Surface- and deep-water hydrography and meltwater events in the mid-latitude North Atlantic Ocean over the past 160,000 years

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An investigation involving stable oxygen and carbon isotopes on core M15612 from the Mid-Atlantic Ridge (44°N, 26°W) shows that distinct negative δ18O anomalies punctuate the planktonic isotope records and correlate with the ice-rafted detritus (IRD) events. Within most IRD layers, detrital carbonate contributes between 1 and 13% and supports the contention that the layers are the southern equivalents of the northern North Atlantic Ocean’s Heinrich events. Anomalies are strongest for the Neogloboquadrina pachyderma (sinistral coiling) isotope record. Systematic changes in the δ18O offset between Globigerina bulloides and Globorotalia inflata signify variations in the mid-latitude thermocline structure. In conjunction with negative benthic δ13C anomalies, the data document a stronger contribution of an 18O-depleted, nutrient-rich water mass during the Heinrich events. The δ13C amplitude of >1% between 25 and 57 ka indicates changes between northern-source (North Atlantic Deep Water) and southern-source (Antarctic Bottom Water) water masses at this site. The Heinrich layers in the mid-latitude core are correlative with those from a core off Portugal. This synchronicity may have resulted from common changes in a North Atlantic thermohaline switch.

INTRODUCTION

A series of distinct IRD layers in the North Atlantic Ocean punctuates glacial and deep-sea sediments (Heinrich 1988; Broecker et al. 1992). The IRD layers, commonly named “Heinrich” layers, have a mineral assemblage containing often more than 20% detrital carbonate material. The Heinrich layers are found mostly between 45° and 55°N latitude (Bond et al. 1992). The detrital carbonate fragments in the Heinrich layers decrease in abundance eastwards, pointing to the Laurentide ice sheet as the most likely source, the ice shedding fragments by way of massive iceberg discharges (Broecker et al. 1992). At times, such icebergs and their discharge reached as far as the Portuguese margin (e.g. Bond et al. 1992; Grouset et al. 2000).

It seems that during the last glacial period, thermohaline circulation was broken repeatedly by massive melting of icebergs (Maslin et al. 1995). This may have been because the production of North Atlantic Deep Water (NADW) ceased, or because the region of sinking changed (Sarnthein et al. 1994). On the basis of depleted benthic δ13C and estimates of decreased sea-surface temperature from faunal assemblages (Maslin et al. 1995), the glacial circulation of the North Atlantic underwent convection breakdown during the Heinrich events.
Some authors (e.g. Lebreiro et al. 1996; Zahn et al. 1997; Baas et al. 1998; Grousset et al. 2000; Thomson et al. 2000) have extended the study of Heinrich layers to the eastern North Atlantic and the far eastern boundary of the presumed iceberg flow. These studies demonstrate that cores from the eastern sector of the North Atlantic contain signals consistent with the Heinrich events, but strongly attenuated.

In this study, we present paleo-oceanographic data from high-resolution measurements of cores from the Mid-Atlantic Ridge, northern North Atlantic and Portuguese margin. Particularly, core M15612 was used to monitor surface- and deep-water properties in association with North Atlantic meltwater events over the past 160,000 years. Our profiles make possible the monitoring of even small changes in surface and deep hydrography on glacial-interglacial to millennial time scales. Finally, the Heinrich events in the cores are correlated and discussed.

MATERIALS AND METHODS

Core M15612 was collected from the Mid-Atlantic Ridge north of the Azores (Fig. 1; Table 1). The site of core M15612 is thus close to the glacial polar front, and its depth position is at the hydrographic front between glacial deep and bottom water masses (Ruddiman 1977). Core SO75-26KL, from the southern Portuguese margin at 37°49'N, and DSDP site 609, at 49°53'N, were used to correlate fluctuations of the IRD events with those of core M15612.

For coarse fraction analysis, core M15612 was continuously sampled at 5 cm intervals, and additional samples were taken at 1.5 cm intervals in the Heinrich layers in order to obtain a higher stratigraphic resolution. Coarse sediment grains (>63 μm) were counted to determine the ratios of biogenic and terrigenous components. An estimate of the relative abundance of planktonic foraminifera (>125 μm) was also made.

