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# Freshwater input and abrupt deglacial climate change in the North Atlantic

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[1] Greenland ice core records indicate that the last deglaciation (~7–21 ka) was punctuated by numerous abrupt climate reversals involving temperature changes of up to 5°C–10°C within decades. However, the cause behind many of these events is uncertain. A likely candidate may have been the input of deglacial meltwater, from the Laurentide ice sheet (LIS), to the high-latitude North Atlantic, which disrupted ocean circulation and triggered cooling. Yet the direct evidence of meltwater input for many of these events has so far remained undetected. In this study, we use the geochemistry (paired Mg/Ca-$\delta^{18}$O) of planktonic foraminifera from a sediment core south of Iceland to reconstruct the input of freshwater to the northern North Atlantic during abrupt deglacial climate change. Our record can be placed on the same timescale as ice cores and therefore provides a direct comparison between the timing of freshwater input and climate variability. Meltwater events coincide with the onset of numerous cold intervals, including the Older Dryas (14.0 ka), two events during the Allerød (at ~13.1 and 13.6 ka), the Younger Dryas (12.9 ka), and the 8.2 ka event, supporting a causal link between these abrupt climate changes and meltwater input. During the Bølling-Allerød warm interval, we find that periods of warming are associated with an increased meltwater flux to the northern North Atlantic, which in turn induces abrupt cooling, a cessation in meltwater input, and eventual climate recovery. This implies that feedback between climate and meltwater input produced a highly variable climate. A comparison to published data sets suggests that this feedback likely included fluctuations in the southern margin of the LIS causing rerouting of LIS meltwater between southern and eastern drainage outlets, as proposed by Clark et al. (2001).

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## 1. Introduction

[2] The discovery of past abrupt climate changes and their possible occurrence during future global warming demands investigation into their causes [e.g., *Alley et al.*, 2003]. Numerous abrupt climate reversals accompanied the transition from the last glacial maximum (~21 ka) to the present warm interglacial period [e.g., *North Greenland Ice Core Project* (*NGRIP*) *members*, 2004], and several of these climate reversals have been associated with a reduction in the strength of the Atlantic meridional overturning circulation (AMOC) and its attendant heat flux [e.g., *Clark et al.*, 2001; *McManus et al.*, 2004; *Ellison et al.*, 2006]. The AMOC consists of a poleward warm saline surface water flow (the North Atlantic Current, NAC) from the tropics, that undergoes subsequent cooling and deep convection at high northern latitudes, feeding a deep southward return flow. Modeling studies [*LeGrande et al.*, 2006; *Meissner and Clark*, 2006; *Stouffer et al.*, 2006; *Clarke et al.*, 2009; *Liu et al.*, 2009], corroborated by proxy data [*McManus et al.*, 2004], indicate that increased freshwater input to the sites of deep convection can weaken AMOC and trigger cooling. Indeed, the abrupt cooling associated with Heinrich Event 1 (~16–17 ka) was likely caused by a collapse of the North American Laurentide ice sheet (LIS), which discharged an armada of icebergs and freshwater into the Atlantic [*Hemming*, 2004], thereby triggering AMOC reduction [*McManus et al.*, 2004]. However, the causes of many of the remaining deglacial climate reversals are uncertain. To constrain the possible involvement of freshwater in these events requires a detailed deglacial history of freshwater input to the northern North Atlantic.

[3] During the last deglaciation (~21–7 ka) melting of ice sheets caused sea level to rise by ~120 m, with the largest contribution from the LIS [*Fairbanks*, 1989; *Peltier*, 2005]. Records of sea level indicate this increase was not smooth, but included periods of rapid sea level rise suggesting faster melting [*Fairbanks*, 1989; *Bard et al.*, 1996; *Hanebuth et al.*, 2000; *Yokoyama et al.*, 2000; *Clark et al.*, 2004]. Prominent increases (between 10 and 20 m in <500 years) occur at 19 ka, ~14.6 ka and possibly ~11 ka, the latter two being termed meltwater pulses 1a and 1b (MWP-1a and MWP-1b), respectively [*Fairbanks*, 1989]. The source and relative contributions to meltwater pulses from different ice sheets is heavily debated, especially concerning MWP-1a [e.g., *Clark et al.*, 1996; *Weaver et al.*, 2003; *Peltier*, 2005]. Sea level fingerprinting suggests the Antarctic ice sheet as the primary source of MWP-1a [*Clark et al.*, 2002; *Bassett et al.*, 2005], countering earlier

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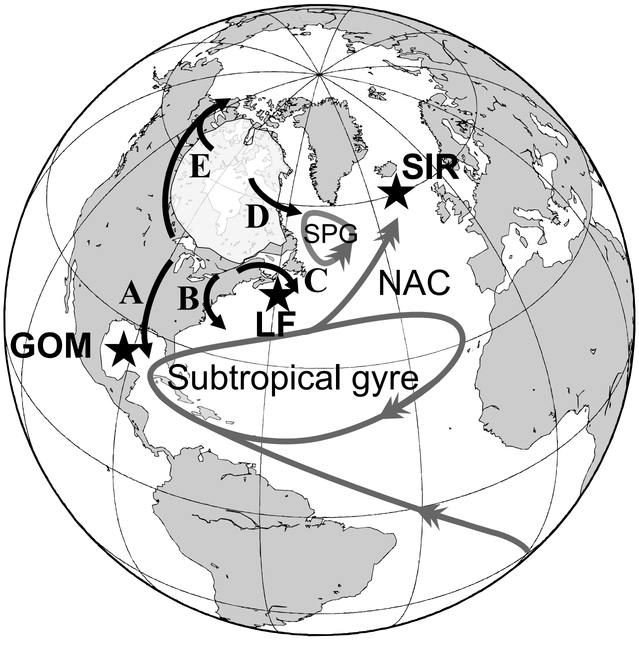


**Figure 1.** Schematic map showing North Atlantic surface circulation, the margin of the LIS at ~13 ka (white shaded region), and meltwater runoff routes (black single-headed arrows) [*Clark et al.*, 2001]. Letters identify meltwater runoff routes: route A, Mississippi River; route B, Hudson River; route C, Saint Lawrence River; route D, Hudson Strait; route E, Mackenzie River. Study sites are indicated by black stars: GOM, Gulf of Mexico [*Aharon*, 2003, 2006]; LF, Laurentian Fan [*Keigwin et al.*, 2005]; SIR, South Iceland Rise (this study). North Atlantic surface circulation (gray double-headed arrows) summarized as warm saline subtropical gyre, cold fresh subpolar gyre (SPG), and the North Atlantic Current (NAC), transporting subtropical water poleward.

suggestions of a LIS source [e.g., *Keigwin et al.*, 1991; *Peltier*, 1994]. Regarding MWP-1b, the existence of an abrupt sea level rise has been questioned by Tahiti coral data [*Bard et al.*, 1996; *Deschamps et al.*, 2009]. Regardless of the contribution the LIS made toward MWP-1a and MWP-1b (recent estimates suggest a LIS contribution to MWP-1a of <30% [*Carlson*, 2009]), disintegration of the LIS undoubtedly provided a significant meltwater flux to the North Atlantic during deglaciation, as implied by planktonic foraminiferal $\delta^{18}$O records from close to the outlets of the LIS [e.g., *Aharon*, 2003, 2006; *Flower et al.*, 2004; *Keigwin et al.*, 1991, 2005]. Fluctuations in the southern margin of the LIS caused episodic rerouting of this continental runoff between southern, eastern and northern outlets [*Clark et al.*, 2001] (Figure 1) and catastrophic drainage of proglacial lakes into the North Atlantic [*Teller et al.*, 2002]. This has led to the established view that the deglacial North Atlantic was subject to numerous freshwater discharge and rerouting events.

[4] It has been suggested that periods of enhanced meltwater input to the Atlantic and episodic rerouting events, via AMOC weakening, triggered several prominent cold intervals of the deglaciation, including the Older Dryas (OD), the intra-Allerød Cold Period (IACP), the Younger Dryas (YD) and the 8.2 ka cold event [*Broecker et al.*, 1989; *Keigwin et al.*, 1991; *Lehman and Keigwin*, 1992; *Barber et al.*, 1999; *Clark et al.*, 2001; *Donnelly et al.*, 2005; *Broecker*, 2006; *Ellison et al.*, 2006; *Stanford et al.*, 2006; *Came et al.*, 2007]. Yet direct evidence of increased LIS meltwater input reaching the northern North Atlantic for most of these events is limited, and largely based on planktonic foraminiferal $\delta^{18}$O [e.g., *Sarnthein et al.*, 1995; *Lehman and Keigwin*, 1992] which may include a temperature component to any signal (notwithstanding recent foraminiferal paired Mg/Ca-$\delta^{18}$O work [*Came et al.*, 2007; *Thornalley et al.*, 2009] recording the final drainage of Lake Agassiz into the North Atlantic, that resulted in the 8.2 ka cold event). Without evidence illustrating that these meltwater events reached the high-latitude North Atlantic, their link to any subsequent abrupt climate event remains hypothetical. Furthermore, determining the phasing between existing marine evidence for increased freshwater fluxes and the high-resolution Greenland ice core climate record has been hindered by uncertainty in chronologies. However, this problem can be solved by reconstructing meltwater input and climate proxy data from the same core, facilitating the accurate determination of relative phasing between the two variables.

[5] In this study we report the deglacial history of meltwater input to the northern North Atlantic using the geochemistry of planktonic foraminifera from a sediment core south of Iceland. Our record can be placed on the same timescale as ice cores and therefore provides a direct comparison between the timing of meltwater input and the abrupt climate changes of the last deglaciation.

## 2. Methods

### 2.1. Core Material

[6] We use material from sediment core RAPiD-15-4P (62°17.58′N, 17°08.04′W, 2133 m water depth), located under the path of the northward flowing NAC, on the South Iceland Rise (Figure 1). The core was sampled at 1 cm intervals throughout the deglaciation. Data from RAPiD-15-4P was combined with published Holocene data from the nearby core RAPiD-12-1K [*Thornalley et al.*, 2009].

[7] Three major Icelandic tephra layers are identified within RAPiD-15-4P by tephra abundance counts, visual examination, and electron microprobe analysis of major element composition. These tephra correspond to the Saksunarvatn Ash, Vedde Ash and a basaltic tephra originating from the Katla volcano during the Older Dryas, found at 1573.0 m in the North Greenland Ice Core Project (NGRIP) ice core [*Mortensen et al.*, 2005]. These tephra have ages of 10.347 ka, 12.171 ka and 14.02 ka, respectively, using the GICC05 age from *Rasmussen et al.* [2006].

