Timing of meltwater pulse 1a and climate responses to meltwater injections

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Received 5 July 2006; revised 18 August 2006; accepted 13 September 2006; published 9 December 2006.

[1] The temporal relationship between meltwater pulse 1a (mwp-1a) and the climate history of the last deglaciation remains a subject of debate. By combining the Greenland Ice Core Project δ18O ice core record on the new Greenland ice core chronology 2005 timescale with the U/Th-dated Barbados coral record, we conclusively derive that mwp-1a did not coincide with the sharp Bølling warming but instead with the abrupt cooling of the Older Dryas. To evaluate whether there is a relationship between meltwater injections, North Atlantic Deep Water (NADW) formation, and climate change, we present a high-resolution record of NADW freshening in the Arctic/Nordic Seas. The combined use of Accelerator Mass Spectrometry (AMS) 14C and U/Th datings in fossil coral studies offers insight into the real time structure of the deglacial sea level rise, and also allows extension of the radiocarbon calibration curve beyond 10 ka (thousands of years before Present, where Present refers to A.D. 1950) [Bard et al., 1990a, 1998, 1996; Fairbanks et al., 2005]. Fossil coral records contain evidence of a dramatic sea level rise in excess of 20 m within the last deglaciation, the so-called meltwater pulse (mwp-1a) [Fairbanks, 1989; Fairbanks et al., 2005; Bard et al., 1990b, 1996; Hanebuth et al., 2000; Siddall et al., 2003], Siddall et al. [2003], have documented that global sea level has risen by 120 m or more since the Last Glacial Maximum. The combined use of Accelerator Mass Spectrometry (AMS) 14C and U/Th datings in fossil coral studies offers insight into the real time structure of the deglacial sea level rise, and also allows extension of the radiocarbon calibration curve beyond 10 ka (thousands of years before Present, where Present refers to A.D. 1950) [Bard et al., 1990a, 1998, 1996; Fairbanks et al., 2005]. Fossil coral records contain evidence of a dramatic sea level rise in excess of 20 m within the last deglaciation, the so-called meltwater pulse (mwp-1a) [Fairbanks, 1989; Fairbanks et al., 2005; Bard et al., 1990b, 1996; Hanebuth et al., 2000; Siddall et al., 2003].

2] Well-dated fossil corals [Fairbanks, 1989, 1990a; Bard et al., 1990a, 1990b, 1996; Fairbanks et al., 2005], and other methods [e.g., Rohling et al., 1998; Hanebuth et al., 2000; Siddall et al., 2003], have documented that global sea level has risen by 120 m or more since the Last Glacial Maximum. The combined use of Accelerator Mass Spectrometry (AMS) 14C and U/Th datings in fossil coral studies offers insight into the real time structure of the deglacial sea level rise, and also allows extension of the radiocarbon calibration curve beyond 10 ka (thousands of years before Present, where Present refers to A.D. 1950) [Bard et al., 1990a, 1998, 1996; Fairbanks et al., 2005]. Fossil coral records contain evidence of a dramatic sea level rise in excess of 20 m within the last deglaciation, the so-called meltwater pulse (mwp-1a) [Fairbanks, 1989; Fairbanks et al., 2005; Bard et al., 1990b, 1996; Hanebuth et al., 2000; Siddall et al., 2003].

A marked step in the last deglaciation is recognized in Greenland ice core records by the sharp Bølling warming (GI-1d) [Björck et al., 1998], dated at around 14.6 ka [Rasmussen et al., 2006]. At around 14 ka, the Bølling warm period was ended abruptly by the Older Dryas cold event (GI-1d) [Björck et al., 1998, 1996; Rasmussen et al., 2006]. On the basis of an inferred age of around 14.6 ka for mwp-1a, this meltwater event has been suggested as a trigger for the Bølling warming [Hanebuth et al., 2000]. Isostatic rebound calculations and ocean-climate modeling results have been used to support the suggestion that mwp-1a was derived from Antarctica and was the direct cause for the Bølling warming [Clark et al., 1996, 2002; Kienast et al., 2003; Weaver et al., 2003]. In contrast, earlier results from the Barbados and Tahiti fossil coral records were used to infer that mwp-1a should instead be associated with the Older Dryas cooling that terminated the Bølling warming [Bard et al., 1996].