Stable oxygen/carbon isotope analyses were performed on several foraminifera species: about 25 specimens of each of the planktonic species *Globigerina bulloides*, *Globorotalia inflata* and *Neogloboquadrina pachyderma* and 2 to 6 specimens of each of the benthic foraminifera species *Cibicidoides wuellerstorfi* and *Cibicidoides kullenbergi*. Prior to isotope analyses, the tests were cracked open and ultrasonically cleaned in reagent grade methanol, dried at 40°C, and transferred to the KIEL CARBOPREP device. The evolved CO₂ was analyzed using the mass spectrometer FINNIGAN MAT 251 in Kiel, Germany. The long-term reproducibility was 0.07‰ for δ¹⁸O and 0.04‰ for δ¹³C as calculated from replicate analyses of internal carbonate standards. All isotope data are expressed in the standard δ notation, which is in ‰ relative to the PDB scale.

The ¹⁴C ages of core M15612 were compiled from the data of Kiefer (1998), which were measured by AMS using the 3MV Tandetron system at the Leibniz Laboratory, University of Kiel (Nadeau et al. 1997). Monospecific samples of *G. inflata* and *N. pachyderma* containing 300 to 1200 tests in the size fractions 315–400 μm (for *G. inflata*) and 250–315 μm (for *N. pachyderma*) were used (Table 2). In order to minimize the bias of bioturbational mixing, samples were picked at relative maxima of the species dated.

Stratigraphic control of core M15612 was mainly based on AMS ¹³C ages (Table 2) and foraminiferal oxygen isotope stratigraphy (Martinson et al. 1987). The initial age model was based on the benthic δ¹⁸O decrease at the end of Younger

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**Table 1. Location of the cores used in this study**

<table>
<thead>
<tr>
<th>Core No.</th>
<th>Latitude (N)</th>
<th>Longitude (W)</th>
<th>Water depth (m)</th>
<th>Local bottom water mass¹</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>M15612</td>
<td>44°21'36&quot;</td>
<td>26°32'36&quot;</td>
<td>3050</td>
<td>NADW</td>
<td>This study</td>
</tr>
<tr>
<td>M15637</td>
<td>28°53'00&quot;</td>
<td>19°14'00&quot;</td>
<td>1884</td>
<td>NADW</td>
<td>Zahn (1986), Kiefer (1998)</td>
</tr>
<tr>
<td>SO75-26KL</td>
<td>37°49'18&quot;</td>
<td>9°19'54&quot;</td>
<td>1099</td>
<td>NADW/MOW</td>
<td>Park (1994), Zahn et al. (1997)</td>
</tr>
<tr>
<td>DSDP 609</td>
<td>49°53'00&quot;</td>
<td>24°14'00&quot;</td>
<td>3884</td>
<td>NADW</td>
<td>Bond et al. (1992, 1993)</td>
</tr>
</tbody>
</table>

¹ NADW = North Atlantic Deep Water, MOW = Mediterranean Outflow Water
Table 2. $^{14}$C-AMS ages for core M15612