### 2.2. Bioturbation Effects

[8] The presence of discrete tephra layers within the core allows the extent of bioturbation to be examined. Following the work of *Bard et al.* [1994] on the Vedde Ash, we compare the downcore distribution of tephra abundance with the modeled effect of intense mixing on an instanta-

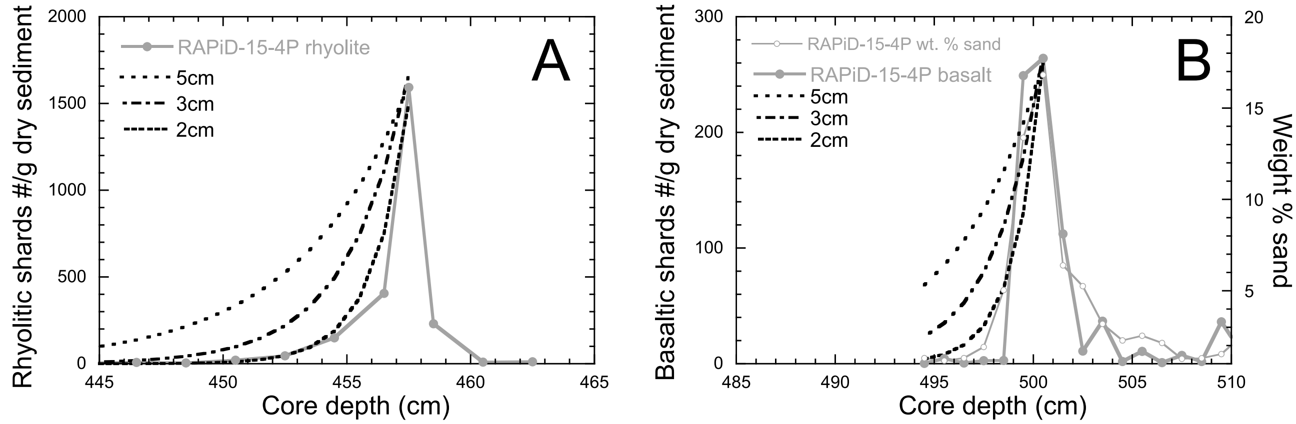


**Figure 2.** Volcanic shard abundance (150–250 $\mu$m) and modeled distribution following an instantaneous input of tephra and subsequent bioturbation using a variety of mixing depths. (a) For the Vedde Ash at 12.171 ka. (b) For the Katla Ash at 14.0 ka, also shown is the weight percent sand fraction which also indicates the upward mixing of tephra, including that within the 63–150 $\mu$m fraction.

neous input of tephra, the results of which are shown in Figure 2. The exponential decrease in the abundance of rhyolite shards subsequent to the input of the Vedde Ash closely matches the expected distribution assuming a mixing interval of 2 cm. The 14.0 ka Katla Ash also shows very little bioturbation (consistent with the observation that this 1–2 cm thick basaltic tephra layer had a relatively sharp base and top within the sediment core) and again suggests a mixing interval of 2 cm or perhaps less. The small modeled mixing depth therefore demonstrates that the effects of bioturbation are minor within RAPiD-15-4P and multicentennial timescale climate signals are faithfully reproduced when sedimentation rates exceed ~10 cm/ka.

### 2.3. Age Model

[9] The age model for RAPiD-15-4P was initially constructed using $^{14}$C accelerator mass spectrometry (AMS) dates on monospecific planktonic foraminifera from abundance maxima (Table 1), and the correlation of the three tephra layers identified within both RAPiD-15-4P and NGRIP. An additional age constraint between the LGM and the onset of the Bølling-Allerød (BA) was provided by benthic $\delta^{18}$O stratigraphy, tying the initial deglacial decrease in benthic $\delta^{18}$O in RAPiD-15-4P [*Thornalley*, 2008] to the same event in the well dated North Atlantic core SU81-18 (38°N, 10°W) [*Waelbroeck et al.*, 2001], located at a latitude where changes in the surface radiocarbon reservoir age ($R_{surf}$) are thought to have been relatively small [*Waelbroeck et al.*, 2001], although increases in $R_{surf}$ of 300–400 years during Heinrich Stadial 1 have been reported in this region [*Siani et al.*, 2001; *Skinner and Shackleton*, 2004]. $R_{surf}$ corrections were applied to the $^{14}$C dates using the deglacial estimates of *Waelbroeck et al.* [2001]. We use a larger $R_{surf}$ (800 years) during the BA than compared to the Holocene (400 years), as suggested by data from *Haflidason et al.* [1995] and the open North Atlantic site of *Peck et al.* [2006], rather than the estimate of 400–500 years from Norwegian coastal waters [*Bondevik et al.*, 2006], because we consider the open ocean sites are more representative of the conditions at RAPiD-15-4P. During the

**Table 1.** The $^{14}$C AMS Dates Used to Support the Final Age Model and Calculate Surface Reservoir Ages[a]

| Laboratory Code | Sample Depth (cm) | Planktic Foraminifera | $^{14}$C Age (years B.P.) | ±1 $\sigma$ Error (years) | Converted Calendar Age Using $R_{surf}$ Estimates From Published Data[b] | Stratigraphic Age[c] | $R_{surf}$[d] (years) |
|---|---|---|---|---|---|---|---|
| SUERC 14086 | 432–433 | *G. bulloides* | 10,198 | 40 | 10,817 ± 450 | 10,764 | 778 ± 340 |
| SUERC 14087 | 458–459 | *Nps* | 10,965 | 40 | 12,023 ± 320 | 11,984 | 715 ± 230 |
| SUERC 14089 | 472–473 | *G. bulloides* | 11,811 | 42 | 13,077 ± 400 | 12,883 | 1,011 ± 350 |
| SUERC 14090 | 500–501 | *Nps* | 12,636 | 43 | 13,763 ± 460 | 14,020 | 536 ± 290 |
| SUERC 14091 | 520–521 | *Nps* | 14,657 | 48 | 14,994 ± 940 | 15,303 | 1,737 ± 740 |
| SUERC 14092 | 565–566 | *Nps* | 18,666 | 66 | 21,292 ± 940 | (21.292)[e] | - |

[a]Radiocarbon ages were converted to calendar ages using Calib 5.1 IntCal [*Reimer et al.*, 2004]. Years B.P., years before A.D. 1950. By using the assigned stratigraphic age (converted to years B.P.), the contemporaneous age difference ($R_{surf}$) between the atmosphere $^{14}$C age (calculated using IntCal04) and the measured $^{14}$C age of the surface marine sample (planktonic foraminifera samples) can be calculated. Error estimates for $\Delta$R include lab error in $^{14}$C measurement, uncertainty in the $^{14}$C calibration curve, error in NGRIP age, and an estimate of error for picking midpoints/tephra layers based on sedimentation rates and modeled mixing intervals.

[b]Used to create age model based on $^{14}$C dates and tephra (excluding percent *Nps* correlation) shown in Figure 3b. Converted calendar age is in years before A.D. 2000.

[c]Stratigraphic age is in years before A.D. 2000.

[d]$R_{surf}$ values are from this study.

[e]No stratigraphic age control older than 17.2 ka; therefore the age is based solely on the $^{14}$C date with an $R_{surf}$ estimate of 800 years.

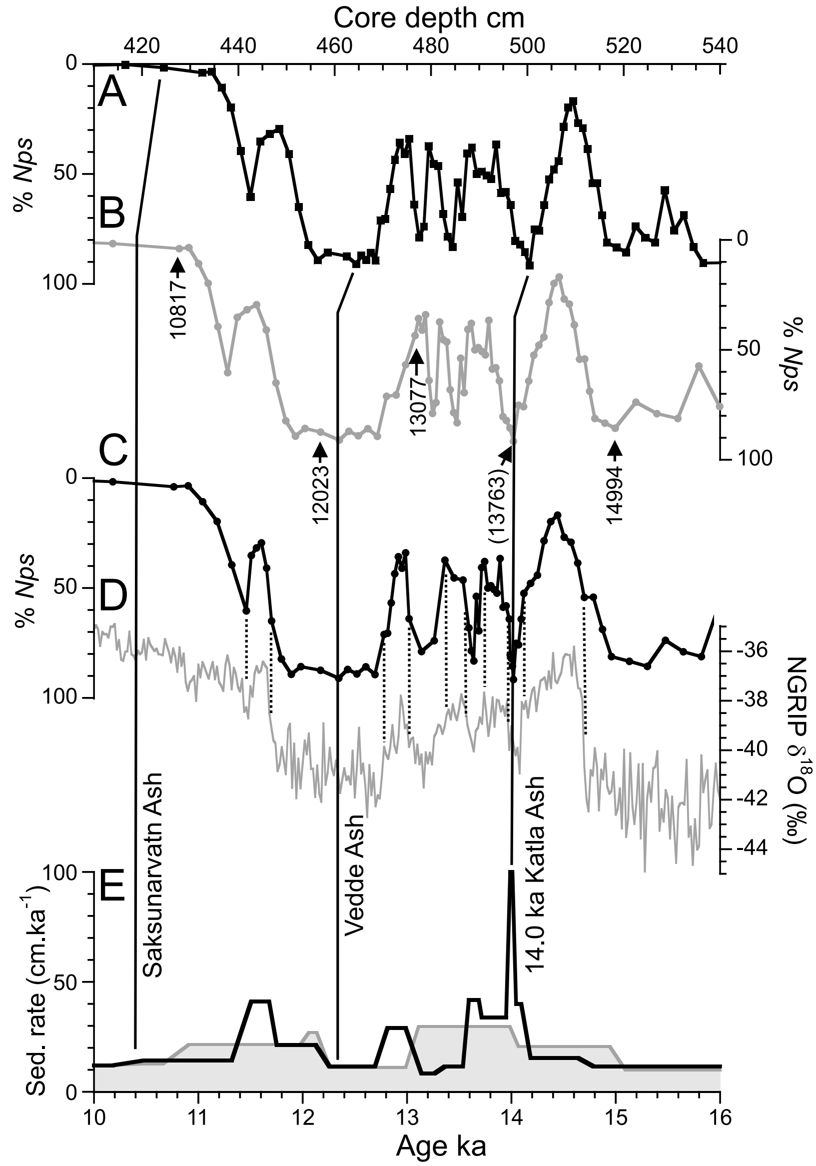


**Figure 3.** Percent abundance of *Nps* plotted on different age models. (a) Percent *Nps* on depth. (b) The percent *Nps* on initial age model, using $^{14}C$ dates (corrected calendar ages shown), tephra tie points (black lines), and benthic $\delta^{18}O$ tie at 17.2 ka. (c) The percent *Nps* on final age model using tephra tie points, correlation of percent *Nps* with NGRIP $\delta^{18}O$ (tie points indicated by dotted lines), and the benthic $\delta^{18}O$ tie point. (d) NGRIP $\delta^{18}O$ [*NGRIP members*, 2004]. (e) Sedimentation rates for the initial (gray line and shaded region) and final (black line) age models.

last glacial maximum, where *Waelbroeck et al.* [2001] do not have constraints, we use a $R_{surf}$ of 800 years, consistent with previous northern North Atlantic studies [*Peck et al.*, 2006; *Knutz et al.*, 2007]. This initial age model, presented in Figure 3b, is subject to large uncertainties associated with the variable $R_{surf}$ of the North Atlantic, likely caused by variable air-sea exchange and flow of the NAC during periods with enhanced sea ice cover and reduced ocean overturning [*Bard et al.*, 1994; *Waelbroeck et al.*, 2001]. Despite this uncertainty, changes in the percent abundance of the polar species *Neogloboquadrina pachyderma* (*Nps*) within RAPiD-15-4P often closely match the climate fluctuations observed within NGRIP, consistent with numerous other studies in the region [e.g., *Lehman and Keigwin*, 1992; *Bond et al.*, 1993; *Haflidason et al.*, 1995; *Rasmussen et al.*, 1996; *Voelker et al.*, 1998; *Knutz et al.*, 2007].

