Centennial- to millennial-scale ice-ocean interactions in the subpolar northeast Atlantic 18–41 kyr ago

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In order to monitor the evolution of the British-Irish Ice Sheet (BIIS) and its influence in surface ocean structure during marine isotopic stages (MIS) 2 and 3, we have analyzed the sediments recovered in core MD04-2829CQ (Rosemary Bank, north Rockall Trough, northeast Atlantic) dated between ~41 and ~18 ka B.P. Ice-rafted debris flux and composition, 40Ar/39Ar ages of individual hornblende grains, multispecies planktonic stable isotope records, planktonic foraminifera assemblage data and faunal-based sea surface temperatures (SSTs) demonstrate a close interaction between BIIS dynamics and surface ocean structure and water properties in this region. The core location lies beneath the North Atlantic Current (NAC) and is ideal for monitoring the shifts in the position of its associated oceanic fronts, as recorded by faunal changes. These data reveal a succession of BIIS-sourced iceberg calving events related to low SST, usually synchronous with dramatic changes in the composition of the planktonic foraminifera assemblage and with variations in the stable isotope records of the taxa Neogloboquadrina pachyderma (sinistral coiling) and Globigerina bulloides. The pacing of the calving events, from typically Dansgaard-Oeschger millennial timescales during late MIS 3 to multicentennial cyclicity from ~28 ka B.P., represents the build-up of the BIIS and its growing instability toward Heinrich Event (HE) 2 and the Last Glacial Maximum. Our data confirm the strong coupling between BIIS instabilities and the temperature and salinity of surface waters in the adjacent northeast Atlantic and demonstrate the BIIS’s ability to modify the NAC on its flow toward the Nordic Seas. In contrast, subsurface water masses were less affected except during the Greenland stadials that contain HEs, when most intense water column reorganizations occurred simultaneously with the deposition of cream-colored carbonate sourced from the Laurentide Ice Sheet.


I. Introduction

Since the discovery of millennial and faster climate variability in Greenland ice cores in the early 1990s [Dansgaard et al., 1993; Greenland Ice Core Project Members, 1993; Grootes et al., 1993], understanding of these climatic oscillations has increased rapidly. The Dansgaard-Oeschger (DO) cycles are the most prominent manifestation of climate instability in the recent geological past as they represent substantial decadal air temperature fluctuations between Greenland stadials (GSs) and interstadials (GIs) of the order of 5°–10°C in the North Atlantic region [Dansgaard et al., 1993]. They coincide with variations in the properties of oceanic water masses recorded in sediment cores in the North Atlantic and beyond. These records reveal the close links between atmospheric changes and ocean dynamics [e.g., Bond et al., 1992, 1993; Cortijo et al., 1997; Keigwin and Lehman, 1994; Vidal et al., 1997]. An increasing number of sediment records recovered offshore the British Isles [Dickson et al., 2008; Haapaniemi et al., 2010; Hibbert et al., 2010; Knutz et al., 2001, 2002, 2007; Peck et al., 2006, 2007a, 2007b, 2008; Peters et al., 2008, 2010; Scourse et al., 2000, 2009] and in the Nordic Seas [Fronval et al., 1995; Kuipers et al., 1998; Lekens et al., 2006, 2009; Meland et al., 2008; Rasmussen et al., 1996a, 1996b; Rasmussen and Thomsen, 2004] show that the Northwestern European Ice Sheet (NWEIS) was an important element of the climate of the last glacial, in that it played a major
role in the freshwater release that led to disruptions in deep-water convection and influenced the Atlantic Meridional Overturning Circulation (AMOC). Of all NWEIS sectors, the British-Irish Ice Sheet (BIIS) was directly situated in, and highly dependent on, the track of moisture and heat transport derived from the North Atlantic Current (NAC) [Hansen and Østerhus, 2000]. As a result, the dynamics of the small-sized, fast responding BIIS were tightly coupled to variations in such transports, making nearby oceanic sediment cores highly sensitive records in which to monitor ocean-ice interactions. Evidence of BIIS growth and instability has been documented in high-resolution sedimentary sequences along its margins [e.g., Dickson et al., 2008; Knutz et al., 2002, 2007; Peck et al., 2006, 2007a, 2007b, 2008; Scourse et al., 2009] and recent modeling studies confirm that the BIIS was the dominant source of ‘background’ icerafted debris (IRD) supplied to the North Atlantic during the last glacial [Bigg et al., 2010], but additional data are needed to achieve a comprehensive regional view of ocean-BIIS interactions.

Here we present new high-resolution paleoceanographic records from a sediment core recovered close to the northwestern margin of the BIIS, aimed at extending results of previous records [Knutz et al., 2002, 2007] well into marine isotopic stage (MIS) 3 in order to depict in fine detail the sequence of BIIS growth and decay between 41 and 18 ka B.P. We present extensive $^{40}$Ar/$^{36}$Ar dating results on hornblende grs than, together with IRD composition, enable IRD to be evaluated. We also compare IRD flux with multispecies planktonic foraminifera stable isotope records and assemblage data, and with faunal-based sea surface temperature estimates (SST) as a means of closely assessing the interactions between ice sheet dynamics and surface water structure. Bringing these two approaches together enables a comprehensive and high-resolution view of the role of the BIIS in the forcing of past short-term climate variability.

### 2. Core Location and Oceanographic Setting

This sediment core (MD04-2829CQ) was sampled every centimeter across the interval 300–1008 cm, yielding a total of 708 sample slices of 1 cm thickness. Each sample was wet sieved with distilled water over a 63 μm mesh and both coarse and fine fractions recovered, dried at 40°C and weighed. Coarse (>63 μm) fractions were split into two aliquots for stable isotope, and petrological and micropaleontological analyses, which were carried out every 2 to 4 cm.

### 3. Materials and Methods

#### 3.1. Core Sampling

The resulting planktonic foraminifera assemblage counts were employed to estimate SST at 10 m water depth by running a transfer function based on a back propagation
artificial neural network (ANN) [Malmgren et al., 2001] trained on the North Atlantic MARGO Project data set [Kucera et al., 2005]. The same data set was used to apply a Modern Analogue Technique (MAT) [Prell, 1985] to the faunal counts. For the ANN, a set of ten neural networks was considered, providing 10 different mean annual SST reconstructions, which were averaged to estimate the SSTs and standard deviations from the calibration data set. The MAT was applied using a set of ten modern analogs, allowing the calculation of a similarity index for MAT SSTs.