<table>
<thead>
<tr>
<th>Depth interval $^1$</th>
<th>Mean depth $^1$</th>
<th>Species</th>
<th># analyzed</th>
<th>$^{14}$C age (yr)</th>
<th>Error +/-</th>
<th>Calendar age (ka)</th>
<th>References for calendar age</th>
</tr>
</thead>
<tbody>
<tr>
<td>0–1.5</td>
<td>0.75</td>
<td>G. inflata</td>
<td>500</td>
<td>2,229</td>
<td>31 31</td>
<td>2.23</td>
<td>Stuiver and Pearson (1993)</td>
</tr>
<tr>
<td>8.0–9.0</td>
<td>8.50</td>
<td>G. inflata</td>
<td>500</td>
<td>4,074</td>
<td>28 28</td>
<td>4.53</td>
<td>Stuiver and Pearson (1993)</td>
</tr>
<tr>
<td>33.0–34.5</td>
<td>34.75</td>
<td>G. inflata</td>
<td>500</td>
<td>12,091</td>
<td>56 56</td>
<td>14.32</td>
<td>linear interp.</td>
</tr>
<tr>
<td>44.0–45.5</td>
<td>44.75</td>
<td>N. pachyderma</td>
<td>1200</td>
<td>14,955</td>
<td>63 62</td>
<td>15.97</td>
<td>linear interp.</td>
</tr>
<tr>
<td>52.0–53.0</td>
<td>52.50</td>
<td>N. pachyderma</td>
<td>1000</td>
<td>16,465</td>
<td>108 107</td>
<td>17.90</td>
<td>linear interp.</td>
</tr>
<tr>
<td>68.0–69.0</td>
<td>68.50</td>
<td>N. pachyderma</td>
<td>650</td>
<td>19,050</td>
<td>110 110</td>
<td>22.08</td>
<td>linear interp.</td>
</tr>
<tr>
<td>72.0–73.0</td>
<td>72.50</td>
<td>N. pachyderma</td>
<td>900</td>
<td>18,394</td>
<td>107 106</td>
<td>23.13</td>
<td>linear interp.</td>
</tr>
<tr>
<td>74.0–75.0</td>
<td>74.50</td>
<td>N. pachyderma</td>
<td>750</td>
<td>20,520</td>
<td>180 180</td>
<td>23.65</td>
<td>H2 top (GISP2)</td>
</tr>
<tr>
<td>81.5–83.0</td>
<td>82.25</td>
<td>N. pachyderma</td>
<td>850</td>
<td>21,255</td>
<td>108 106</td>
<td>24.58</td>
<td>linear interp.</td>
</tr>
<tr>
<td>86.0–87.5</td>
<td>86.75</td>
<td>N. pachyderma</td>
<td>1000</td>
<td>22,180</td>
<td>102 101</td>
<td>25.62</td>
<td>linear interp.</td>
</tr>
<tr>
<td>98.0–99.0</td>
<td>98.50</td>
<td>G. inflata</td>
<td>380</td>
<td>23,938</td>
<td>139 137</td>
<td>28.33</td>
<td>linear interp.</td>
</tr>
<tr>
<td>106.5–107.5</td>
<td>107.00</td>
<td>N. pachyderma</td>
<td>1200</td>
<td>26,638</td>
<td>287 277</td>
<td>30.18</td>
<td>H3 base (GISP2)</td>
</tr>
<tr>
<td>133.5–135.0</td>
<td>134.25</td>
<td>G. inflata</td>
<td>300</td>
<td>32,027</td>
<td>273 264</td>
<td>37.11</td>
<td>linear interp.</td>
</tr>
<tr>
<td>150.0–151.0</td>
<td>150.50</td>
<td>N. pachyderma</td>
<td>1000</td>
<td>34,814</td>
<td>503 473</td>
<td>39.92</td>
<td>linear interp.</td>
</tr>
</tbody>
</table>

$^1$ Depth intervals and mean depth in cm.

Dryas cold event (11.6 cal. ka; Meese et al. 1994) and at the onset of stage 4 maximum (71.1 cal. ka). Holocene age control points were established from $^{14}$C dates and then converted to calendar ages (Stuiver et al. 1998).

RESULTS

Stratigraphy

The upper IRD records of core M15612 show 6 distinct IRD maxima, which reveal clearly Heinrich layers 1 (H1) to 6 (H6), although two peaks (H3 and H6) are of relatively low magnitude (Fig. 2). An additional age control point was obtained by assigning the IRD records in core M15612 to the well-documented ages of the corresponding oxygen-isotope minima in the GISP ice core. In this way, $^{14}$C ages of the Heinrich events were obtained and H1, H2, H3 and H4 were dated at 12.3–15.0, 20.0–21.3, 26.6 and 33.8–35.4 ka, respectively.

According to the age models, the stable isotope records of the core cover the time interval from the penultimate full-glacial stage 6 to the Holocene. The glacial-interglacial transitions of Termination 1 and II are documented by $\delta^{18}$O shifts of up to 2% in the planktonic and benthic records (Fig. 2a). The largest isotope shifts are documented in the $\delta^{18}$O record obtained from G. bulloides, which is a near-surface dweller, and thus records a strong temperature signal. G. inflata, on the other hand, lives at greater water depths and records attenuated temperature signals (Fig. 3a). Systematic changes in the $\delta^{18}$O offset between these planktonic species signify changes in the mid-latitude thermocline structure. Negative excursions of the planktonic and benthic $\delta^{18}$O records occur together with significant increases in abundance of N. pachyderma, pointing to the influence of meltwater or subpolar water of the site of core M15612.

Following on from the above results, we calculate that, over the past 160,000 years, mean sedimentation rate was 3.05 cm/kyr (Fig. 2). This rate is very low compared with those at the upper continental slope off Portugal (15.9 cm/kyr for core SO75-26KL): sedimentation rates at the continental shelf and slope are one order of magnitude higher than those computed for Mid-Atlantic core M15612 and the North Atlantic cores (e.g. 2.5 cm/kyr for core M15637; Kiefer 1998). The low sedimentation rate of core M15612 is surely due to pelagic accumulation rates in open ocean plains. However, sedimentation rates increase dramatically to 18 cm/kyr during H1 and 12 cm/kyr during H2, respectively, reflecting twofold to sixfold increase in sediment flux during Heinrich events (Fig. 2c).