[4] To resolve this controversy, we compare the Greenland ice core climate proxy record on the new layer-counted Greenland ice core chronology 2005 (GICC05) timescale [Rasmussen et al., 2006], with the latest U/Th-dated history of deglacial sea level change [Fairbanks et al., 2005; Peltier and Fairbanks, 2006]. To evaluate the implications of the resolved relative timings between these records with respect to climate change mechanisms asso-

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ciated with meltwater injections, we present a new high-resolution record of NADW flow intensity during the last deglaciation, as recorded in core TTR13-AT451G (hereinafter referred to as TTR-451) from Eirik Drift (south of Greenland) (Figure 1). The results are then placed within the context of recent suggestions about the location and depth-distribution of meltwater pulses [Aharon, 2005; Tarasov and Peltier, 2005]. Modern observations of surface freshening and sea ice reductions in the Arctic Ocean and Nordic Seas [Dickson et al., 2002; Lindsay and Zhang, 2005] and reduced formation of NADW [Dickson et al., 2002; Bryden et al., 2005] underline the need for a better understanding of the potential ocean-climate responses to meltwater injections.

2. Revised Ages of mwp-1a and Greenland Climate Events

The fossil reef series indicates a sharp global sea level rise of about 20 m at around 14 ka (Figure 2b): meltwater pulse 1a (mwp-1a) [Fairbanks, 1990]. Gaussian smoothing through the recently improved data series for Barbados [Fairbanks et al., 2005; Peltier and Fairbanks, 2006] illustrates the history of sea level change through time (Figure 2b), and its first time derivative offers insight into rates of sea level change (Figure 2c). The latter clearly identifies mwp-1a between 14.17 and 13.61 ka, with a peak rate of 4.3 cm yr$^{-1}$ sea level rise close to 13.86 ka. The youngest indicator of low pre-mwp-1a sea level is sample RGF 9-8-2, which was originally U/Th-dated at 14.235 ± 0.050 ka (1σ) [Bard et al., 1990a, 1990b] and was recently re-dated at 14.082 ± 0.028 ka (1σ) using higher precision mass spectrometry and new decay constants [Fairbanks et al., 2005]. This single young point has been questioned [Kienast et al., 2003; Weaver et al., 2003] since Sunda shelf data suggest that mwp-1a started considerably earlier, at around 14.6 ka [Hanebuth et al., 2000]. However, the Sunda shelf record relies upon AMS$^{14}$C ages, which are subject to much greater uncertainty than U/Th chronologies, especially when corrections for variable $^{14}$C reservoir ages are considered. Recent improvements in the resolution of the Barbados record have added six new low (pre-mwp-1a) sea level points between 14.60 and 14.08 ka that comprehensively validate the age of sample RGF 9-8-2 (Table 1,
Stable oxygen isotope ratios ($\delta^{18}O$) of ice from Greenland ice cores reflect temperature and air mass variations over the ice sheet. More negative values generally represent colder conditions, and less negative values represent warmer conditions. Recent layer-counting has provided a common timescale for three key ice cores from the Greenland ice sheet back to 14.73 ka [Rasmussen et al., 2006]. The resultant new GICC05 timescale is in good overall agreement with that of the fourth key record from the Greenland ice sheet (GISP2) [Meese et al., 1997]. In the GRIP ice core $\delta^{18}O$ record, a sharp shift near the bottom of the layer-counted series (Figure 2a, black) marks the abrupt onset of the Bølling (GI-1e) [Björck et al., 1998] warm period. In the GICC05 timescale this warming is dated at 14.64 ka, with a maximum error of 186 years [Rasmussen et al., 2006], which is in close agreement with its layer-counted age of 14.65 ± 0.09 ka in Cariaco Basin [Lea et al., 2003]. These tight constraints place the warming some 60 years earlier than the previous, less certain, ss09sea ice core timescale [Johnsen et al., 2001] (Figure 2a, thin grey).