3.4. Dating of $^{40}$Ar/$^{39}$Ar on Hornblende Grains

[11] A total of 421 individual hornblende fragments from the >150 μm fraction were picked when present and analyzed to provide for provenance data of IRD. Grains from the same horizon were grouped in the same sample, yielding a total of 184 samples covering the interval from ~18 to ~40 ka B.P., that were sent to the Cd-lined core facility (CLICIT) at the Oregon State University reactor (USA) for their coirradiation with the hornblende monitor standard Mmh5 (525 Ma [Samson and Alexander, 1987]). Argon activity analyses were performed at the Ar Geochemistry Laboratory, Lamont-Doherty Earth Observatory (LDEO, USA), where individual grains were fused with a CO$_2$ laser and ages were calculated from the resulting Ar isotope ratios, corrected for mass discrimination, interfering nuclear reactions, procedural blanks and atmospheric Ar contamination following the standard procedure of McDougall and Harrison [1999].

3.5. Isotope Measurements

[12] Aliquots for stable isotope analyses were dry-sieved over 250 and 355 μm meshes. Around 10–15 specimens of the planktonic foraminifera Neogloboquadrina pachyderma (sinistral coiling), mostly the encrusted type, were obtained every 2 cm, while up to 15–20 individuals of Globigerina bulloides were picked when possible. Around 25 specimens of Turborotalita quinqueloba (>150 to <250 μm) were also
Table 1. AMS Radiocarbon Ages for MD04-2829CQ and Their Calibrated Correspondence as Well as Their Calendar Age According to the Tuned-to-GISP 2 Age Model and the Difference in Years Between Both Age Models as Plotted in Figure 3b

<table>
<thead>
<tr>
<th>Laboratory Code</th>
<th>Depth (cm)</th>
<th>Material</th>
<th>14C Age (years)</th>
<th>Error Age (±1σ years)</th>
<th>Calendar Age (cal years B.P.)</th>
<th>Error Age (± years)</th>
<th>Age GISP 2 (cal years B.P.)</th>
<th>Difference Age Models (years)</th>
</tr>
</thead>
<tbody>
<tr>
<td>SUERC-8793</td>
<td>312.5</td>
<td>N. pachyderma l.c.</td>
<td>16,732</td>
<td>65</td>
<td>19,450</td>
<td>86</td>
<td>18,498</td>
<td>952</td>
</tr>
<tr>
<td>SUERC-8794</td>
<td>376.5</td>
<td>N. pachyderma l.c.</td>
<td>17,254</td>
<td>69</td>
<td>20,024</td>
<td>107</td>
<td>19,380</td>
<td>644</td>
</tr>
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<td>SUERC-8795</td>
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<td>N. pachyderma l.c.</td>
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<td>70</td>
<td>20,174</td>
<td>91</td>
<td>19,595</td>
<td>579</td>
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<tr>
<td>SUERC-8797</td>
<td>422.5</td>
<td>N. pachyderma l.c.</td>
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<td>73</td>
<td>20,470</td>
<td>94</td>
<td>20,024</td>
<td>446</td>
</tr>
<tr>
<td>SUERC-8798</td>
<td>438.5</td>
<td>N. pachyderma l.c.</td>
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<td>76</td>
<td>20,806</td>
<td>128</td>
<td>20,242</td>
<td>564</td>
</tr>
<tr>
<td>SUERC-8799</td>
<td>457.5</td>
<td>N. pachyderma l.c.</td>
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<td>78</td>
<td>21,127</td>
<td>141</td>
<td>20,373</td>
<td>754</td>
</tr>
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<td>SUERC-8802</td>
<td>495.5</td>
<td>N. pachyderma l.c.</td>
<td>18,312</td>
<td>80</td>
<td>21,239</td>
<td>146</td>
<td>20,649</td>
<td>590</td>
</tr>
<tr>
<td>SUERC-8803</td>
<td>510.5</td>
<td>N. pachyderma l.c.</td>
<td>18,569</td>
<td>83</td>
<td>21,626</td>
<td>165</td>
<td>20,758</td>
<td>868</td>
</tr>
<tr>
<td>SUERC-8804</td>
<td>534.5</td>
<td>N. pachyderma l.c.</td>
<td>18,670</td>
<td>82</td>
<td>21,802</td>
<td>162</td>
<td>21,159</td>
<td>643</td>
</tr>
<tr>
<td>SUERC-8805</td>
<td>544.5</td>
<td>N. pachyderma l.c.</td>
<td>18,804</td>
<td>84</td>
<td>22,012</td>
<td>132</td>
<td>21,436</td>
<td>576</td>
</tr>
<tr>
<td>SUERC-8807</td>
<td>560.5</td>
<td>N. pachyderma l.c.</td>
<td>19,597</td>
<td>92</td>
<td>22,793</td>
<td>130</td>
<td>21,879</td>
<td>914</td>
</tr>
<tr>
<td>SUERC-8808</td>
<td>592.5</td>
<td>N. pachyderma l.c.</td>
<td>20,328</td>
<td>101</td>
<td>23,813</td>
<td>143</td>
<td>22,764</td>
<td>1049</td>
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<td>SUERC-8809</td>
<td>618.5</td>
<td>N. pachyderma l.c.</td>
<td>20,512</td>
<td>104</td>
<td>24,021</td>
<td>132</td>
<td>23,252</td>
<td>769</td>
</tr>
<tr>
<td>SUERC-8812</td>
<td>636.5</td>
<td>N. pachyderma l.c.</td>
<td>20,696</td>
<td>106</td>
<td>24,203</td>
<td>121</td>
<td>23,539</td>
<td>664</td>
</tr>
<tr>
<td>SUERC-8813</td>
<td>648.5</td>
<td>N. pachyderma l.c.</td>
<td>21,383</td>
<td>117</td>
<td>25,107</td>
<td>178</td>
<td>23,911</td>
<td>1196</td>
</tr>
<tr>
<td>SUERC-8814</td>
<td>664.5</td>
<td>N. pachyderma l.c.</td>
<td>21,447</td>
<td>117</td>
<td>25,190</td>
<td>171</td>
<td>24,408</td>
<td>782</td>
</tr>
<tr>
<td>SUERC-8815</td>
<td>720.5</td>
<td>N. pachyderma l.c.</td>
<td>23,505</td>
<td>151</td>
<td>27,726</td>
<td>220</td>
<td>26,441</td>
<td>1285</td>
</tr>
<tr>
<td>SUERC-8816</td>
<td>762.5</td>
<td>N. pachyderma l.c.</td>
<td>23,779</td>
<td>174</td>
<td>29,160</td>
<td>234</td>
<td>28,112</td>
<td>1048</td>
</tr>
<tr>
<td>SUERC-8817</td>
<td>784.5</td>
<td>G. bulloides</td>
<td>25,711</td>
<td>197</td>
<td>30,465</td>
<td>269</td>
<td>29,792</td>
<td>1673</td>
</tr>
<tr>
<td>SUERC-10904</td>
<td>807.5</td>
<td>N. pachyderma l.c.</td>
<td>26,480</td>
<td>146</td>
<td>31,318</td>
<td>210</td>
<td>30,017</td>
<td>1301</td>
</tr>
<tr>
<td>SUERC-10899</td>
<td>816.5</td>
<td>N. pachyderma l.c.</td>
<td>26,963</td>
<td>154</td>
<td>31,839</td>
<td>224</td>
<td>30,679</td>
<td>1160</td>
</tr>
<tr>
<td>SUERC-10900</td>
<td>880.5</td>
<td>N. pachyderma l.c.</td>
<td>30,073</td>
<td>223</td>
<td>35,092</td>
<td>254</td>
<td>34,146</td>
<td>946</td>
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<td>SUERC-10901</td>
<td>916.5</td>
<td>N. pachyderma l.c.</td>
<td>32,900</td>
<td>306</td>
<td>37,905</td>
<td>344</td>
<td>36,755</td>
<td>1150</td>
</tr>
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<td>SUERC-10902</td>
<td>950.5</td>
<td>N. pachyderma l.c.</td>
<td>35,136</td>
<td>399</td>
<td>40,092</td>
<td>421</td>
<td>39,020</td>
<td>1072</td>
</tr>
</tbody>
</table>