Composition of coarse fraction

Core M15612 contains predominantly pelagic foraminiferal ooze with minor contributions from pteropod shells. This uniform sediment distribution is interrupted only in the upper section, between 10 and 40 ka, and near the base of the core, where the Heinrich layers occur (Fig. 3c). Detrital carbonate contributes 1–13% of the detrital fraction of the Heinrich layers. H1 and H2 consist of over 90% terrigenous material, of which detrital carbonate comprises up to 10%. Within the Heinrich layers, the terrigenous fraction accounts for 25% (H6) to nearly 100% (H2), H3 and H6 are more pronounced in the >63 pm fraction than in the >125 pm fraction (Fig. 3c). The largest part of the Heinrich layers consist of transparent angular quartz grains, angular to subangular feldspar and dolomite particles. The IRD layers in the deeper section between 125 and 150 cm are composed of 20 to 45% IRD fractions, but these do not contain any detrital carbonates (see below).
Faunal analyses of core M15612 show that the Heinrich layers are associated with a distinct pattern of foraminiferal abundances (Fig. 3b). *N. pachyderma* tends to abundance maxima in or near the Heinrich layers. In the same intervals, the rest of the planktonic foraminiferal community almost disappears, further supporting the existence of an exotic polar or subpolar environment that could be tolerated only by *N. pachyderma*.

**δ¹³C anomalies in core M15612**

During the past 70,000 years, δ¹³C records from core M15612 show clear response to the Heinrich events and variations in sea-surface temperature (Fig. 4a, b). Therefore, these δ¹³C values primarily monitor variability of surface- and deep-water masses, although they also trace the intensity of ventilations in the North Atlantic.

Planktonic (*G. bulloides*) δ¹³C is depleted up to 1% compared to benthic (*C. wuellerstorfi*) δ¹³C, but the benthic δ¹³C values commonly decrease 0.2–0.5‰ during the Heinrich events. The close correlation of minimum δ¹³C values and maximum IRD abundances in core M15612 suggests that thermohaline convection in the subpolar North Atlantic was reduced during massive iceberg discharge and meltwater flow. A similar pattern was also observed on the Portuguese margin (Zahn et al. 1997). However, superimposed on the benthic δ¹³C patterns is a long-term trend with positive δ¹³C values around 1–1.5‰ between 25 and 57 ka. The δ¹³C amplitude of more than 1‰ in this time interval indicates changes between northern-source (North Atlantic Deep Water) and southern-source (Antarctic Bottom Water) water masses at this site.
**DISCUSSION**

**Correlation between IRD layers at Mid-Atlantic core M15612 and DSDP 609**

As mentioned above, all IRD layers from core M15612 are characterized by the maximum abundance of lithic fragments, abundance maximum of *N. pachyderma* and relative high percentages of dolomite except for H3 and H6, in which the IRD layers are detrital carbonate free. Examination using optical microscopy did not reveal any detrital calcite in the IRD layers of the core. It is generally thought that, through progressive melting of the icebergs, detrital calcite-rich carbonates would be released and would partly dissolve in the water column. The fact that IRD layers in core M15612 contain little or no detrital calcite could be explained by the possibility that ice-rafted detrital calcite is more susceptible to dissolution than dolomite.

Compared to DSDP 609, the Heinrich layers in core M15612 contain only small amounts of dolomite in the detrital carbonate fraction. The two IRD layers in the lower part of the core (between 125 and 144 ka) show no indication of the presence of detrital carbonates at all. The IRD layers in the DSDP 609 indicate that H1, H2, H4 and H5 contain layers with 20–25% detrital carbonates, whereas carbonate is absent in H3 and H6, as shown in core M15612 (Fig. 5).

Similar to the signals observed at DSDP 609, Heinrich events H1 to H6 in core M15612 coincide with distinct minima of the planktonic foraminiferal fauna and maximum abundances of the left-coiling form of *N. pachyderma* (Fig. 3). However, negative anomalies of the *N. pachyderma* δ¹⁸O values in the core...
Fig. 4. Correlation of the data between 10-70 ka in core M15612: a) Foraminiferal δ¹³C records and IRD patterns in the >63 μm and >125 μm fractions. b) SST for summer and winter estimated from planktonic foraminiferal assemblage (S1MMAX; for details see Kiefer (1998)).