[10] Significant differences in the timing of inferred abrupt climate changes do however exist between RAPiD-15-4P and NGRIP, particularly during the Allerød (~13 ka). There is no evidence suggesting reworking of material in RAPiD-15-4P during this interval. Furthermore, the general structure of the percent *Nps* record matches NGRIP except with regards to the timing of events. Rather than invoke extreme climate scenarios to explain the asynchronous

**Table 2.** Final Age Model for RAPiD-15-4P[a]

| | Sediment Depth (cm) | NGRIP Age (ka) | NGRIP Error (ka) | Estimated Error in RAPiD-15-4P Absolute Age (ka) |
|---|---|---|---|---|
| Saksunarvatn ash | 426.5 | 10.347 | 0.089 | 0.134 |
| NGRIP $\delta^{18}$O tie point 1 | 442.5 | 11.46 | 0.097 | 0.140 |
| NGRIP $\delta^{18}$O tie point 2 | 452.5 | 11.703 | 0.099 | 0.141 |
| Vedde ash | 462.5 | 12.171 | 0.114 | 0.152 |
| NGRIP $\delta^{18}$O tie point 3 | 469.5 | 12.78 | 0.134 | 0.167 |
| NGRIP $\delta^{18}$O tie point 4 | 476.5 | 13.02 | 0.141 | 0.245 |
| NGRIP $\delta^{18}$O tie point 5 | 479 | 13.32 | 0.149 | 0.250 |
| NGRIP $\delta^{18}$O tie point 6 | 482 | 13.58 | 0.156 | 0.185 |
| NGRIP $\delta^{18}$O tie point 7 | 487 | 13.70 | 0.159 | 0.188 |
| NGRIP $\delta^{18}$O tie point 8 | 496.5 | 13.98 | 0.166 | 0.194 |
| Katla basalt | 500.5 | 14.02 | 0.167 | 0.174 |
| NGRIP $\delta^{18}$O tie point 9 | 504.5 | 14.12 | 0.170 | 0.197 |
| NGRIP $\delta^{18}$O tie point 10 | 513.5 | 14.70 | 0.186 | 0.273 |
| Benthic $\delta^{18}$O tie point | 542.5 | (17.2) | - | 0.283 |

[a]NGRIP age is GICC05 timescale, kiloyears before A.D. 2000. Error in NGRIP age from *Rasmussen et al.* [2006]. Error in RAPiD-15-4P includes NGRIP error and estimates on the accuracy of tying percent *Nps* to NGRIP, which primarily depends upon bioturbation and sedimentation rates.

behavior between Iceland and Greenland, it is more probable that these differences are caused by variability in the $R_{surf}$ of the northern North Atlantic and uncertainty in determining its value. The final age model for RAPiD-15-4P (with age control points provided in Table 2) is therefore constructed using the established procedure of correlating sharp abundances changes in percent *Nps* within RAPiD-15-4P to the abrupt climate transitions recorded by NGRIP [e.g., *Haflidason et al.*, 1995; *Peck et al.*, 2006; *Knutz et al.*, 2007], while maintaining the tephra age control points and the benthic $\delta^{18}$O tie at 17.2 ka. Because the core has been tied to NGRIP, dates throughout this study are given in thousands of years before A.D. 2000, unless stated otherwise. The close similarity between the percent *Nps* record and NGRIP $\delta^{18}$O and the absence of any unrealistic changes in sedimentation rate (the only abrupt change follows a large input of ash at 14.0 ka) provides confidence in the age assignment. Figure 4 illustrates that this stratigraphic age model is, within error, consistent with the radiocarbon dates.

[11] Following the examples of *Waelbroeck et al.* [2001], *Siani et al.* [2001], *Peck et al.* [2006] and *Knutz et al.* [2007], we use the stratigraphic age constraints to estimate the $R_{surf}$ south of Iceland. Values for calculated $R_{surf}$ are given in Table 1. They are generally in excellent agreement with previous North Atlantic studies [*Bard et al.*, 1994; *Voelker et al.*, 1998; *Waelbroeck et al.*, 2001; *Peck et al.*, 2006; *Knutz et al.*, 2007] supporting increased $R_{surf}$ at the end of Heinrich Stadial 1 (1737 ± 740 years) and shortly following the Vedde Ash (715 ± 230 years). Throughout the BA, $R_{surf}$ increased from 536 ± 290 years to 1011 ± 350 years, perhaps related to the gradual deterioration of climate from the Bølling to the Allerød [e.g., *NGRIP members*, 2004].

## 2.4. Reconstruction of Meltwater Input Using Planktonic Mg/Ca-$\delta^{18}$O

[12] The input of freshwater sourced from melting ice sheets to the ocean produces low seawater $\delta^{18}$O ($\delta^{18}O_{sw}$) that can be detected using planktonic foraminifer $\delta^{18}$O. Because $\delta^{18}O_{sw}$ and salinity are affected by similar processes (such as freshwater input and evaporation), a near linear relationship is observed in the modern ocean [*LeGrande and Schmidt*, 2006], thus $\delta^{18}O_{sw}$ can be utilized as a proxy for salinity, although changing end-members can cause changes in this relationship through time, preventing the accurate quantitative reconstruction of salinity when boundary conditions differ significantly from the modern. $\delta^{18}O_{sw}$ was reconstructed using paired Mg/Ca-$\delta^{18}$O analysis on *Globorotalia inflata* and *Globigerina bulloides*, in which the Mg/Ca ratio of the calcite was used to remove the calcification temperature effect on $\delta^{18}$O [*Elderfield and Ganssen*, 2000]. $\delta^{18}O_{sw}$ was then corrected for global ocean shifts caused by ice volume changes. The global ice volume corrected seawater $\delta^{18}$O values ($\delta^{18}O_{sw\text{-}ivc}$) were calculated using a 1‰ whole ocean change between the LGM and the latest Holocene scaled to a 120 m sea level change using the sea level versus age record of *Fairbanks* [1989]. South of Iceland, *G. inflata* lives at the base of the seasonal thermocline (~100–200 m depth) [*Ganssen and Kroon*, 2000; *Cléroux et al.*, 2007], while *G. bulloides* is a near surface dweller (~0–50 m depth) [*Ganssen and Kroon*, 2000].

[13] Typically ~15–40 tests of *G. inflata* (>250 $\mu$m fraction) were analyzed for $\delta^{18}$O and Mg/Ca ratios, following published methods [*Barker et al.*, 2003], and screening for contaminating ferromanganese overgrowths, clay minerals and silicate particles. The Mg/Ca calibrations used by *Thornalley et al.* [2009] were employed: Mg/Ca = $m$ exp (0.1 $T$), where $T$ is calcification temperature and $m$ is 0.675 for *G. inflata* and 0.794 for *G. bulloides*. $\delta^{18}O_{sw}$ was calculated using the paleotemperature equation of *Kim and O'Neil* [1997], and a VPDB-to-SMOW $\delta^{18}$O conver-

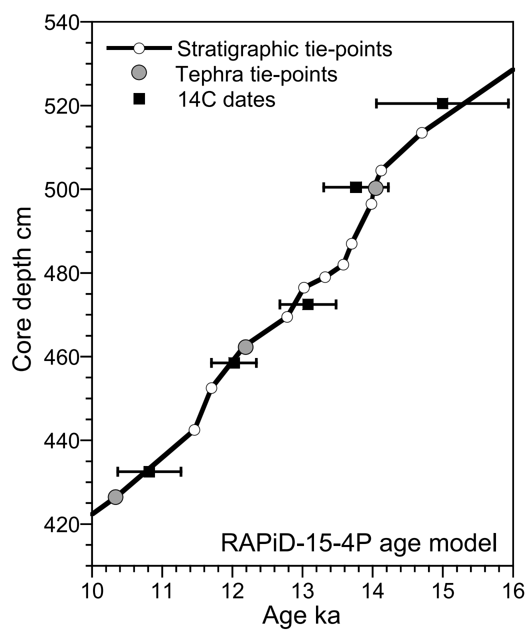


**Figure 4.** Final age model, showing depth versus age relationship. The $^{14}$C dates are corrected for surface reservoir changes and converted to calendar years before A.D. 2000, consistent with the final stratigraphic age model being tied to NGRIP GICC05 age.

sion of 0.27‰ [*Hut*, 1987]. *G. bulloides* does not precipitate calcite in equilibrium with seawater and therefore an offset of 0.6‰ has been applied [*Bemis et al.*, 1998].

[14] Oxygen isotope ratios were determined via gas source mass spectrometry relative to the Vienna Peedee belemnite (VPDB) standard, where analytical precision based on long-term replicates is better than 0.08‰. Analytical precision of Mg/Ca ratios based on replicates of foraminiferal standards is 3%. The estimated standard deviations in absolute temperature and $\delta^{18}O_{sw\text{-}ivc}$ are 1.5°C and 0.34‰, respectively, while those in relative temperature and $\delta^{18}O_{sw\text{-}ivc}$ are 1.0°C and 0.26‰. The relative error estimates include measurement errors, sample heterogeneity, carbonate ion effects and ice volume effect uncertainty but ignore calibration errors, which should be more constant downcore.

### 2.5. Interpreting Deglacial High-Latitude North Atlantic *G. bulloides* and *G. inflata* Mg/Ca-$\delta^{18}$O

[15] At present, RAPiD-15-4P is bathed by the warm saline NAC and faunal assemblages are dominated by transitional (including *G. inflata*) and warm subpolar (including *G. bulloides*) species, while the cold subpolar/polar species *N. pachyderma* (sinistral, *Nps*) is virtually absent (Holocene abundances rarely exceeding 1%). However, downcore records of the percent *Nps* document intervals when this cold species dominated the faunal assemblage (up to 90%), and by inference, the polar front was predominantly situated to the south of the study site, and sea surface temperatures (SSTs) were close to 4°C (95% abundance of *Nps* indicates an SST of <4°C [*Bé and Tolderlund*, 1971]). Intervals containing high percent *Nps* within northern North Atlantic cores are correlated with cold periods recorded in the Greenland ice cores, and abrupt fluctuations in percent *Nps* imply rapid alternation between polar and subpolar/transitional water masses [e.g., *Lehman and Keigwin*, 1992; *Bond et al.*, 1993; *Haflidason et al.*, 1995; *Rasmussen et al.*, 1996; *Voelker et al.*, 1998; *Knutz et al.*, 2007].