*Converted using the online calibration program of Fairbanks et al. [2005], January 2007 version, and assuming a constant 400 year marine reservoir correction.

picked every 2 cm in the interval 780–830 cm. Prior to analyses samples were submersed in hydrogen peroxide (3%) for 30 min and briefly ultrasonicated in ethanol. Excess liquid and residues were quickly removed and the tests dried at low temperatures (40°C). T. quinqueloba samples were not cleaned or crushed prior to analysis due to their fragility and small size. All samples were analyzed on a ThermoFinnigan MAT 252 mass spectrometer coupled to a fully automated CARBO Kiel carbonate preparation device at the Stable Isotope Laboratory of Cardiff University (UK). Isotope results are reported in standard delta notation relative to Vienna Pee Dee Belemnite (VPDB), using NBS-19 for calibration. The external reproducibility of the δ18O determination is ±0.06 ‰.

[13] Possible contamination of isotopic results by clay-size dolomitic carbonate fragments attached to shell walls [Hodell and Curtis, 2008] was reduced to a minimum since all N. pachyderma sin. samples from cream-colored carbonate bearing IRD horizons (mainly HE 2 and HE 4) were crushed before cleaning, as well as G. bulloides samples corresponding to HE 4.

[14] Extra analyses on specimens from intermediate horizons within intervals with anomalous isotopic values (e.g., spikes in the G. bulloides δ18O record) were performed when possible, yielding a final 1 cm resolution on those specific intervals. All paleodata will be publicly available through the World Data Centre for Marine Environmental Sciences (http://www.wdc-mare.org; http://www.pangaea.de).

4. Age Model

[15] The chronostratigraphic framework of core MD04-2829CQ is based on the calibration to calendar years (cal years B.P.) of 24 AMS 14C dates determined from monospecific samples containing more >1000 specimens (equivalent >10 mg of carbonate) of N. pachyderma sin. or G. bulloides (Table 1). Samples for dating were selected from sections where abundance of either taxon maximized in order to facilitate comparison of our results to those from the more recent NGRIP ice core δ18O record [Grootes and Stuiver, 1995] (Figure 2). The resulting age model was then fine tuned and extended beyond current AMS 14C coverage by correlating relative abundance of N. pachyderma sin. to the GISP 2 Greenland ice core δ18Orec record [Grootes and Stuiver, 1997] (Figures 2 and 3a and Table 2.). The GISP 2 record has estimated age uncertainties of about ±2% for the age interval studied here [Meese et al., 1997], and shows a good long-term agreement with the more recent NGRIP ice core [Svensson et al., 2008]. Therefore it was selected for correlation instead of the NGRIP in order to facilitate comparison of our results to those from cores MD01-2461 and DAPC2, also tuned to GISP 2 [Knutz et al., 2007; Peck et al., 2006, 2007a, 2007b, 2008]. The name convention of the INTIMATE group for the GSs and Gls is used [Lowe et al., 2007].

[16] Tuning N. pachyderma sin. abundance maxima to Greenland stadials is a common procedure in North Atlantic records [e.g., Bond et al., 1993; Bond and Lotti, 1995; Knutz et al., 2007; Peck et al., 2006] and is based on the obser-
Table 2. Depths and Corresponding Calendar Ages of the Tie Points Used for the Fine Tuning of the Radiocarbon-Based Age Model to the GISP 2 Oxygen Isotope Record

<table>
<thead>
<tr>
<th>Depth of Pointers (cm)</th>
<th>Age GISP 2 (cal years B.P.)</th>
<th>Stratigraphic Position</th>
<th>Sedimentation Rates (cm kyr⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>300</td>
<td>18,334 GS 1, top of record</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>369</td>
<td>19,280 GS 1</td>
<td></td>
<td>72.90</td>
</tr>
<tr>
<td>434</td>
<td>20,215 GS 1</td>
<td></td>
<td>69.87</td>
</tr>
<tr>
<td>522</td>
<td>20,851 GS 1</td>
<td></td>
<td>137.62</td>
</tr>
<tr>
<td>600</td>
<td>23,017 base GI 2</td>
<td></td>
<td>36.13</td>
</tr>
<tr>
<td>633</td>
<td>23,475 base GI 2</td>
<td></td>
<td>71.69</td>
</tr>
<tr>
<td>687</td>
<td>25,151 GS 3</td>
<td></td>
<td>32.20</td>
</tr>
<tr>
<td>759</td>
<td>28,053 base GI 3</td>
<td></td>
<td>24.87</td>
</tr>
<tr>
<td>794</td>
<td>29,132 base GI 4</td>
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<td>32.34</td>
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<tr>
<td>815</td>
<td>30,681 base “DO 4.1”</td>
<td></td>
<td>13.59</td>
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<tr>
<td>826</td>
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<td>864</td>
<td>33,672 base GI 6</td>
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<td>879</td>
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<tr>
<td>936</td>
<td>38,467 base GI 8</td>
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<td>12.25</td>
</tr>
<tr>
<td>960</td>
<td>39,492 HE4</td>
<td></td>
<td>23.53</td>
</tr>
<tr>
<td>1007</td>
<td>41,153 top GI 10</td>
<td></td>
<td>28.24</td>
</tr>
</tbody>
</table>