Fig. 5. Occurrence of the Heinrich events (horizontal bars) in cores used in this study. Solid bars indicate IRD events with detrital carbonates, shaded bars indicate IRD layers with no detrital carbonates.
are recorded from H1 to H4. For H5 and H6, these values show only weak response to the IRD event. Surprisingly, *N. pachyderma* shows the same δ¹⁸O response as recorded by *G. bulloides* and *G. inflata*. The observation that *G. bulloides* and *G. inflata* show different isotope signals from that of *N. pachyderma* may be the result of ecological factors: that is, *N. pachyderma* prefers a polar environment similar to that existing during the Heinrich events, whereas *G. bulloides* and *G. inflata* are boreal species and presumably could not tolerate polar environments. Therefore, the last two species decreased sufficiently in abundance (or vanished completely) to be no longer reliable recorders of surface-subsurface water conditions. If this were indeed the case, the δ¹⁸O values of both species during the Heinrich events would be strongly disturbed by bioturbation, which mixes in shells from pre- and post-IRD event sections, in which the two species are abundant.

Subtle differences between the IRD records of core M15612 and DSDP 609 may be ascribed to the fact that core M15612 is located close to the southern limit of maximum iceberg flow, whereas DSDP 609 is located in the maximum flow path of icebergs. Also, the magnitude of the various IRD events in core M15612 would depend on the position of the axis of the iceberg flow path and this, in turn, may be a function of climate and magnitude of iceberg shedding.

**Correlation between IRD layers from the open North Atlantic and Upper continental margin off Europe and Africa**

The records of core SO75-26KL from the upper Portuguese continental margin document the existence of distinct IRD layers during MIS (Marine-oxygen Isotope Stage) 2/3 (Zahn et al. 1997; Schönfeld and Zahn 2000). On the other hand, core M15637 does not contain any Heinrich-IRD layers (Figs. 1, 5). The IRD layers in core SO75-26KL contain less detrital minerals than those observed at the Mid-Atlantic core M15612. The mineral assemblage of some IRD layers in the core is very similar to that of the Heinrich layers in core M15612. For example, the dolomite contained in H1 and H4 in core SO75-26KL suggests that both layers are correlative with the open North Atlantic events. In contrast, H2 in core SO75-26KL does not contain detrital carbonate (Fig. 5). This is in contrast to the findings from the open North Atlantic, where H2 contains significant amounts of dolomite; enhanced dolomite content is also documented for H2 in core M15612 (Fig. 3). In fact, the broader dolomite maximum observed for H2 in core M15612 implies a stronger input of dolomite compared to the other peaks in the same core. Yet, dolomite is missing from what is inferred here to be the Portuguese-margin equivalent of H2. Thus, iceberg flow during this full-glacial event may not have been as severe as the one recorded for Termination I, so that only a few icebergs would have reached the Portuguese margin, and at this distance they would have lost most of their original paleo-oceanographic signals.

The close correlation between the benthic δ¹³C decreases and occurrences of Heinrich events from the eastern North Atlantic are suggestive of mid-depth circulation changes that were related to changing surface ocean conditions (Zahn et al. 1997). Also, δ¹³C records from the study area support the possibility of a linkage between sea-surface conditions and thermohaline overturns in the North Atlantic. A similar pattern of stable isotope overturns from both the area north of the Azores and an area of the eastern North Atlantic off Portugal suggests common paleoclimatic changes during the most recent glacial-interglacial climate cycles, indicating large-scale forcing elements in the climate system.

**CONCLUSIONS**

Coarse-fraction counting shows that IRD layers exist in the upper section of core M15612, which are clearly associated with H1 to H6 observed at open North Atlantic sites. These layers show paleo-oceanographic features which are identical to Heinrich layers. These include maximum abundance of planktonic foraminifera *N. pachyderma* and occurrence of detrital carbonates. Except for H3, H6 and IRD layers during Termination II, the dolomite accounts for between 1-13% of terrigenous material.

Salinity excursions, as suggested by distinctly negative δ¹⁸O anomalies, are documented in all IRD layers. According to the results from δ¹³C analysis, benthic δ¹³C values decrease during the Heinrich events, documenting reduced ventilation in the North Atlantic. In contrast, benthic δ¹³C values increase 1-1.5% between 25 and 57 ka, implying hydrographic changes between northern-source (NADW) and southern-source (AABW) water masses at this site.

The Heinrich layers in core M15612 are correlative with the IRD layers from core SO75-26KL and DSDP 609. These layers from the Portuguese margin correlate with H1, H2 and H4 from core M15612 and DSDP 609. This synchronicity may have been derived from common changes in a North Atlantic thermohaline switch.

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