[16] During the deglacial section of RAPiD-15-4P (11–17 ka) the average percent *Nps* is 60 ± 21%, fluctuating between ~20% and ~90% (the remaining assemblage is composed of subpolar and transitional species (including *G. bulloides*, *Turborotalia quinqueloba*, *N. pachyderma* (dextral) and *G. inflata*)). These subpolar and transitional species are associated with an increased influence of warm Atlantic water [e.g., *Rasmussen and Thomsen*, 2008] and their presence within the RAPiD-15-4P implies episodic increased influence of this water mass; incursions were infrequent during cold intervals when polar water dominated, but still regular enough to allow a continuous downcore record of *G. bulloides* and *G. inflata* to be produced. The presence of warm Atlantic water in the high-latitude North Atlantic is consistent with the suggestion by *Rasmussen et al.* [1996] and *Rasmussen and Thomsen* [2004] that there was an inflow to the Nordic Seas of a relatively warm Atlantic water mass beneath a cold, low-salinity, surface layer during stadials, and during years when there was a decrease in the influence of surface polar water, relatively warm Atlantic water may have been present at near surface depths. *Sarnthein et al.* [1995] also suggest Atlantic water flowed into the Nordic Seas during the YD, but during HE1 and the LGM, inflow to the Nordic Seas was heavily restricted.

[17] Analysis of downcore sediment records should consider that each sample (typically 1 cm thick) contains material that integrates conditions over many years (~25–100 years within RAPiD-15-4P, with planktonic foraminifera only recording conditions during their seasonal blooms). Thus, the occurrence of two species within a single sample does not imply that the foraminifera lived contemporaneously. For example, it is more likely that the transitional species *G. inflata* and the warm subpolar species *G. bulloides* grew during periods when conditions were favorable for their growth, namely intervals when there was a greater influence and northward incursion of warm Atlantic water, while *Nps* would have bloomed during years when cold subpolar and polar water dominated, and subsequent bioturbation of this 1 cm sample mixed the species together within the core. With this bias in mind, it can be realized that Mg/Ca-$\delta^{18}$O analysis performed on an individual species of planktonic foraminifera, from a site that experienced frequent and large fluctuations in climate that caused significant faunal changes, does not reveal the mean conditions at the site integrated over many annual blooms. Rather, temperature and $\delta^{18}O_{sw}$ signals reconstructed from *G. bulloides* and *G. inflata* Mg/Ca-$\delta^{18}$O will primarily record conditions at the site during incursions of Atlantic waters and monitor any freshening of Atlantic water caused by the input of LIS meltwater. (Note: In contrast, faunal assemblage based SST reconstruction techniques, which utilize the species' preferences for different water masses/climate conditions, do provide an estimate of mean conditions over many years, through examining changes in relative abundance.)

## 3. Results

[18] *G. bulloides* and *G. inflata* $\delta^{18}$O, Mg/Ca derived temperatures and $\delta^{18}O_{sw\text{-}ivc}$ data are presented in Figure 5. *G. bulloides* and *G. inflata* display similar long-term trends in $\delta^{18}$O, with *G. bulloides* recording lighter values, consistent with its shallower habitat depth.

[19] Between 17 ka and ~9 ka, Mg/Ca derived temperatures for the two species range between 6.5°C and 11.5°C, with typical values of 8°C–10°C. The near-identical, warm, and near-constant temperatures recorded by the two species during the deglaciation suggest they are both calcifying during intervals when there is an increased influence of Atlantic water, producing conditions favorable for their growth. Mg/Ca derived temperatures prior to the BA warming and during the YD cold interval are, as expected, much warmer than implied by the high percent *Nps* (~4°C), yet as discussed, this is an artifact of the habitat bias of the species, and the Mg/Ca-derived temperatures should not be interpreted as a multidecadal/centennial mean summer SST. Warmer temperatures are recorded during the early Holocene, and it is interesting to note that the warming into the Holocene recorded by *G. bulloides* begins during the mid YD, similar to the trend observed in *G. bulloides* Mg/Ca data from off the coast of the British Isles [*Peck et al.*, 2008], and also consistent with evidence for increased

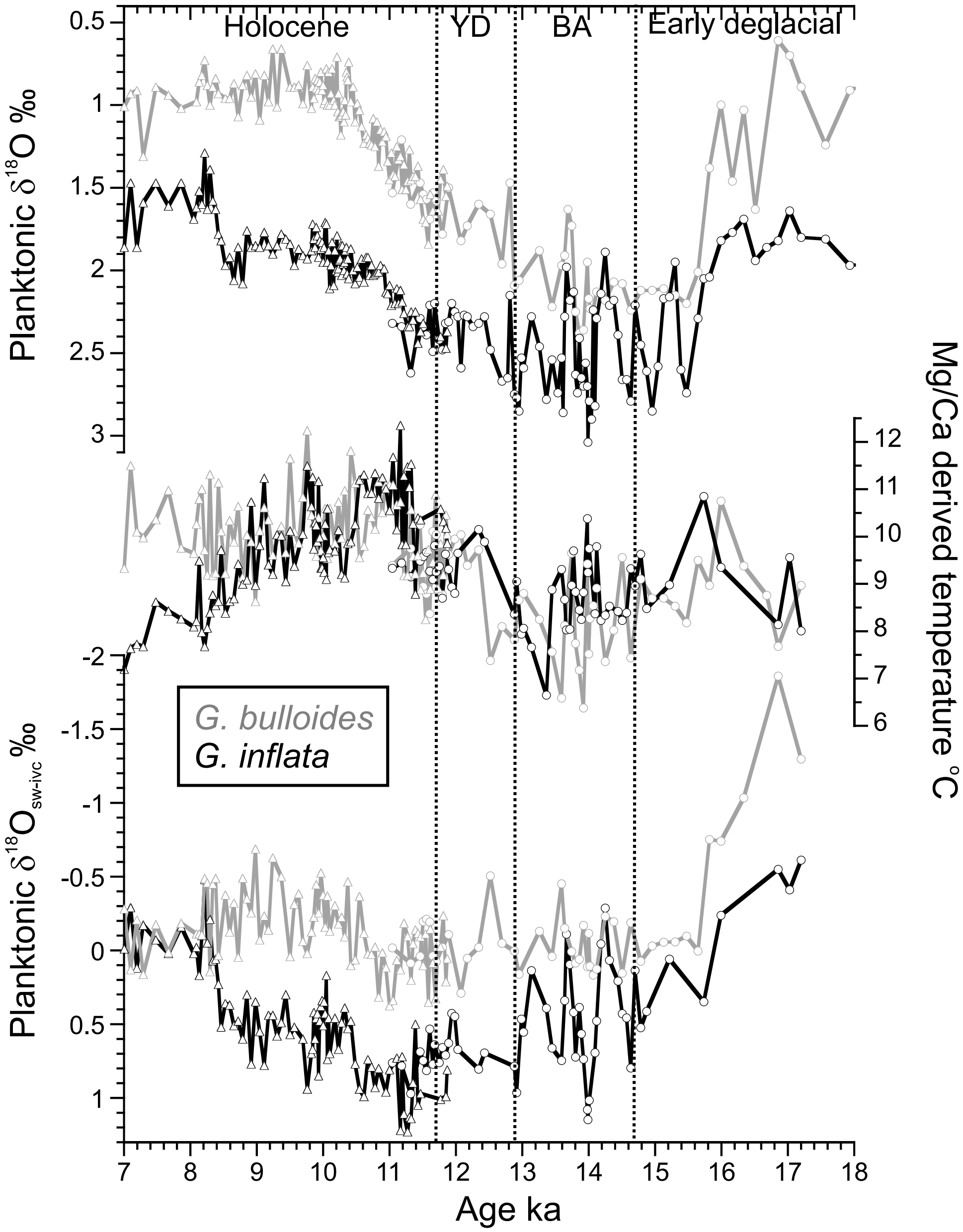


**Figure 5.** Geochemical records for *G. bulloides* and *G. inflata* from RAPiD-15-4P. Calibrations and conversion are discussed in section 2 of the main text. Here $\delta^{18}O_{sw\text{-}ivc}$ is seawater $\delta^{18}O$ corrected for whole ocean changes in $\delta^{18}O$ caused by global land-based ice volume changes.

inflow of subpolar waters into the Nordic Seas and atmospheric circulation changes during the late YD [*Bakke et al.*, 2009]. The surface ocean warming recorded at all these sites was likely caused by an increased northward flow of the NAC, presumably coupled to the small, gradual, increase in AMOC strength during the YD following the initial slowdown, as suggested by Pa/Th ratios from Bermuda Rise [*McManus et al.*, 2004] and atmospheric $^{14}C$ data [*Hughen et al.*, 1998].

[20] The derived $\delta^{18}O_{sw\text{-}ivc}$ records for both species indicate fresh conditions between 17.0 ka and 15.8 ka, with *G. bulloides* recording extremely light $\delta^{18}O_{sw\text{-}ivc}$ values (down to −1.9‰). Both species then show an abrupt increase in $\delta^{18}O_{sw\text{-}ivc}$ at 15.8 ka, after which *G. bulloides* $\delta^{18}O_{sw\text{-}ivc}$ remains relatively constant, except for a number of low-amplitude (∼0.4‰) centennial events between 13 and 15 ka, a 0.8‰ decrease during the early YD (only one data point), and a previously reported ∼0.5‰ broad decrease throughout the early Holocene [*Thornalley et al.*, 2009]. The inferred lower salinity of *G. bulloides*, compared to *G. inflata*, throughout the deglaciation is likely caused by its shallower, near surface habitat, where it will be greater influenced by low-salinity surface waters [*Thornalley et al.*, 2009]. *G. inflata* $\delta^{18}O_{sw\text{-}ivc}$ values show a long-term trend suggesting more saline background conditions between 15.5 ka and 8.4 ka. Between 15.8 and 12.6 ka, super-