*Explanation and references in text and in Figure 3. The approximate stratigraphic position of the selected intervals (Figure 3a) and the partial sedimentation rates (Figure 2) are also included.

viation that this taxon is dominant in present-day polar and subpolar assemblages [Be, 1977; Ruddiman and McIntyre, 1981]. Hence it is considered a proxy for both arctic water masses with seasonal ice cover [Johannesen et al., 1994] and polar water masses with perennial sea ice cover [Elliot et al., 1998; Kuijpers et al., 1998; Pflaumann et al., 1996].

The resulting chronostratigraphic framework (Figure 2) shows that our records span an interval of about 23 kyr, extending between mid MIS 3 (~41.1 ka B.P.) and mid MIS 2 (~19.3 ka B.P.). The age model has independently been constrained by the identification of the Laschamp excursion (~41 ka B.P.) [Lund et al., 2005] in the magnetic record (E. Moreno, personal communication, 2010). Mean sedimentation rates for the studied interval are approximately 31 cm kyr⁻¹, yielding a mean time step of ~65 years for each 2 cm interval, with maximum rates of up to 137 cm kyr⁻¹ recorded during MIS 2. Sedimentation rates increase from ~26 ka B.P. toward the Last Glacial Maximum (LGM), in a similar fashion to those recorded in nearby cores MD95-2006, DAPC2 and ENAM32 [Knutz et al., 2001, 2007; Lassen et al., 1999] and sites in the southern Norwegian Sea [Nielsen et al., 2007; Rasmussen et al., 1996b], probably reflecting a combination of changing bottom current intensity and terrigenous sediment supply to the core site.

Significant temporal discrepancies between the initial 14C-calibrated and the calendar year timescale derived from tuning to the GISP 2 age model (Figure 2) suggest that local marine reservoir ages (ΔR) have shifted within a range of ~430 to ~1700 years, far larger than the modern global average 400 year reservoir correction (Figure 3b). Mean implied ΔR values during MIS 2 are typically ~700 years, while in MIS 3 the divergence increases to ~1000 years. Maximum ΔR values are found close to H events. High reservoir ages during colder periods are systematically registered in records recovered off the British Isles [Knutz et al., 2007; Lekens et al., 2006; Peck et al., 2006] and in other regions of the glacial North Atlantic, due to a combination of sea-ice and meltwater induced reduction of air-sea gas exchange, ventilation and isotope equilibration [Austin et al., 1995; Bard et al., 1994; Voelker et al., 1998, Sarnthein et al., 2007; Waelbroeck et al., 2001]. Increased contribution of 14C-depleted Antarctic waters reaching the high latitude North Atlantic [Cao et al., 2007] possibly have added to the elevated reservoir ages. The effect of these differences apparently decreases toward lower latitudes, deglacial warm periods and the Holocene [Cao et al., 2007; Waelbroeck et al., 2001].

5. Results

5.1. Abundance Distribution of Planktonic Foraminifera

Changes in the relative abundances of planktonic foraminifera closely reproduce the variability seen in the GISP 2 δ18O ice record (Figure 4). Large fluctuations in the assemblage composition characterize the period between 41 and 27 ka B.P., in accordance with the high-amplitude GS-GI cycles 4 to 11. N. pachyderma sin. dominates during HE and GSs, while transient increases of temperate/subtropical taxa (e.g., Globorotalia scitula) mark the warmer GSs, dominated by G. bulloides and T. quinqueloba. Abundance of N. pachyderma (dental coiling; abundances not shown here) also increases during GSs but never reaches values above 5%.

Figure 2. Age models of core MD04-2829CQ. (a) Solid triangles, raw radiocarbon AMS dates; (b) thin black line, AMS 14C dates calibration to calendar ages; and (c) thick black line, tuned-to-GISP2 age model. The grey area evidences the divergence between the AMS 14C calibrated and the tuned-to-GISP2 age models. The sedimentation rates according to the tuned-to-GISP2 age model are also plotted. Positions of HE 2–4 are included for reference.
N. pachyderma sin. is the dominant taxon between 27 and 18 ka B.P., showing small abundance shifts during the LGM interval coincident with centennial-scale climate variations observed in Greenland. Major exceptions are two brief abrupt increases of the cold/temperate species Turborotalita quinqueloba and G. bulloides coincident with GI 2.

5.2. Faunal-Based Sea Surface Temperatures

SST from ANN and MAT techniques generated very similar records and values, with the occasional exception, for example peak warmth during GI 3 and 4, where ANN estimates suggest higher temperatures up to 3°C higher than MAT (Figure 4). A few levels from both reconstructions, however, are affected by no-analog situations, as indicated by MAT similarity values below 0.7 and ANN standard deviations larger than 1σ [Kucera et al., 2005] (Figure 4e). Most of these intervals also show the highest degree of disagreement between the two techniques and so should be considered with some caution. Nevertheless, it is also the case that these no-analog situations typically occur within warm intervals, when assemblages are dominated by G. bulloides and T. quinqueloba with less than 5% of N. pachyderma dex., an unusual assemblage with no present-day analog in the MARGO Atlantic database [M. Kucera, personal communication, 2008]. However, low N. pachyderma dex. abundance, even during GIs, is reported in other cores recovered from the Rockall Trough and surrounding area [e.g., Knutz et al., 2007; Kuipers et al., 1998; van Kreveld et al., 2000] suggesting a regional ecological feature in response to oceanographic conditions. Therefore, we consider that these SST estimates provide a generally faithful trend reconstruction, supported by the similarity of the curves to that of the GISP 2 δ¹⁸O ice record.