Figure 2. (a) The GRIP ice core $\delta^{18}O$ record. The thin grey line is the record plotted versus the previous (modeled) ss09sea timescale [Johnsen et al., 2001]. The heavy black line is plotted versus the new GICC05 timescale on the basis of layer-counting down to 14.73 ka [Rasmussen et al., 2006]. (b) The sea level record based on U/Th-dated corals from Barbados [Fairbanks et al., 2005; Peltier and Fairbanks, 2006] (see also Table 1). Error bars indicate 2σ limits. (c) Record of rate of sea level change, determined as the first time derivative of the solid line in Figure 2b.

Auxiliary materials are available in the HTML. doi:10.1029/2006PA001340.
Table 1. U/Th and \(^{14}\)C Ages of the Samples Identifying Meltwater Pulse 1a in the Barbados Coral Record

<table>
<thead>
<tr>
<th>Sample Code</th>
<th>Species</th>
<th>RSL, m</th>
<th>Mean U/Th Age, years B.P.</th>
<th>2(\sigma), years</th>
<th>N</th>
<th>Mean (^{14})C Age, years B.P.</th>
<th>2(\sigma), years</th>
<th>N</th>
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<tbody>
<tr>
<td>RGF 12-9-5</td>
<td>A. palmata</td>
<td>-57.59</td>
<td>12,203</td>
<td>76</td>
<td>1</td>
<td>10,485</td>
<td>70</td>
<td>1</td>
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<td>RGF 12-15-4</td>
<td>A. palmata</td>
<td>-63.78</td>
<td>12,993</td>
<td>34</td>
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<td>11,464</td>
<td>78</td>
<td>4</td>
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<td>A. palmata</td>
<td>-69.06</td>
<td>13,555</td>
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<td>11,903</td>
<td>56</td>
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<td>A. palmata</td>
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<td>13,574</td>
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<td>1</td>
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<td>A. palmata</td>
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<td>11,909</td>
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<td>A. palmata</td>
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<td>12,735</td>
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<td>1</td>
<td>12,795</td>
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<td>A. palmata</td>
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<td>12,780</td>
<td>50</td>
<td>2</td>
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<td>A. palmata</td>
<td>-92.49</td>
<td>14,408</td>
<td>40</td>
<td>1</td>
<td>12,780</td>
<td>80</td>
<td>1</td>
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<td>RGF 15-5-3</td>
<td>A. palmata</td>
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<td>14,573</td>
<td>30</td>
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<td>12,730</td>
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<td>1</td>
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<td>A. palmata</td>
<td>-93.89</td>
<td>14,539</td>
<td>46</td>
<td>1</td>
<td>12,680</td>
<td>60</td>
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<td>RGF 9-21-11</td>
<td>A. palmata</td>
<td>-106.09</td>
<td>18,176</td>
<td>112</td>
<td>1</td>
<td>15,405</td>
<td>60</td>
<td>1</td>
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</table>

*Radiocarbon ages are reported without reservoir age subtraction. Ages are after Fairbanks et al. [2005]. All reported \(^{14}\)C ages are conventional \(^{14}\)C ages before 1950 (5568 year half-life, corrected for fractionation). Errors are given at the 2\(\sigma\) level for both \(^{14}\)C and \(^{230}\)Th ages. All \(^{230}\)Th ages are expressed as years before 1950. The relative sea levels (RSL) are the depth of Acropora palmata recovery corrected for a constant uplift of 0.34 m/kyr (\(^{230}\)Th ages are used for the uplift correction). Mwp-1a is bracketed in Barbados between samples from cores 9 and 12.

Note that the total uncertainty of the GICC05 timescale, which amounts to only 186 years at around the Bølling warm transition, mainly consists of the maximum counting error (related to the number of annual layers that were difficult to interpret) and a possible bias (because the rules used for identifying annual layers cannot be independently validated). Comparison of the GICC05 timescale with the INTCAL04 radiocarbon calibration curve based on volcanic horizons [Rasmussen et al., 2006] indicates that the maximum counting error in GICC05 is a conservative and adequate measure of the total uncertainty of the GICC05 timescale in this interval.