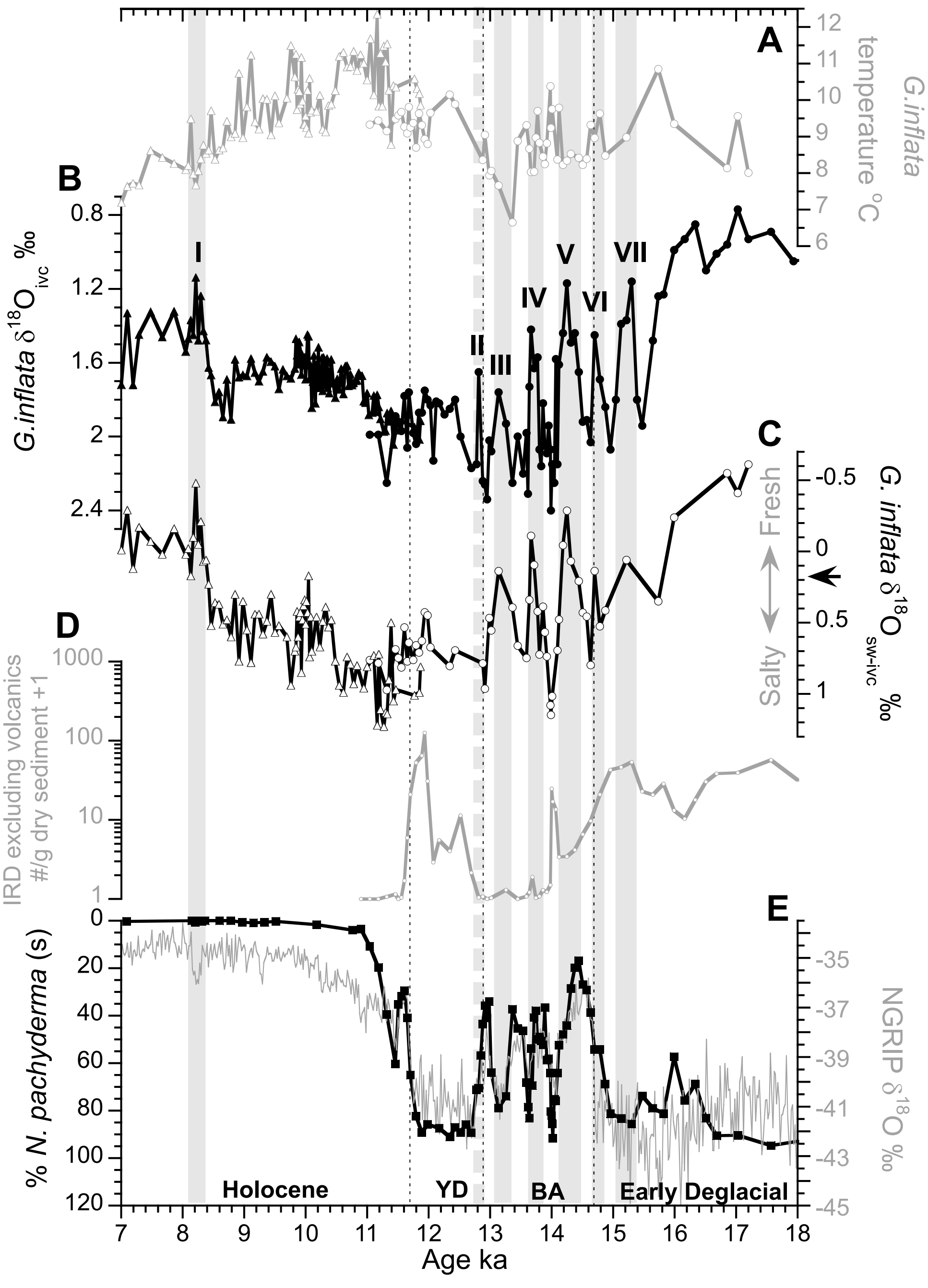


**Figure 6.** (a–e) Proxy records for the South Iceland Rise between 7 and 18 ka. Data from RAPiD-15-4P, circles (11–18 ka) and RAPiD-12-1K [*Thornalley et al.*, 2009], triangles (7–12 ka). Freshwater events indicated by numbering I–VII. Arrow beside axis in Figure 6c indicates core top *G. inflata* $\delta^{18}O_{sw-ivc}$. Figure 6d shows number of IRD grains (150–250 $\mu$m fraction, excluding volcanics) per gram of dry sediment in RAPiD-15-4P. Figure 6e shows NGRIP $\delta^{18}O$ (gray line with no symbols) [*NGRIP members*, 2004]; percent abundance of *N. pachyderma* (black line with solid squares) in RAPiD-15-4P, South Iceland Rise, tied to NGRIP.

imposed upon the high deglacial background *G. inflata* $\delta^{18}O_{ivc}$ ($\delta^{18}O$ ice volume corrected) and $\delta^{18}O_{sw-ivc}$ values (Figure 6), there are six prominent light $\delta^{18}O_{ivc}$ events (five of which are also captured by our slightly lower resolution $\delta^{18}O_{sw-ivc}$ data), indicating brief periods (~100–500 years) of freshening south of Iceland. A previously reported [*Thornalley et al.*, 2009] light $\delta^{18}O_{sw-ivc}$ event also occurs at 8.2 ka, coinciding with the final drainage of Lake Agassiz [*Barber et al.*, 1999; *Ellison et al.*, 2006; *Came et al.*, 2007]. The $\delta^{18}O_{sw-ivc}$ events (labeled I–VII; Figure 6), spaced ~500 years apart, vary in magnitude between 0.4 and 1.4‰. No consistent temperature trends are observed across the events, although several events (I, III and IV) are associated with a transient cooling of ~1.5°C. Similar large amplitude variability is probably not observed in *G. bulloides* $\delta^{18}O_{ivc}$ and $\delta^{18}O_{sw\text{-}ivc}$ because this species is already calcifying within a relatively low salinity near surface layer (with $\delta^{18}O_{sw\text{-}ivc}$ values similar to the minimum values recorded by *G. inflata* during the freshwater events) therefore the episodic input of freshwater, as recorded by *G. inflata*, will cause only minimal change in *G. bulloides* $\delta^{18}O_{ivc}$ and $\delta^{18}O_{sw\text{-}ivc}$ (hinted at by the low-amplitude centennial changes within the *G. bulloides* $\delta^{18}O_{sw\text{-}ivc}$ record).

[21] The presence of large and abrupt centennial timescale changes in *G. inflata* $\delta^{18}O_{ivc}$ and $\delta^{18}O_{sw\text{-}ivc}$ excludes the possibility that *G. inflata* tests were subject to significant bioturbation, preferentially mixing *G. inflata* from abundance maxima associated with warm intervals (low percent *Nps*), to abundance minima associated with cold intervals (high percent *Nps*).

## 4. Discussion

### 4.1. Background Salinity Changes

[22] The fresh conditions recorded in *G. inflata* and *G. bulloides* $\delta^{18}O_{sw\text{-}ivc}$ (Figure 5) during the early deglacial (17–15.8 ka) are likely caused by the influx of icebergs to the North Atlantic during HE1. Following HE1, and persisting throughout the Bølling-Allerød (BA, 12.9–14.7 ka), YD (11.7–12.9 ka), and early Holocene, the heavy *G. inflata* $\delta^{18}O_{sw\text{-}ivc}$ values (~0.5–1.0‰) suggest high background salinity south of Iceland during intervals when Atlantic water was advected to the site. High subsurface salinity south of Iceland during the early Holocene has been proposed to be caused by an increase in the dynamic contribution of saline subtropical gyre (STG) water to the NAC, at the expense of fresh subpolar gyre (SPG) water [*Thornalley et al.*, 2009]. This likely occurred because freshwater fluxes to the Labrador Sea prevented deep convection, thereby reducing baroclinic circulation and weakening SPG circulation, such that it retracted westward [*Thornalley et al.*, 2009]. Consistent with the absence of deep convection in the Labrador Sea prior to ~8 ka [*Hillaire-Marcel et al.*, 2001], we propose that throughout the deglaciation SPG circulation was weak because of high freshwater flux to the Labrador Sea, allowing STG water to dominate the NAC. This will have produced a saline water mass at subthermocline depths in the NE Atlantic when there were significant northward incursions of the NAC. High background NAC salinity following HE1 is also in agreement with the proposed advection of saline subtropical water to high latitudes several hundred years prior to the BA warming [*Carlson et al.*, 2008], and a modeled increased Agulhas leakage of saline Indian Ocean water to the Atlantic during the deglaciation [*Knorr and Lohmann*, 2007], both helping to restart a vigorous AMOC.

### 4.2. Freshwater Events and Their Likely Source

[23] It is proposed that *G. inflata* and *G. bulloides* are calcifying during increased influence of warm Atlantic water to the south of Iceland, and are thus recording the changing properties of the NAC, albeit with *G. bulloides* also being affected by near surface fresh water. Furthermore, it is suggested that the centennial timescale freshwater events recorded by *G. inflata* are caused by episodic increases in freshwater input to the NAC. To support this hypothesis, more local sources of freshwater must be excluded as primary controls on *G. inflata* $\delta^{18}O_{sw\text{-}ivc}$.

[24] Because the *G. inflata* freshwater events do not correlate with maxima in percent *Nps*, and contemporaneous changes in $\delta^{18}O_{sw\text{-}ivc}$ of the near surface dwelling *G. bulloides* are of a much lower amplitude, it is unlikely that the freshwater events are caused by increased influence of near surface, fresh, subpolar and polar waters. Also, since ice-rafted debris (IRD) south of Iceland was predominantly sourced from ice sheets that lay to the north (i.e., the Icelandic and Fennoscandinavian ice sheets) [*Robinson et al.*, 1995], and maxima in IRD do not coincide with the freshwater events I–VII (instead coinciding with cold intervals) (Figure 6), the freshwater events are highly unlikely to have been sourced from these ice sheets. The most plausible scenario is therefore episodic enhanced input of freshwater into the North Atlantic and advection northward via the NAC. This can be confirmed by examining modeling studies and paleoceanographic records (Figure 7).

[25] Model simulations of the drainage chronology of the North American ice sheet [*Tarasov and Peltier*, 2005] indicate significant freshwater discharge from the LIS into the North Atlantic through the southern and eastern drainage routes during the early deglacial. Similarly, planktonic $\delta^{18}O$ data from both the Gulf of Mexico (GOM) [*Aharon*, 2003; *Flower et al.*, 2004; *Aharon*, 2006] and the Laurentian Fan (LF) [*Keigwin et al.*, 2005] (Figure 7) record numerous periods of enhanced freshwater runoff between 15.8 and 12.6 ka. Is it feasible that these freshwater inputs reached the high-latitude North Atlantic? The 8.2 ka freshwater discharge event from Hudson Strait (route D in Figure 1) has been detected along the path of the NAC at numerous locations between 150 and 800 km southwest of Iceland [*Ellison et al.*, 2006; *Came et al.*, 2007; *Thornalley et al.*, 2009], proving that freshwater input from the Hudson Strait can reach the northern North Atlantic. Freshening of the STG during the BA has been reported in lower-resolution studies [*Keigwin et al.*, 1984, 1991; *Schmidt et al.*, 2004; *Carlson et al.*, 2008], although this freshening has been attributed to several other factors. The advection of freshwater from the GOM into the STG and its subsequent transport across the North Atlantic, via the NAC, to south of Iceland has been demonstrated in modeling studies [*Roche et al.*, 2007; *Otto-Bliesner and Brady*, 2009]. Yet during advection across the North Atlantic, freshwater input from the GOM outlet was likely diluted [*Manabe and Stouffer*, 1997; *Tarasov and Peltier*, 2005; *Roche et al.*, 2007], with recent modeling studies suggesting freshwater input via the GOM is approximately half as effective in reducing AMOC in comparison to an input into the subpolar North Atlantic [*Otto-Bliesner and Brady*, 2009]. In summary, there were clearly numerous rerouting events and/or periods of enhanced freshwater input to the North Atlantic between 15.8 and 12.6 ka, which feasibly reached the northern North Atlantic via advection within the NAC, and these freshwater inputs are the most probable source