Figure 3. (a) Correlation of the relative abundance of N. pachyderma sin. recorded in core MD04-2829CQ with the GISP 2 oxygen isotope record [Grootes and Stuiver, 1997]. Dotted arrows connect tie points used for the correlation, while black triangles indicate the horizons dated with AMS ¹⁴C. (b) Difference in years between the calibrated AMS ¹⁴C ages and the calendar ages for the same horizons as given by the tuned-to-GISP2 age model. The horizontal dashed line marks the average modern mean reservoir age. The position of HE 2 to HE 4 is plotted for reference.
cores south of the Iceland/Faroe Ridge for the same time interval [Weinelt et al., 2003].

5.3. Lithic Composition and Abundance

The flux of total IRD to the core site is continuous throughout the 41–18 ka B.P. interval (Figure 5); however warmer GIs are characterized by very low abundances of mineral grains while increased fluxes occur during HE, GSs and MIS 2. The sequence of centennial-scale IRD events superimposed on a background of increasing IRD fluxes from ~28 ka B.P. toward HE 2 is a major aspect of the record. This feature is also observed in the IRD record of cores SU90-09 and MD95-2002 [Auffret et al., 2002; Grousset et al., 2000, 2001] in the southeastern margin of the main IRD belt, and is similar to the increases shown in cores DAPC2 [Knutz et al., 2002, 2007] and MD01-2461 [Peck et al., 2006] before HE 1. Also, spectral analysis of the record (multitaper method [Pardo-Igúzquiza...
et al., 1994] with seven data tapers and spectral confidence levels located using the robust AR(1) modeling of median-smoothed spectra, not included here) shows significant centennial-scale oscillations during MIS 2, which follow a rough 500 year cyclicity, while between 30 and 41 kyr the frequency of IRD events decreases to around 1100 years. The dominant components of the mineral grain assemblage, transparent quartz and haematite-stained grains, are likely sourced in a range of locations around the North Atlantic that are impossible to isolate. The remaining grains comprise a wide range of lithologies. HE 4 and 2 stand out as peaks of finely crystalline, cream-colored carbonate (>750 grains cm\(^{-2}\) kyr\(^{-1}\) at HE 4; Figure 5d), classically diagnostic of HEs, identified as sourced from dolomitic limestones outcropping in the Hudson Bay Province of the LIS [Andrews and Tedesco, 1992; Bond et al., 1992]. Sedimentary lithics originating from the BIIS and igneous rock fragments typically increase during cold periods. Basalt grains, petrographically similar to those described as sourced in the Tertiary Volcanic Provinces outcropping in western Scotland [Knutz et al., 2007], show an abundance distribution that parallels that of the BIIS component (Figures 5b and 5c). Other volcanic grains, usually associated with the Icelandic Ice Sheet (IIS), do not show a distinct increase toward HE 2 as the other groups (Figure 5b). Some peaks could be related to tephra layers identified in other cores of this area [Rasmussen and Thomsen, 2004].

**Figure 5.** Correlation of the major lithic records of core MD04-2829CQ to GISP 2. (a) GISP 2 ice core record, (b and c) fluxes of the main IRD groups, and (d) total IRD and LIS-sourced cream-colored carbonate fluxes. Grey vertical bars mark HE (dark grey) and some of the Greenland stadial (light grey) horizons.

5.4. **Hornblende Ages**

Hornblende grains are present throughout most of the record, although their abundance is higher at certain intervals such as HE 4 and between 26 and 23 ka B.P. (including HE 2), and 22 and 20 ka B.P. (Figure 6). A total of 419 individual grains from 184 samples covering the interval...
1600–1900 Myr
(Paleoproterozoic)
Church Prox (N America)
W Greenland
Lavelle Complex (NW Scotland)
200–600 Myr
(Paleozoic)
Appalachian (N America)
Caledonides & Greenland
Ellesmere Island
Scotsman Area

MAIN GROUPS
PRESENT
(Hemming et al., 1998)

~18 to ~40 ka B.P. produced good quality Ar\(^{40}/\text{Ar}^{39}\) ages, of which 10 were discarded as grains were suspected to be pyroxenes.

[27] By dividing the sediment record into several stratigraphic intervals selected to separate HE from periods between HE (Figure 7a, <23.5 ka B.P.; Figure 7b, HE 2; Figure 7c, 24.3–29 ka B.P.; Figure 7d, cream-colored carbonate peak time span within HE 2; Figure 7e, 31–38.6 ka B.P.; Figure 7f, HE 3; Figure 7g, >40 ka B.P.; Figure 7h, HE 4) and grouping the hornblende grains accordingly, we constructed histogram graphs in order to study the population composition in each period (Figure 7). We selected the histogram age bins based on known major geological provinces around the North Atlantic [Hemming et al., 1998, 2002; Hemming and Hajdas, 2003].

[28] The results highlight the importance of the Paleo- proterozoic (1600–1900 Ma) and Paleozoic (200–600 Ma) hornblende grain populations, while the Mesoproterozoic (1400–1650 Ma) and the Early Paleoproterozoic (1900–2400 Ma) are represented by far fewer grains. The Paleoproterozoic population is dominant during most of the record, although the quantity of grains within all groups tends to vary simultaneously and the Paleozoic population is also present in high numbers during most periods. These results contrast with sediment core V23-14 from the western North Atlantic [Hemming and Hajdas, 2003], in where Paleopro-
terozoic populations dominate within each HE and in the intervals between them, with far less numbers of Paleozoic (Appalachian) and Mesoproterozoic (Grenville) grains. Moreover, Paleoproterozoic grains in our core are not restricted to the intervals synchronous with the deposition of cream-colored carbonates (Figures 7d and 7h), which opposes their common apparent source in the Hudson Strait area in the Canadian Shield [Andrews and Tedesco, 1992; Gwiazda et al., 1996; Hemming, 2004].

5.5. Oxygen Isotopes

Structural similarity with the GS-GI cyclicity is not visible within the G. bulloides and N. pachyderma sin. 

![Histogram plots of hornblende populations in different intervals of core MD04-2829CQ](image)

**Figure 7.** (a-h) Histogram plots of hornblende populations in different intervals of core MD04-2829CQ (see text and Figure 6 for interval significance). In each histogram, n is the number of hornblende grains considered for that specific age interval.

δ¹⁸O records in our core (Figures 8 and 9b), as also noted in other records from the region [e.g., Lassen et al., 1999; Lekens et al., 2006; Rasmussen et al., 1996b]. Both curves show a trend of increasing δ¹⁸O between HE’s, attaining maximum values, indicative of coldest and/or most saline conditions, immediately before the HEs. Oxygen isotope values of G. bulloides remain on average 0.5% lighter than those of N. pachyderma sin., while both records converge immediately before HE 3 and 2, and during the LGM (Figure 9b).

Contrary to the δ¹⁸O record of G. bulloides the N. pachyderma sin. δ¹⁸O record shows only small amplitude changes (~0.2%) (Figures 8 and 9b), coincident with
the faunal and SST records. The rather higher amplitude changes along the *G. bulloides* δ¹⁸O typically coincide with transient cold events and increased fluxes of IRD (Figure 9).

[31] Heinrich Events stand out in the *N. pachyderma* sin. record as isolated negative δ¹⁸O anomalies of ~0.5 and ~1‰ that last 200 to 300 years at HE 2 and HE 4 respectively, and a longer ~0.75‰ reduction in δ¹⁸O during HE 3 that leads the IRD event by almost 1000 years (Figure 9). While, the HEs IRD peaks coincide with only subtle negative δ¹⁸O excursions in *G. bulloides* δ¹⁸O that are superimposed on a longer-term trend toward lower δ¹⁸O.

[32] The short *T. quinqueloba* record (Figure 9b) seems to follow the GS-GI cyclicity, showing light peaks simultaneously to warm periods in the GISP 2 δ¹⁸O ice record. Similarly to the other two planktonic isotopic records, a large (~1‰) depletion characterizes the interval before the onset of HE 3.

6. Discussion

6.1. Ice Sheet Dynamics

6.1.1. Iceberg Provenance

[33] The composition of the most abundant group of IRD in core MD04-2829CQ, besides the ubiquitous quartz and haematite-stained grains (Figure 5c), has been previously associated to various outcrops in the British Isles [Peck et al., 2007a; Scourse et al., 2000, 2009]. The presence of soft chalk grains of Cretaceous age (C. Koch, personal communication, 2008) and of basalts, identified as originated in the Tertiary Volcanic Province of western Scotland, in the MD04-2829CQ record support this origin. Additionally,
Figure 9
the occurrence of occasional fine gravel sized (4–8 mm) IRD during periods of highest IRD flux points to the existence of iceberg sources within a distance of 10–100 km [Knies et al., 2001; Smith and Andrews, 2000], consistent with the assumed expansion of BIIS margins during peak glacial conditions [Knutz et al., 2001; Sejrup et al., 2005]. This would also account for the sustained IRD fluxes and for the near invariable composition of IRD assemblages registered through the entire record. Alternative nearby rock fragment sources for a BIIS-like assemblage, such as the Late Paleozoic to Tertiary sediments that form the basement of the Rockall Basin within the Rockall Trough, are not considered since top layers consist mainly of recent detritic unconsolidated material [Morewood et al., 2004].

[34] We also require a suitable source, or combination of sources, to explain the almost permanent flux of both Paleoproterozoic and Paleozoic hornblende grains to the core site (Figure 6). The Churchill Province units from the Hudson Strait region and the southern Labrador area are a likely origin for both populations [Gwiadza et al., 1996; Hemming et al., 1998, 2002; Hemming, 2004]. However, LIS-sourced, cream-colored carbonates are only present at transient (200–300 years) episodes within the GSs that contain HE 2 and HE 4 (Figures 5d and 9f). These intervals correlate to geochemical and magnetic evidence from other northeast Atlantic cores [Groussset et al., 2000; Peck et al., 2007a; Peters et al., 2008], which show that the influence of the LIS signal in the northeast Atlantic is exclusive to those horizons. Thus, while it is possible that the Paleoproterozoic and Paleozoic hornblende population at those events may have been mostly or entirely sourced in the LIS, lack of further evidence of a LIS imprint in our core during non-HS intervals suggests that other sources of Paleoproterozoic and Paleozoic hornblende grains were also contributing. Those other potential areas include Greenland and the Lewisian Complex outcropping in the Outer Hebrides (northwest Scotland) for the Paleoproterozoic, and Scandinavia and the British Isles for the Paleozoic. The Lewisian Complex seems the most probable source of Paleoproterozoic grains due to its proximity and to the documented flow path of ice streams of the westernmost BIS margin [Sejrup et al., 2005] (Figure 1). It is also notable that the interval of highest concentration of Paleoproterozoic hornblende grains (Figure 6) is simultaneous to the steady increase in the IRD fluxes and the centennial-scale events in the period 27–20 ka B.P., documented as the interval of BIIS advance to the shelf break [Wilson and Austin, 2002], supporting a West Scotland origin for most Paleoproterozoic hornblende grains. A northeastern Atlantic source such as Scandinavia and the British Isles is also more likely for the Paleozoic population of our core, since grains of this age range are more common in the Nordic Seas [Hemming et al., 2002]. This likely explains the differences in hornblende distributions from our core (Figure 7) to those of core VM23-14 in the Central North Atlantic [Hemming and Hajdas, 2003] in where Paleozoic populations are not as important.

6.1.2. BIIS Evolution

[35] The frequency and scale of iceberg discharge events recorded in core MD04-2829CQ are likely connected to BIIS ice volume and its growth throughout the last glacial period. The smaller, millennial-scale discharges of roughly equal magnitude during MIS 3–related cold events contrast with the increasingly larger, centennial-scale IRD peaks dominating MIS 2 (Figures 5d and 9f), and indicate substantial changes in the extent of nearby BIIS and the development of fringing marine ice shelves, as suggested by Scourse et al. [2009]. Land-based evidence from Scotland shows an interstadial period between 38.1 and 32 ka B.P. (the Alesund/Denkamp/Sadnes interstadial), widely recognized in Western Europe and equivalent in age to GIs 8 to 5 [Whittington and Hall, 2002]. This warmer period was followed in land by major climate deterioration and ice advance to the west and south after ∼32 ka B.P. [Brown et al., 2007; Finlayson et al., 2010]. In the marine record, a combination of enhanced NWEIS build-up rate and general sea level lowering due to decreasing insolation from ∼28 ka B.P. seems the probable cause of this shift in the frequency of iceberg calving from that point. Both circumstances apparently triggered continuous centennial-scale readjustments of the extended marine-based NWEIS margins during this younger period, also evidenced in nearby sediment records off northeast Europe and Iceland [Fronval et al., 1995; Knutz et al., 2007; Lassen et al., 1999; Lekens et al., 2006; Peck et al., 2006; Rasmussen et al., 1996a; Scourse et al., 2009; van Kreveld et al., 2000], which are all consistent with short-term cooling episodes in Greenland. This high-frequency response is also observed prior to HE 1 [Knutz et al., 2007; Peck et al., 2006]. Knutz et al. [2007] suggest that increasing 65°N summer insolation following the LGM was a potential triggering mechanism for the centennial IRD cycles observed in core DAPC2 showing early BIIS disintegration prior to HE 1. Our data from core MD04-2829CQ indicate that both increasing and decreasing 65°N summer insolation are associated with similar readjustments in the BIIS margins and point to a sustained metastability of the BIIS throughout MIS 2.