For the first time, the age models of both ice cores and sea level evolution are constrained with sufficient accuracy to permit sensible comparisons. If the 186-year uncertainty in the GICC05 age of 14.64 ka for the Bølling onset was entirely systematically biased toward the younger bound, then its youngest possible age would be 14.45 ka. Note that it is unlikely that the errors in the GICC05 timescale are distributed in such a systematic manner, especially given the excellent agreement with ages from Cariaco Basin [Lea et al., 2003]. On the basis of RGF 9-8-2, the maximum age for the onset of mwp-1a would be 14.11 (1\(\sigma\)) or 14.14 ka (2\(\sigma\)). Hence mwp-1a apparently lags the Bølling onset by 5 to 6 centuries. Even when pushing the edges of the confidence limits in the two timeframes, mwp-1a lags the Bølling onset by 3 centuries. This firmly corroborates the age offset suggested independently by radiocarbon results, which by themselves are insufficiently precise to be conclusive owing to reservoir age uncertainties (see Auxiliary Materials). Consequently, there is no longer a reasonable case to assume that mwp-1a coincided with the Bølling onset. We therefore reject the hypothesis that mwp-1a may have triggered the abrupt onset of the Bølling warm period [Kienast et al., 2003; Weaver et al., 2003].

The new time constraints instead confirm a previous suggestion [Bard et al., 1996] that mwp-1a was associated with the termination of the Bølling warm period (Figures 2a–2c). Meltwater addition started within the Bølling, possibly as a direct response to high-latitude warming during that period [McManus et al., 2004], and culminated in a meltwater peak at 13.9 ka that coincides with the sharp “Older Dryas” (GI-1d) [Björcck et al., 1998] cooling event. At issue, however, is whether there was a change in NADW flow intensity coincident with mwp-1a and the Older Dryas.

3. Methods

Core TTR-451 was recovered from Eirik Drift in 2003 during a research cruise of R/V Professor Logachev (latitude 55°31’N, longitude 44°54’W, water depth 1927 m) (Figure 1). As a contourite deposit formed where the combined flow of NADW from the Nordic Seas rounds the tip of southern Greenland [Chough and Hesse, 1985], Eirik Drift is a key monitoring site for NADW flow intensity. We present AMS\(^{14}\)C datings (Table 1, Figure 3), and magnetic property measurements for core TTR-451 (Figures 3b and 4e), along with ice rafted debris (IRD) counts of lithic grains (>150 \(\mu\)m) per gram of dry sediment (Figure 4d). Details of methods for each type of analysis are given below.

3.1. Chronology of Core TTR-451

The chronology of TTR-451 is primarily constrained by seven AMS\(^{14}\)C datings of monospecific samples of left-coiling *N. pachyderma* that have been calibrated using Calib5.0.1 with a reservoir age correction \(\Delta R = 0\) (Table 2). Within these constraints, the age model is fine-tuned using a robust correlation between the TTR-451 magnetic susceptibility (\(k\)) record and the GRIP ice core \(\delta^{18}\)O record (see correlation lines in Figure 3), similar to methods used previously [e.g., Dokken and Jansen, 1999; Kissel et al., 1999]. The fine-tuned age model remains well within the 2\(\sigma\) bounds on four of the calibrated radiocarbon datings (squares in Figure 3a), just within the 2\(\sigma\) bounds for one dating, and just outside the 2\(\sigma\) bounds for two datings (Figure 3a). We note that the fine-tuned age model remains well within the 2\(\sigma\) bounds of the calibrated AMS\(^{14}\)C during intervals for which \(\Delta R\) values is generally thought to have been close to zero [Hughen et al., 2000; Bondevik et al., 2006]. Our fine-tuned age model suggests somewhat enhanced \(\Delta R\) values for the three radiocarbon datings from the Younger Dryas (YD) and Heinrich event 1 (H1).
intervals (Figure 3a), again in agreement with previous suggestions that $^{14}$C reservoir ages were substantially increased during those times [Hughen et al., 2000, 2004; Bondevik et al., 2006]. We approximate $D_R$ for those datings using the difference between the ages derived from the magnetic susceptibility tie points (absolute calendar ages) and the calibrated AMS $^{14}$C datings (Table 2). The inferred $D_R$ value for the Younger Dryas ($D_R = 437$ years, Table 2) compares well (just within $1\sigma$) with independently derived $D_R$ estimates of 371 years, based on cores from the Norwegian margin [Bondevik et al., 2006].