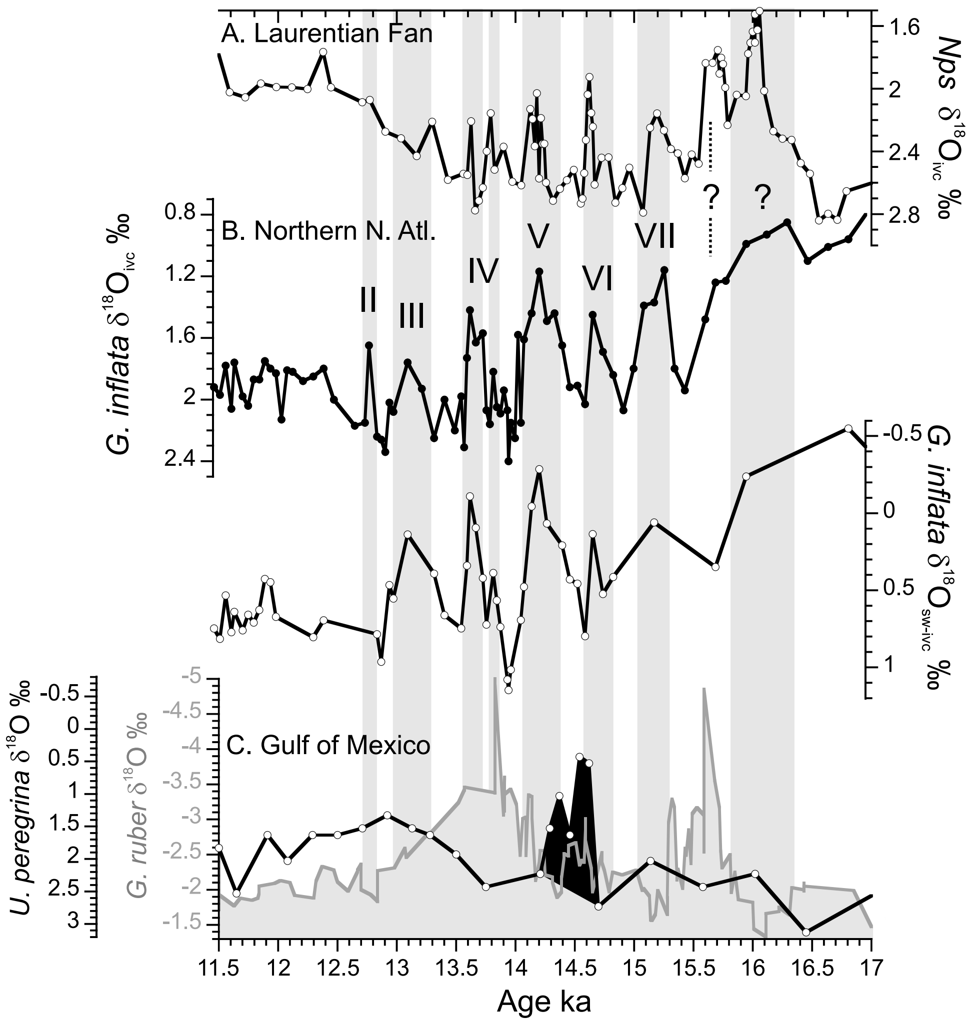


**Figure 7.** Comparison of proxy records of freshwater input to Atlantic between 11.5 and 17 ka. Records plotted on their original timescales, and RAPiD-15-4P records are placed on a before A.D. 1950 age scale, consistent with the other records. (a) Laurentian Fan *Nps* $\delta^{18}O_{ivc}$ from core 14GGC [*Keigwin et al.*, 2005]. (b) *G. inflata* records from RAPiD-15-4P, South Iceland Rise. (c) Planktonic (*G. ruber*, gray line, gray shading, and no symbols) [*Aharon*, 2003] and benthic (*Uvigerina peregrina*, black line with open circles) [*Aharon*, 2006] $\delta^{18}O$ from the Gulf of Mexico; black shading highlights the proposed hyperpycnal discharge of meltwater into the Gulf of Mexico during MWP-1a.

of freshwater to explain the events recorded by *G. inflata* at our site.

### 4.3. Freshwater Events and Climate Feedback During the Deglaciation

[26] In Figure 7, the timing of the freshwater events south of Iceland are compared to published records from the Gulf of Mexico [*Aharon*, 2003, 2006], monitoring the southern LIS meltwater route, and from the Laurentian Fan, which is sensitive to freshwater discharge from the Saint Lawrence River and, to a lesser extent, the more distal Hudson Straits via the Labrador Current [*Keigwin et al.*, 2005]. Despite uncertainties in respective age models (typically between 100 and 200 years), the freshwater events recorded south of Iceland show a striking correlation with freshwater events recorded at the LF. The lower resolution of the RAPiD-15-4P data between 15.5 ka and 16.5 ka may explain why the two events seen in the LF data are not resolved, and similarly why events II and III in RAPiD-15-4P are poorly expressed or absent in the LF data. Moreover, opposing temperature changes during freshwater release may have damped $\delta^{18}O_{ivc}$ variability in the LF data and masked any $\delta^{18}O_{sw}$ changes. In addition, the lower amplitude of the LF $\delta^{18}O_{ivc}$ events in comparison to south of Iceland may be an artifact of *Nps* underestimating near surface salinity changes [*Simstich et al.*, 2003]. Comparing the South Iceland Rise data to the GOM data, the only good correlation is with event IV. Light $\delta^{18}O$ peaks in the GOM planktonic data at 15.6 ka, and benthic data at 14.6 ka occur during intervals of heavy $\delta^{18}O_{ivc}$ south of Iceland. The GOM planktonic data also show a broad freshwater peak between 14.2 ka and 12.8 ka.

[27] Based on this comparison, the freshwater events recorded by *G. inflata* in RAPiD-15-4P are primarily sourced from freshwater discharges from the LIS that were directed out of the eastern routes. However, it is also likely that freshwater discharge out of the GOM and its subsequent northward advection by the NAC, will have contributed to the freshening south of Iceland, although a significant proportion of this freshwater probably did not reach south of Iceland.

[28] *Clark et al.* [2001] proposed that the routing of LIS meltwater to the North Atlantic was controlled by fluctuations in the southern margin of the LIS. During warm intervals, the ice margin retreated and thus opened up the eastern drainage outlets. The increased discharge of freshwater close to sites of deep ocean convection within the North Atlantic was proposed to have caused a reduction in AMOC. The reduction in AMOC led to cooling and a subsequent readvance of the LIS margin, causing meltwater to be rerouted through the southern outlet, allowing AMOC recovery and circum North Atlantic warming. While the LIS margin was located at an intermediate position (43°N–49°N), this feedback mechanism may have caused centennial to millennial timescale fluctuations in climate. Rerouting events have been identified for each of the abrupt centennial-millennial climate reversals of the deglaciation, including the three cold intervals during the BA [*Clark et al.*, 2001].

[29] If the hypothesis of *Clark et al.* [2001] is correct, then the northern North Atlantic should freshen during warm intervals (when retreat of the LIS margin allows meltwater discharge out of the eastern outlets), and become more saline during cold intervals (when advance of the LIS shuts off the eastern drainage route). Furthermore, the onset of cooling should occur during peak fresh conditions in the northern North Atlantic. Our data, illustrated in Figure 8, provide convincing evidence in support of this hypothesis. During warm periods, *G. inflata* $\delta^{18}O_{ivc}$ and $\delta^{18}O_{sw\text{-}ivc}$ values decrease (freshen) south of Iceland, due to enhanced melting of the LIS and retreat of its margin, allowing drainage out of the east. During cold events *G. inflata* $\delta^{18}O_{sw\text{-}ivc}$ values increase (rising salinity), or remain high (saline), suggesting decreasing meltwater input during cold intervals caused by decreased melting and LIS margin advance blocking off the eastern drainage route. Minima in $\delta^{18}O_{ivc}$ and $\delta^{18}O_{sw\text{-}ivc}$ (peak freshening) coincide with the onset of cold intervals, supporting a causal link between increased meltwater input to the northern North Atlantic and climate deterioration, likely via a reduction in AMOC strength and/or enhanced sea ice formation. Freshwater event VII (~15.3 ka) indicates melting and retreat of the LIS occurred prior to the BA, and is consistent with the Hudson River rerouting event (R7 at ~15.5 ka) of *Clark et al.* [2001]. Potentially, this freshwater event may have triggered the decrease in ventilation of the deep North Atlantic at ~15.4 ka recorded by deep sea coral radiocarbon [*Robinson et al.*, 2005] although uncertainty in chronology prevent accurate assessment of phasing.

[30] Our records therefore support the hypothesis of *Clark et al.* [2001] stating that the abrupt climate changes of the deglacial are caused by meltwater rerouting induced by fluctuations in the position of the LIS margin, producing a feedback between climate and meltwater input. Model studies have confirmed that increased freshwater input to the northern North Atlantic caused by rerouting results in AMOC reduction and associated cooling [*Meissner and Clark*, 2006].

### 4.4. Onset of the Younger Dryas

[31] The onset of the Younger Dryas is marked at our site in *G. inflata* $\delta^{18}O_{ivc}$ by a period of freshening at ~12.9 ka that is briefer (~100 years or less) and weaker than those in the BA. (The latter were of course not followed by 1200 year cold intervals.) The low resolution of the $\delta^{18}O_{sw\text{-}ivc}$ data hinders our interpretation of the $\delta^{18}O_{ivc}$ signals, especially regarding the initial freshwater pulse at 12.9 ka, and therefore conclusions are tentative. The data can be interpreted in two ways.

[32] 1. The absence of a strong freshwater signal south of Iceland (briefer and weaker than earlier events) is consistent with the hypothesis that the YD was initially triggered by increased LIS meltwater input to the Arctic, accompanied by only minor meltwater flux to the Atlantic [*Tarasov and Peltier*, 2005]. Supporting evidence for this hypothesis is yet to be found from Arctic records [e.g., *Poore et al.*, 1999; *Hall and Chan*, 2004] although the low resolution of these studies limits their ability to detect centennial timescale events. High-resolution planktonic $\delta^{18}O$ data from the Laptev Sea have instead suggested an early YD flood event from the paleo–Lena River into the Arctic, but the precise timing of the event is poorly constrained [*Spielhagen et al.*, 2005]. Conclusive paleoceanographic data supporting an Arctic freshwater trigger for the YD therefore remains elusive. Furthermore, other mechanisms (aside from the AMOC reduction associated with a brief freshwater input) would also need to be called upon during the remainder of the YD to cause prolonged AMOC weakening and climate cooling [*Stouffer et al.*, 2006].

[33] 2. *Broecker et al.* [1989] proposed that the YD was caused by rerouting of LIS meltwater from the Mississippi to the Saint Lawrence River valley. Reconstructed sea surface salinities from dinocyst assemblage data for the Gulf of Saint Lawrence have been used to refute this hypothesis [*de Vernal et al.*, 1996]. In contrast, light planktonic $\delta^{18}O$ values from the LF data of *Keigwin et al.* [2005] during the YD suggest increased drainage of meltwater down the Saint Lawrence, but without temperature estimates it is difficult to quantify the magnitude of the possible freshening. (Light $\delta^{18}O_{ivc}$ values are also recorded south of Iceland during the YD from 12.4 ka onward, similar to the observed trend in the LF data, although the low-resolution Mg/Ca and $\delta^{18}O_{sw\text{-}ivc}$ data suggest warming rather than freshening at our site.)