[36] The absence of large peaks in the total IRD flux for HE 3 and HE 4 in core MD04-2829CQ, reflecting the relatively small size of the BIIS during MIS 3, contrasts with those of records inside the main IRD belt between 40 and 55°N [after Ruddiman, 1977; Hemming, 2004], in which HEs represent the most important discharge episodes [Bond et al., 1993; Bond and Lotti, 1995]. LIS-sourced IRD accumulation during HE 2 and HE 4 decreases toward the northeast North Atlantic and outside this belt, and is also supported by smaller fluxes of LIS-sourced cream-colored carbonates at Rosemary Bank in contrast to those from core MD01-2461 in the Porcupine Bight, offshore southwest

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Figure 9. Compilation of some of the records obtained for core MD04-2829CQ: (b) surface (G. bulloides and T. quinquelaoba) and subsurface (N. pachyderma sin.) oxygen isotopes (with extra N. pachyderma sin. and G. bulloides samples in open rhombi and grey squares, respectively; explanation in the text), (d) N. pachyderma sin. relative abundances, and (e) summer SST estimates at 10 m depth as in Figure 4e, and (f) total IRD and LIS-sourced cream-colored carbonate fluxes. (a) The GISP 2 oxygen isotope record and (c) the June 65°N mean insolation are plotted for reference. Grey vertical bars mark the position of HE (dark grey) and some of the Greenland stadial (light grey) horizons.
Ireland [Peck et al., 2006], and of ODP Site 609 and core VE23-81 further offshore on the west flank of the Rockall Trough [Bond and Lotti, 1995]. These decreasing flux gradients indicate that the LIS only played a fairly minor role in IRD accumulation at Rosemary Bank that was limited to short intervals within HE 4 and HE 2. Recent data on mineral magnetic measurements from core MD95-2006 recovered from the nearby Barra Fan [Peters et al., 2008] support this interpretation.

6.2. Ice Sheet–Upper Ocean Connection

[37] Today N. pachyderma sin. thrives in the Nordic Sea and Arctic Ocean at the base of the summer seasonal surface mixed layer at ~50 m [Bauch et al., 1997; Jonkers et al., 2010]. In the subpolar North Atlantic, G. bulloides calcifies within 0–60 m water depth, blooming in the summer months [Cortijó et al., 1997; Ganssen and Kroon, 2000]. Field and modeling evidence from present-day and LGM assemblages in polar and subpolar regions [Fraile et al., 2009a, 2009b; Jonkers et al., 2010; Köhfeld et al., 1996] suggest that glacial blooms of both G. bulloides and of N. pachyderma sin. occurred during the summer months. However, given the habitat preferences of the two species we can interpret the G. bulloides δ18O record as being a summer SST signal and N. pachyderma sin. δ18O as representing mean annual subsurface conditions [Peck et al., 2008]. The ~0.5‰ offset between both δ18O records at this site prior to the LGM (Figure 9b) points to a strong gradient between surface and subsurface waters during the summer months, with denser (cooler and/or saltier) waters below lighter (warmer/less saline) surface waters. At the LGM the two δ18O records converge suggesting that both species are living in the same water mass, suggestive of year-round mixing of the upper waters/no seasonal thermocline.

[38] A comparison of both isotopic records with the equivalent from core MD01-2461 in the Porcupine Seabight (SW Ireland) [Peck et al., 2006] (Figure 8) reveals clear similarities in upper water column structure at the two sites. The surface isotopic signals seem to represent roughly equivalent water masses at both locations. Since G. bulloides from the LGM seem to calcify in summer in both areas [Fraile et al., 2009b], the small occasional difference between both records can be interpreted as a minor latitudinal temperature gradient and/or meltwater lenses.

[39] Comparable N. pachyderma sin. δ18O records from the two sites suggest similar subsurface water mass properties at both locations up until ~27 ka B.P. After this time, the δ18O N. pachyderma sin. record at the more southerly site becomes up to 0.5‰ lighter than in Rosemary Bank, evidencing clear differences between both records.

[40] A likely explanation for the different behavior of both records from ~27 ka B.P. relies in the migration of the Polar Front. The comparison of N. pachyderma sin. relative abundance records (Figure 8) during MIS 3 shows near synchronous variations between both sites until ~27 ka B.P., allowing us to infer an unstable Polar Front migrating rapidly along the western BIIS margin, across both sites, simultaneously to GS-GI cyclicity. Additional evidence of the migration of the Polar Front as far south as 40°N during HE and some of the GS is observed in records from the western Iberian Margin [Eynaud et al., 2009; Salgueiro et al., 2010; Voelker et al., 2009]. In contrast, at the onset of MIS 2 the persistent dominance of N. pachyderma sin. in the faunal assemblage and heavier δ18O values at Rosemary Bank are consistent with a scenario of BIIS build-up and a likely almost stable position of the Polar Front south of core MD04-2829CQ. Periodic migration of the Polar Front south of MD01-2461 is reflected in our records by the convergence of both isotopic and faunal signals, at 25.7 ka B.P., 25 ka B.P. and HE 2. After HE 2 the Polar Front remains between the two sites but appears to periodically retreat north of core MD04-2829CQ when N. pachyderma sin. relative abundance from this site decreases to converge with MD01-2461, although this is not expressed in the isotopic records. BIIS advance at ~21.5 ka B.P. (also observed by Knutz et al. [2007] and Peck et al. [2006]) is associated with a final advance of the Polar Front toward MD01-2461, while three further Polar Front retreats northward of Rosemary Bank are noted at 20.1, 19.9 and 19 ka B.P.

[41] Cold events such as the ones at ~25.7 and ~21.5 ka B.P. produced meltwater plumes both north and south of the BIIS, as reflected in the light δ18O peaks in G. bulloides recorded at both sites. At HE 2, the amount of melt waters and icebergs seems to be much larger at the southern front of the BIIS, as shown by the large isotopic spike and IRD flux recorded in core MD01-2461 (Figure 8) [Peck et al., 2007a] and in core MD95-2002 further southeast along the continental margin [Grousset et al., 2000; Auffret et al., 2002].