3.2. IRD Counts

Dried samples were weighed, wet sieved (>63 µm), and oven dried. The dried sand fraction was dry sieved and each fraction weighed. Suitable aliquots of the coarse (>150 µm) fraction were weighed and the particles counted under a binocular microscope. On average, ~500 grains were counted per sample, and the grain types distinguished. The IRD counts are shown in Figure 4d, and a distinct ~500 year IRD minimum reflects the Bølling warm period, lending further support to our age model for core TTR-451.
Table 2. AMS$^{14}$C Measurements for Core TTR-451

<table>
<thead>
<tr>
<th>Depth, cm</th>
<th>Species</th>
<th>Radiocarbon Conventional Age</th>
<th>Calibrated Age (calib 5.0.1.)</th>
<th>2σ Age Range</th>
<th>Inferred ΔR</th>
<th>2σ Range for the ΔR</th>
</tr>
</thead>
<tbody>
<tr>
<td>12.00</td>
<td>N. pachyderma (lc)</td>
<td>10,905 ± 60 B.P.</td>
<td>12,535 B.P.</td>
<td>12,213–12,715</td>
<td>437</td>
<td>96–650</td>
</tr>
<tr>
<td>40.00</td>
<td>N. pachyderma (lc)</td>
<td>12,220 ± 55 B.P.</td>
<td>13,692 B.P.</td>
<td>13,495–13,800</td>
<td>801</td>
<td>470–1204</td>
</tr>
<tr>
<td>57.50</td>
<td>N. pachyderma (lc)</td>
<td>12,690 ± 55 B.P.</td>
<td>14,141 B.P.</td>
<td>13,986–14,527</td>
<td>949</td>
<td>760–1408</td>
</tr>
<tr>
<td>76.00</td>
<td>N. pachyderma (lc)</td>
<td>12,825 ± 55 B.P.</td>
<td>14,420 B.P.</td>
<td>14,154–14,843</td>
<td>10,905 ± 60 B.P.</td>
<td>12,535 B.P.</td>
</tr>
<tr>
<td>83.50</td>
<td>N. pachyderma (lc)</td>
<td>13,460 ± 55 B.P.</td>
<td>15,411 B.P.</td>
<td>15,136–15,769</td>
<td>801</td>
<td>470–1204</td>
</tr>
<tr>
<td>94.25</td>
<td>N. pachyderma (lc)</td>
<td>14,750 ± 60 B.P.</td>
<td>17,196 B.P.</td>
<td>16,720–17,568b</td>
<td>499</td>
<td>760–1408</td>
</tr>
<tr>
<td>102.25</td>
<td>N. pachyderma (lc)</td>
<td>14,890 ± 60 B.P.</td>
<td>17,432 B.P.</td>
<td>16,975–17,839</td>
<td>76.00</td>
<td>40.00</td>
</tr>
</tbody>
</table>

$^{a}$AMS$^{14}$C measurements measured from monospecific samples of N. pachyderma, left coiling (lc), and their calibrated (calendar) ages for core TTR-451. Inferred ΔR values are also presented for AMS$^{14}$C datings that represent the Younger Dryas and Heinrich event 1.

$^{b}$The 2σ errors for the ages of the GRIP δ$^{18}$O ice core record on the ss09sea timescale are not constrained [Johnsen et al., 2001] but are considered larger than those for the GICC05 timescale (186 years at the Bølling transition).