[34] Enhanced freshwater fluxes out of the Saint Lawrence during the YD have been reconstructed using various geochemical tracers in planktonic foraminifera [*Carlson et al.*, 2007], with U/Ca and Sr isotope ratios indicating significant freshwater rerouting during the middle and late YD, but only a small increase at the onset of the YD. The $\delta^{18}O_{sw}$ reconstruction of *Carlson et al.* [2007] shows large (~1.8‰) and abrupt freshening at the onset of the YD,

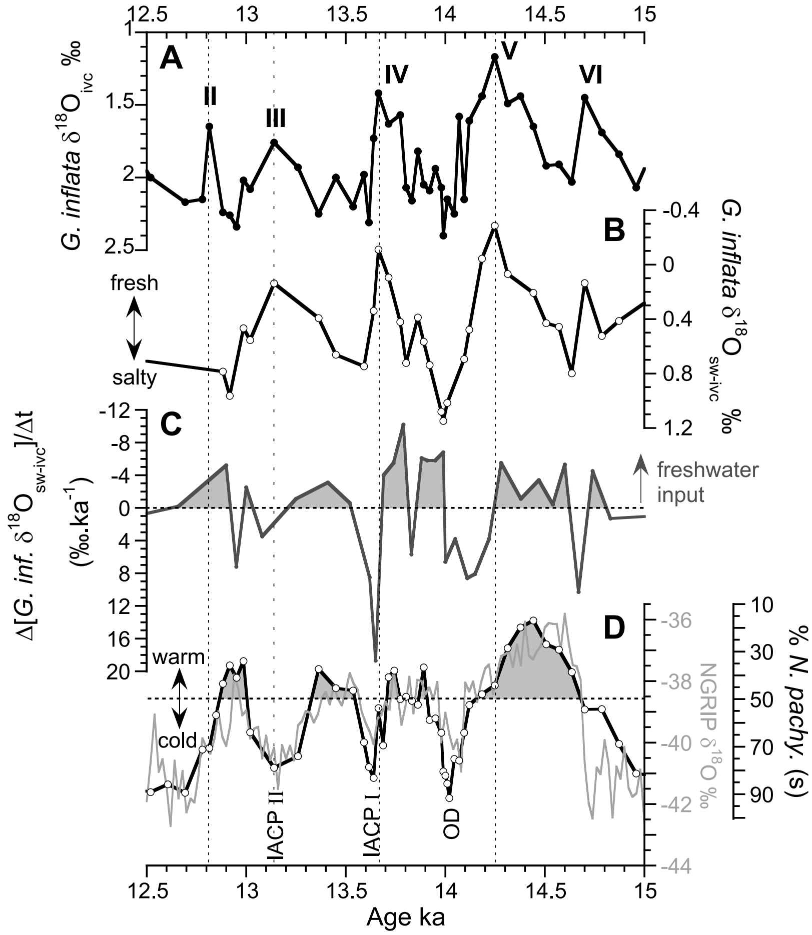


**Figure 8.** Records from the South Iceland Rise (RAPiD-15-4P) illustrating the phasing between meltwater input to the northern North Atlantic and climate records between 12.5 and 15 ka. (a) *G. inflata* $\delta^{18}O_{ivc}$, numbering of fresh water events as in Figure 6. (b) *G. inflata* $\delta^{18}O_{sw\text{-}ivc}$. (c) First differential of *G. inflata* $\delta^{18}O_{sw\text{-}ivc}$ with respect to time (gray line and shading). Gray shading highlights periods of decreasing *G. inflata* $\delta^{18}O_{sw\text{-}ivc}$, i.e., net freshening. (d) Climate (temperature) proxies: NGRIP $\delta^{18}O$ (gray line) [*NGRIP members*, 2004] and percent *Nps* (black line with open circles) in RAPiD-15-4P. Shading highlights warm intervals, where the percent *Nps* is below 50%. During warm intervals $\delta^{18}O_{sw\text{-}ivc}$ decreases (freshening), whereas during cold intervals $\delta^{18}O_{sw\text{-}ivc}$ increases (rising salinity). Dashed lines highlight that minima in $\delta^{18}O_{sw\text{-}ivc}$ (peak freshening) occur at the onset of cold intervals (peak freshening during IACP II (dotted line) may not coincide with the onset of IACP II because of low resolution).

although the reliability of this reconstruction (and the ΔMg/Ca reconstruction) largely depends upon the robustness of using an SST estimate from dinocyst assemblage data to correct for temperature effects on $\delta^{18}O$ (and Mg/Ca) in *G. bulloides*.

[35] Given the lack of meltwater discharge through the Mississippi during the YD, it seems probable that meltwater was rerouted through the Saint Lawrence; however evidence from the Saint Lawrence suggesting that rerouting was the initial trigger for the YD is equivocal. If the brief and relatively weak YD onset freshwater event we observe at our site was caused by this rerouting and was indeed the initial trigger for the YD, it seems likely that the background climate system was preconditioned by earlier meltwater inputs, so that the AMOC was primed for a strong response to a weak trigger [*Lehman and Keigwin*, 1992; *Fanning and Weaver*, 1997; *Knorr and Lohmann*, 2007].

[36] Until further records are produced which provide compelling evidence for an Arctic or Atlantic freshwater trigger, the role that freshwater played in initiating the YD will be debated. Despite uncertainty surrounding the initial trigger and the magnitude of later freshwater fluxes out of the eastern LIS outlets, IRD evidence from the Norwegian Sea [e.g., *Haflidason et al.*, 1995] and south of Iceland (Figure 6) suggests that throughout the middle to late YD the release of icebergs into the Nordic Seas, with associated meltwater and sea ice formation, may have contributed to a weaker AMOC and prolonged the YD until the final switch

to the interglacial state occurred at the onset of the Holocene [*Bakke et al.*, 2009].

### 4.5. Relationship With MWP-1a

[37] MWP-1a has been identified as a ∼20 m rise in sea level in <500 years [*Fairbanks*, 1989; *Edwards et al.*, 1993; *Bard et al.*, 1996; *Hanebuth et al.*, 2000] close to the onset of the BA. The precise timing of this event remains controversial. New dates from Barbados coral records are used by *Stanford et al.* [2006] to argue that MWP-1a occurred between 14.08 ± 0.05 ka and 13.63 ± 0.03 ka (these ages are kiloyears before present (B.P.) A.D. 1950), with the sea level jump being defined by the end of one coral record and the start of another; while the continuous Sunda Shelf mangrove records suggest an earlier timing of 14.6 ka (B.P. A.D. 1950) [*Hanebuth et al.*, 2000], supported by new Tahiti coral data [*Deschamps et al.*, 2009]. Benthic $\delta^{18}O$ data recording a dense hyperpycnal flow of LIS meltwater into the GOM associated with MWP-1a occurs between 14.7 and 14.2 ka (B.P. A.D. 1950) [*Aharon*, 2006], consistent with the earlier timing of MWP-1a.

[38] The source of MWP-1a is also uncertain. Originally the LIS was presumed to have been the most likely source [e.g., *Keigwin et al.*, 1991; *Peltier*, 1994] given its large size. Certain ice sheet models support a large contribution by the LIS to MWP-1a [*Peltier*, 2005; *Tarasov and Peltier*, 2005], and there is clear evidence for meltwater discharge during MWP-1a from the LIS from planktonic foraminifera $\delta^{18}O$ records proximal to major LIS meltwater outlets [*Aharon*, 2003, 2006; *Flower et al.*, 2004; *Keigwin et al.*, 2005]. However, relatively small retreat of the LIS margin during MWP-1a is reconstructed, and steady state ice sheet models suggest only a fraction of meltwater during MWP-1a was derived from the LIS [*Clark et al.*, 1996]. Sea level fingerprinting techniques suggest the primary source of MWP-1a was the Antarctic ice sheet [*Clark et al.*, 2002; *Bassett et al.*, 2005], although the conclusions of these studies have themselves been challenged [*Peltier*, 2005; *Tarasov and Peltier*, 2006]. Employing a freshwater runoff–ocean mixing model, and planktonic foraminifera $\delta^{18}O$ records from close to the major LIS outlets, *Carlson* [2009] has recently estimated that the LIS contributed <30% (or <5.3 m sea level equivalent) toward MWP-1a.

[39] Constraining the timing and source of MWP-1a is important if we are to assess its role within deglacial climate evolution. An early timing and Antarctic source of MWP-1a has been linked to a strengthening of the AMOC and initiation of the BA [*Weaver et al.*, 2003; *Kienast et al.*, 2003], while the later timing suggested by *Stanford et al.* [2006] combined with a significant LIS contribution has been linked to a minor AMOC weakening and the onset of the Older Dryas cold interval at 14.1 ka. Meltwater delivery from the LIS to the Gulf of Mexico during MWP-1a likely occurred as a dense hyperpycnal flow at 14.2–14.7 ka (B.P. A.D. 1950) [*Aharon*, 2006], and although it was suggested this mode of delivery may explain the absence of a strong reduction in AMOC [*Tarasov and Peltier*, 2005], later modeling studies have demonstrated that only approximately <6 m equivalent sea level rise can be delivered by this method without triggering a reduction in AMOC that results in a detectable North Atlantic temperature decrease [*Roche et al.*, 2007].

[40] *G. inflata* records from south of Iceland show numerous freshwater events in the northern North Atlantic with relatively constant amplitudes (based on $\delta^{18}O_{sw\text{-}ivc}$), and no obvious single event corresponds to a dramatic increase in freshwater input associated with MWP-1a. Instead, the *G. inflata* records are primarily recording when meltwater is discharged out of the eastern outlets. Previously, a hyperpycnal discharge of meltwater into the GOM between 14.2 ka and 14.7 ka (B.P. A.D. 1950) has been associated with the LIS contribution to MWP-1a. The largest amplitude event in the *G. inflata* records, event V between 14.05 ka and 14.55 ka (B.P. A.D. 1950), occurs at a similar time to the hyperpycnal flow, consistent with increased melting and an LIS component to MWP-1a during this time interval. The lag between peak discharge into the GOM and peak freshening south of Iceland (Figure 7), is likely caused by the time taken for the LIS margin to retreat and allow discharge of meltwater through the eastern outlets (once this position has been reached, discharge out of the southern route may be expected to decrease, as is observed). Alternatively, the timing of MWP-1a according to *Stanford et al.* [2006] would coincide with the peak in GOM planktonic $\delta^{18}O$ centered at 13.9 ka (B.P. A.D. 1950) and with the slightly smaller event IV at 13.55–13.85 ka (B.P. A.D. 1950) in RAPiD-15-4P. The timing of the hyperpycnal flow into the GOM between 14.7 ka and 14.2 ka (B.P. A.D. 1950) [*Aharon*, 2006] ($\delta^{18}O$ minima at 14.6 ka), followed by peak freshening south of Iceland at 14.2 ka (B.P. A.D. 1950) leads us to favor an early timing for the LIS contribution to MWP-1a, the majority occurring during the hyperpycnal flow between 14.7 ka and 14.2 ka (B.P. A.D. 1950). This increase in LIS meltwater discharge, contributing toward MWP-1a, was likely caused by warming associated with the Bølling interval.

## 5. Conclusions

[41] Using high-resolution planktonic paired Mg/Ca-$\delta^{18}O$ measurements we have reconstructed the $\delta^{18}O_{sw}$ of Atlantic waters advected to the northern North Atlantic throughout the deglaciation, and compared the timing of enhanced freshwater input with abrupt climate change.

[42] Warm Atlantic waters were episodically advected to south of Iceland during the deglaciation. (Note that Atlantic incursions were much less frequent throughout cold intervals when high percent *Nps* abundances suggest a general predominance of polar waters.) Following HE1, the high inferred background salinity of the northward advected Atlantic water is likely caused by saline subtropical waters dominating the NAC at the expense of subpolar water. This was probably caused by weak subpolar gyre circulation, associated with an absence of deep convection in the Labrador Sea, and consistent with findings during the early Holocene. During intervals of advection of warm Atlantic water, near surface waters are fresher than the subsurface water, due to a greater influence of fresh subpolar and polar water.

[43] Superimposed upon the inferred high background subsurface salinity, we observe six freshwater events between 15.8 and 12.6 ka, spaced approximately 500 years apart. The phasing of these events with climate proxies and records of freshwater release from the Saint Lawrence river [*Keigwin et al.*, 2005] provide strong support for the hypothesis of *Clark et al.* [2001]. During warming, the LIS retreats and freshwater is redirected through eastern outlets into the northern North Atlantic. Peak freshening coincides with the onset of abrupt cooling (likely caused by a reduction in AMOC and/or sea ice growth in response to the freshwater input). Throughout the ensuing cold interval, LIS melting reduces and the ice margin advances south, blocking off the eastern outlet and reducing freshwater discharge to the high latitudes. This eventually allows AMOC recovery and climate amelioration, and the feedback loop resumes. *Clark et al.* [2001] have proposed that these fluctuations may occur whenever the LIS margin is located in an intermediate position (43°N–49°N).

[44] Regarding the initial trigger for the onset of the Younger Dryas, we observe only a relatively weak freshwater event at ~12.9 ka. If this was caused by eastward rerouting of LIS meltwater, it seems likely that earlier preconditioning of the climate system is required to explain such a strong AMOC response to a weak trigger. Alternatively, an Arctic freshwater trigger has been proposed for the initial cause of the Younger Dryas, although convincing data in support of this hypothesis has yet to be found.

[45] Controversy surrounds the source and timing of MWP-1a. It is likely that any LIS contribution to MWP-1a was predominantly released through the southern outlet, and therefore our site is poorly located to examine this event. Our records instead primarily record numerous meltwater discharges through the eastern outlets, consistent with enhanced LIS melting during the Bølling-Allerød, but show little or no indication of a single, particularly strong, meltwater discharge event reaching the northern North Atlantic.

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## References


Aharon, P. (2003), Meltwater flooding events in the Gulf of Mexico revisited: Implications for rapid climate changes during the last deglaciation, *Paleoceanography*, *18*(4), 1079, doi:10.1029/2002PA000840.

Aharon, P. (2006), Entrainment of meltwaters in hyperpycnal flows during deglaciation superfloods in the Gulf of Mexico, *Earth Planet. Sci. Lett.*, *241*, 260–270, doi:10.1016/j.epsl.2005.10.034.

Alley, R. B., et al. (2003), Abrupt climate change, *Science*, *299*, 2005–2010, doi:10.1126/science.1081056.

Bakke, J., et al. (2009), Rapid oceanic and atmospheric changes during the Younger Dryas cold period, *Nat. Geosci.*, *2*, 202–205, doi:10.1038/ngeo439.

Barber, D. C., et al. (1999), Forcing of the cold event of 8,200 years ago by catastrophic drainage of Laurentide lakes, *Nature*, *400*, 344–348, doi:10.1038/22504.

Bard, E., et al. (1994), The North Atlantic atmosphere–sea surface $^{14}$C gradient during the Younger Dryas climatic event, *Earth Planet. Sci. Lett.*, *126*, 275–287, doi:10.1016/0012-821X(94)90112-0.

Bard, E., et al. (1996), Deglacial sea-level record from Tahiti corals and the timing of global meltwater discharge, *Nature*, *382*, 241–244, doi:10.1038/382241a0.

Barker, S., M. Greaves, and H. Elderfield (2003), A study of cleaning procedures used for foraminiferal Mg/Ca paleothermometry, *Geochem. Geophys. Geosyst.*, *4*(9), 8407, doi:10.1029/2003GC000559.

Bassett, S. E., G. A. Milne, J. X. Mitrovica, and P. U. Clark (2005), Ice sheet and solid Earth influences on far-field sea-level histories, *Science*, *309*, 925–928, doi:10.1126/science.1111575.

Bé, A. W. H., and D. S. Tolderlund (1971), Distribution and ecology of living planktonic foraminifera in surface waters of the Atlantic and Indian oceans, in *The Micropaleontology of the Oceans*, edited by B. M. Funnell and W. R. Riedel, pp. 105–149, Cambridge Univ. Press, Cambridge, U. K.

Bemis, E. B., H. J. Spero, J. Bijma, and D. W. Lea (1998), Reevaluation of the oxygen isotopic composition of planktonic foraminifera: Experimental results and revised paleotemperature equations, *Paleoceanography*, *13*, 150–160, doi:10.1029/98PA00070.

Bond, G., et al. (1993), Correlations between climate records from North Atlantic sediments and Greenland ice, *Nature*, *365*, 143–147, doi:10.1038/365143a0.

Bondevik, S., et al. (2006), Changes in North Atlantic radiocarbon reservoir ages during the Allerød and Younger Dryas, *Science*, *312*, 1514–1517, doi:10.1126/science.1123300.

Broecker, W. S., J. P. Kennett, B. P. Flower, J. T. Teller, S. Trumbore, G. Bonani, and W. Wolfli (1989), Routing of meltwater from the Laurentide Ice Sheet during the Younger Dryas cold episode, *Nature*, *341*, 318–321, doi:10.1038/341318a0.

Broecker, W. S. (2006), Was the Younger Dryas triggered by a flood?, *Science*, *312*, 1146–1148, doi:10.1126/science.1123253.

Came, R. E., D. W. Oppo, and J. F. McManus (2007), Amplitude and timing of temperature and salinity variability in the subpolar North Atlantic over the past 10 k.y., *Geology*, *35*, 315–318, doi:10.1130/G23455A.1.

Carlson, A. E. (2009), Geochemical constraints on the Laurentide Ice Sheet contribution to meltwater pulse 1A, *Quat. Sci. Rev.*, *28*, 1625–1630, doi:10.1016/j.quascirev.2009.02.011.

Carlson, A. E., et al. (2007), Geochemical proxies of North American freshwater routing during the Younger Dryas cold event, *Proc. Natl. Acad. Sci. U. S. A.*, *104*, 6556–6561, doi:10.1073/pnas.0611313104.

Carlson, A. E., et al. (2008), Subtropical Atlantic salinity variability and Atlantic meridional circulation during the last deglaciation, *Geology*, *36*, 991–994, doi:10.1130/G25080A.1.

Clark, P. U., R. B. Alley, L. D. Keigwin, J. M. Licciardi, S. J. Johnsen, and H. Wang (1996), Origin of the first global meltwater pulse following the Last Glacial Maximum, *Paleoceanography*, *11*, 563–577, doi:10.1029/96PA01419.

Clark, P. U., et al. (2001), Freshwater forcing of abrupt climate change during the last glaciation, *Science*, *293*, 283–287, doi:10.1126/science.1062517.

Clark, P. U., J. X. Mitrovica, G. A. Milne, and M. E. Tamisiea (2002), Sea-level fingerprinting as a direct test for the source of global meltwater pulse 1A, *Science*, *295*, 2438–2441.

Clark, P. U., A. M. McCabe, A. C. Mix, and A. J. Weaver (2004), Rapid rise of sea level 19,000 years ago and its global implications, *Science*, *304*, 1141–1144, doi:10.1126/science.1094449.

Clarke, G. K. C., A. B. G. Bush, and J. W. M. Bush (2009), Freshwater discharge, sediment transport, and modelled climate impacts of the final drainage of Lake Agassiz, *J. Clim.*, *22*, 2161–2180, doi:10.1175/2008JCLI2439.1.

Cléroux, C., E. Cortijo, J.-C. Duplessy, and R. Zahn (2007), Deep-dwelling foraminifera as thermocline temperature recorders, *Geochem. Geophys. Geosyst.*, *8*, Q04N11, doi:10.1029/2006GC001474.

Deschamps, P., et al. (2009), Synchroneity of meltwater pulse 1A and the Bølling onset: New evidence from the IODP "Tahiti Sea-Level" expedition, *Geophys. Res. Abstr.*, *11*, EGU2009-10233.

de Vernal, A., C. Hillaire-Marcel, and G. Bilodeau (1996), Reduced meltwater outflow from the Laurentide ice margin during the Younger Dryas, *Nature*, *381*, 774–777.

Donnelly, J. P., et al. (2005), Catastrophic meltwater discharge down the Hudson Valley: A potential trigger for the intra-Allerød cold period, *Geology*, *33*, 89–92, doi:10.1130/G21043.1.

Edwards, R. L., et al. (1993), A large drop in atmospheric $^{14}C/^{12}C$ and reduced melting in the Younger Dryas, documented with $^{230}Th$ ages of corals, *Science*, *260*, 962–968, doi:10.1126/science.260.5110.962.

Elderfield, H., and G. Ganssen (2000), Past temperature and $\delta^{18}O$ of surface ocean waters inferred from foraminiferal Mg/Ca ratios, *Nature*, *405*, 442–445, doi:10.1038/35013033.

Ellison, C. R. W., M. R. Chapman, and I. R. Hall (2006), Surface and deep ocean interactions during the cold climate event 8,200 years ago, *Science*, *312*, 1929–1932, doi:10.1126/science.1127213.

Fairbanks, R. G. (1989), A 17,000-year glacio-eustatic sea level record: Influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation, *Nature*, *342*, 637–642, doi:10.1038/342637a0.

Fanning, A. F., and A. J. Weaver (1997), Temporal-geographical meltwater influences on the North Atlantic conveyor: Implications for the Younger Dryas, *Paleoceanography*, *12*, 307–320, doi:10.1029/96PA03726.

Flower, B. P., D. W. Hastings, H. W. Hill, and T. M. Quinn (2004), Phasing of deglacial warming and Laurentide Ice Sheet meltwater in the Gulf of Mexico, *Geology*, *32*, 597–600, doi:10.1130/G20604.1.

Ganssen, G. M., and D. Kroon (2000), The isotopic signature of planktonic foraminifera from NE Atlantic surface sediments: Implications for the reconstruction of past oceanic conditions, *J. Geol. Soc. London*, *157*, 693–699.

Haflidason, H., H. P. