}C$ 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 $\delta^{13}C$ 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 North Atlantic. The more gradual increase in benthic $\delta^{13}C$ suggests that thermohaline circulation wound up slowly and resumed its full strength as much as 3 kyr 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 $\delta^{13}C$ 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 $\delta^{18}O$ in TS diagrams with computed δ_C 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 $\delta^{13}C$ 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 ^{14}C-AMS ages for IRD 1 and 2 of 13.54±.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 $\delta^{13}C$ 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 $\delta^{18}O$ at Site 609 and minima in thermohaline circulation as indicated by minimum benthic $\delta^{13}C$ 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 $\delta^{13}C$ 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 northward 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 $\delta^{13}C$ 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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