3.3. Magnetic Measurements

[12] The ratio of susceptibility of the anhysteretic remanent magnetization (κARM) to low-field magnetic susceptibility (κ) for core TTR-451 is shown in Figure 4e. These data were obtained from discrete samples that have been cross-validated by comparison with additional u-channel κARM measurements (not shown) and the higher-resolution whole core magnetic susceptibility record (Figure 3b). The whole core magnetic susceptibility was measured with a Bartington Instruments MS2E1 point sensor that was placed in contact with the surface of a split core (labeled as “whole core” in Figure 3b). The discrete sample magnetic susceptibility measurements were made using a Kappabridge KLY-4 magnetic susceptibility meter, and κARM was imparted using a 50 $\mu$T bias field and an alternating field of 100 mT, with measurements made using a 2G-Enterprise cryogenic magnetometer.

4. Results

[13] Previous studies have invoked a relationship between large meltwater pulses into the Atlantic Ocean and coincident cooling over large parts of the Northern Hemisphere, through reduction of deepwater formation in the Nordic Seas and the associated oceanic poleward heat transport [e.g., Broecker, 1991; Rahmstorf, 1995]. Our new proxy record for the strength of NADW overflow from Eirik Drift (Figure 4e) allows us to test the hypothesis that mwp-1a may have caused the termination of the Bølling warm period.

[14] The κARM/κ ratio is a proxy for magnetic mineral grain size when sediments are magnetically dominated by (titanom) magnetite [Banerjee et al., 1981; Verosub and Roberts, 1995]. Previous work supports the assumption of constant magnetic mineralogy along the NADW flow path in the vicinity of Eirik Drift [Kissel et al., 1999; Laj et al., 2002]. Eirik Drift represents a key region for deposition of suspended matter transported out of the Nordic Seas, especially for transportation via deepwater overflow through Denmark Strait (between Iceland and Greenland). The magnetic mineral content of sediments in that region originates from a single common source (the Nordic basaltic province) [Kissel et al., 1999; Laj et al., 2002]. Our new κARM/κ record from Eirik Drift therefore reflects variations in the size of the coarsest magnetic grains that can be carried by NADW out of the Nordic Seas and that settle out on Eirik Drift as NADW rounds the southern tip of Greenland. Lower κARM/κ values represent coarser grain sizes [Banerjee et al., 1981], and therefore increased intensity of NADW formation (Figure 4e; note the inverted scale).

[15] Our TARM/κ record for core TTR-451 is shown in Figure 4e, along with another (independently dated) record of NADW flow rate, which is based on the $^{234}$Pa/$^{230}$Th ratio in core GGC5 from Bermuda Rise [McManus et al., 2004]. The suggestion that magnetic mineral concentration and grain size reflect changes in NADW flow strength [Kissel et al., 1999; Laj et al., 2002] is empirically corroborated by the high degree of structural similarity between our κARM/κ record and the $^{234}$Pa/$^{230}$Th record from core GGC5 (Figure 4e). Note that the apparent offset at the Bølling transition results only from sample resolution (this is verified by a sharper shift in our higher-resolution, “whole core” measurements (not shown) that agrees closely with the $^{234}$Pa/$^{230}$Th shift at that time). Compared to the $^{234}$Pa/$^{230}$Th record, our κARM/κ record offers a higher temporal resolution between the onset of the Bølling and the onset of the Younger Dryas (Figure 4e). A distinct NADW slowdown is suggested during the cooling from about 14.2 ka that culminated in the Older Dryas at around 14.0 ka. This suggests that the Bølling warm period may have been terminated by a reduction in deepwater formation.
in the Nordic Seas, and consequently reduced oceanic poleward heat transport, coincident with mwp-1a.

5. Discussion and Wider Implications

[16] Our compilation of records demonstrates that direct climate impacts are not proportional to either the magnitude, or the rate, of meltwater addition (compare Figure 4a with Figures 4b and 4c), and neither is the NADW flow intensity (compare Figure 4e with Figures 4b and 4c). First, the dramatic mwp-1a event is found to be associated with only a slowdown (not shutdown) of NADW flow, and only a brief (100–150 years) direct climatic anomaly. Second, the widespread and long-lasting Younger Dryas cold reversal (Figure 4a) clearly lacks a discernible meltwater pulse (Figures 4b and 4c), but it was characterized by significant NADW slowdown (not shutdown) (Figure 4e). Third, there is little evidence [Hanebuth et al., 2000] for rapid sea level rise/meltwater addition during the 23 kyr interval centered on 16 ka (“Heinrich event 1” [Hemming, 2004; McManus et al., 2004]), but this major cold period was marked by near NADW shutdown [McManus et al., 2004] (Figure 4e).