[42] Contrary to the N. pachyderma sin. record, the small response of G. bulloides to the events recorded during HE 2 and MIS 2 in core MD04-2829CQ likely show a partially biased record affected by the harsh conditions of average years, with low to very low abundances (Figure 4) and/or summer blooming restricted to the warmer years.

[43] Isotopic data from core MD04-2829CQ suggest that MIS 3 GS conditions were characterized by cooling synchronously to iceberg calving from the BIIS (Figure 9b). The frequency and scale of iceberg discharge events and hence the input of low-salinity waters to the surface layers water are greatly enhanced after ~31 ka B.P. as may be inferred from the increased magnitude of the light G. bulloides δ18O excursions and of simultaneous maxima in IRD flux during the centennial-scale BIIS-sourced events prior to HE 2 and during the LGM. The absence or significantly smaller scale of the related decreases in the N. pachyderma sin. δ18O record during most of these cold intervals is remarkable since the associated IRD flux peaks are of similar or even greater magnitude to those recorded during HE 3 and 4. This may reflect a stronger water column stratification resulting in the migration of N. pachyderma sin. to deeper depths below the halocline [Köhfeld et al., 1996]. Alternatively, Elliot et al. [1998] attributed an analogous N. pachyderma sin. δ18O signal to a combination of decreased meltwater flux and its origin in coastal ice sheets, therefore showing a less fractionated isotopic signature. However, since the BIIS is in all cases the likely source of meltwater and the ice sheet size was far larger after ~28 ka B.P. than during MIS 3, the degree of distance-based isotopic fractionation between ice accumulation areas and the ice sheet margins should have increased rather than decreased as the BIIS increased its ice volume, and for
the interval before ∼28 ka B.P., the relatively low fractionation may have reduced the meltwater signal recorded in the *N. pachyderma sin. δ18O*. The magnitude of the *G. bulloides δ13C* light anomalies and IRD fluxes between 28 and 18 ka B.P. in MD04-2829CQ suggest that a meltwater lens may have been a persistent feature at that time/site, although, because of the low abundances of *G. bulloides*, this interpretation should be treated with some caution. Overall, it seems probable that the freshening associated with the BIIS IRD surging only influenced the surface waters and did not fully affect deeper layers. The scale of some of these calving events (such as those at ∼25.7 ka B.P. and ∼21.5 ka B.P.) must have involved much of the western margin of the BIIS since simultaneous meltwater plumes were also recorded in core MD01-2461 (Figure 8). Larger fluxes of IRD at MD01-2461 during these two events [Peck et al., 2007a] compared with our core further north may be indicative of higher instability in the southwestern BIIS margin at those times.

[44] As highlighted in Figure 9 it is evident that in our core HE 4 and 2 are characterized by prominent light peaks in the isotopic record of *N. pachyderma sin.* coincident with the arrival of LIS-sourced material. These excursions are not accompanied by large depletions of the isotopic values of *G. bulloides* as during centennial-scale cooling events; rather, negative values in *G. bulloides δ18O* are less clearly defined, and tend to lag those of *N. pachyderma sin.* It has been suggested that convergence of surface and subsurface δ18O values during HE represents strong mixing of the upper water column down to ∼100–200 m [Simstich et al., 2003], and a recent study suggest that this deepening of the surface mixed layer might have been caused by a brief increase in storminess during GSs containing HEs [Rashid and Boyle, 2007]. However, convergence of our isotopic records is not evident in HE 4 and HE 2 and only HE 3 shows possible evidence of this process although leading the maximum IRD flux by a few hundreds of years.

[45] HE 3 stands out in our record as a distinct event due to the absence of LIS-sourced material and to the particular isotopic signals recorded in this interval. Before this event and for a ∼500 year interval, δ18O values from three different species representing diverse depths in the upper water masses decrease simultaneously (Figure 9b). A prominent negative anomaly is also observed in the *N. pachyderma sin. δ18O* record at MD01-2461 around the same time (Figure 8), with an associated increase in benthic δ18O and decrease in benthic δ13C values and 14C marine reservoir ages [Peck et al., 2006, 2007b]. We note that these anomalies coincide with extremely low IRD fluxes at MD04-2829CQ (Figure 9f), consisting predominantly of mica flakes, and of a faunal assemblage dominated almost entirely by *T. quinquelaoba* in the planktonic foraminiferal fraction (Figure 4). Together, all these indicators seem to show either the presence of an exceptionally warm or less saline water mass apparently unconnected to BIIS calving events, since its IRD composition is unique. This event can be correlated to the so-called “DO 4.1” in Greenland and Antarctic ice core records [EPICA Community Members, 2006] and represents a warming phase in those records and in the Western Mediterranean [Sierro et al., 2009]. Warming is not obvious from our records as *T. quinquelaoba* is characterized as an extremely resistant species that can survive to very low salinity conditions [Simstich et al., 2003], and has been previously interpreted as a marker for the position of the Arctic front [Johannessen et al., 1994; Eynaud et al., 2009]. Moreover, simultaneously to our event Lekens et al. [2006] registered a “low salinity” event focused in the southern North Sea apparently caused by the breakage of ice dams and flooding of freshwater glacial lakes from the North Sea. These authors consider this low δ18O as representing HE 3 in this area, but our records indicate that, if being the same event, it preceded HE 3 itself (or at least BIIS IRD response to HE 3) by 2–3 centuries, in line with the observations of van Kreveld et al. [2000] in the Irminger Sea.

### 7. Conclusions

[46] This study emphasizes the high sensitivity of the BIIS as an important component of the climate system of the Last Glacial in the North Atlantic. The interval between 41 and 18 ka B.P. was characterized in the Northeast Atlantic by a succession of iceberg calving events sourced in the BIIS synchronous with decreases in SSTs that reflect the southern advances of the Polar Front. Those events occurred with millennial frequencies from 41 to 28 ka B.P. when the BIIS was not fully developed, and at centennial timescales from 28 to 18 ka B.P. reflecting the ice sheet building up and simultaneous and rapid development of fringing marine ice shelf margins, resulting in characteristic metastability. The data confirm the strong coupling between BIIS instabilities and surface ocean circulation during MIS 3/2, and demonstrate BIIS ability to modify the North Atlantic Current on its flow toward the Nordic Seas; petrological evidence indicates a superimposed contribution of the LIS during HE 2 and 4.

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