[17] Nonlinear responses of ocean circulation to the magnitude and rate of meltwater additions may be expected in a system with different quasi-stable climate states and abrupt meltwater-driven transitions [Rahmstorf, 1995; Ditlevsen, 1999]. However, it has also been argued that the location and nature of meltwater entry may be more important for NADW formation than the magnitude or rate [Moore, 2005; Tarasov and Peltier, 2005]; smaller surface-bound additions in critical locations could outweigh large additions elsewhere and over greater depth ranges (hypervincal additions).

[18] A considerable component of meltwater associated with mwp-1a is thought to have entered the ocean via the Gulf of Mexico [Flower et al., 2004], in a hypervincal manner [Aharon, 2005], and strong mixing with seawater would have reduced its impact on NADW formation [Tarasov and Peltier, 2005; Aharon, 2005]. Conversely, surface (iceberg) meltwater injection from Hudson Strait during Heinrich event 1, although not dramatically evident in the sea level record [Fairbanks, 1989; Aharon, 2005; Peltier and Fairbanks, 2006] (Figures 4b and 4c), may have sufficiently affected the Nordic Seas to cause a collapse of NADW formation (Figure 4e). The pronounced Younger Dryas event, associated with severely reduced NADW formation (Figure 4e), has also been ascribed to a surface (iceberg) meltwater flux, that was small enough to remain undetectable in terms of sea level, into this critical region (from the Arctic [Moore, 2005; Tarasov and Peltier, 2005]). A meltwater signal into that region has been detected on the SE Greenland shelf [Jennings et al., 2006].

[19] The combined results presented here demonstrate that NADW formation and its associated climatic impacts are not simply governed by the magnitude and/or rate of meltwater addition. If indeed the climate forcing was dependent on freshwater input, then (small) freshwater additions targeted on the Arctic/Nordic Seas appear to carry a much greater risk of disrupting NADW formation. Alternatively, it should be considered that the inferred nonlinear responses of ocean circulation to the magnitude and rate of meltwater additions indicate that meltwater input was not necessarily the primary driver. To evaluate this, new records are required that constrain other potential aspects of the ocean-climate interaction, such as sea ice feedbacks [Li et al., 2005; Wunsch, 2006].

6. Conclusions

[20] Comparison of the GRIP ice core δ18O record on the new layer-counted GICC05 timescale with the better constrained U/Th-dated sea level record conclusively demonstrates that mwp-1a coincided with the Older Dryas and not the Bolling warming. By combining a new proxy record of NADW flow intensity from Eirik Drift with the Barbados sea level and Greenland ice core δ18O records, we show that at the time of mwp-1a and the Older Dryas, there was a brief reduction (not a shutdown) in NADW flow intensity. However, our combination of proxy records also demonstrates that more extreme cooling events (Heinrich event 1 and the Younger Dryas), which were not associated with meltwater pulses large enough to significantly affect the sea level record, were characterized by nearly collapsed NADW formation. This suggests either that there is a fundamental nonlinearity between the rate and magnitude of meltwater injection and the rate of NADW formation, with perhaps a greater importance of the location and depth-distribution of meltwater injection, or that the primary mechanism for some climate transitions lies elsewhere in the ocean-climate-atmosphere system (e.g., sea ice feedbacks).

[21] Acknowledgments. This study was supported by the Natural Environment Research Council as part of the RAPID and QUEST programmes. We acknowledge the crew and scientists onboard R/V Professor Logachev, BOSCORF, and Leibniz Labor, Kiel, for the AMS14C datings. We would also like to thank the referees who helped to improve our manuscript. R. G. Fairbanks’ sea level and radiometric dating programs were supported by U.S. NSF grants ATM03-27722 and OCE99-11637. This is LDEO contribution number 6965.

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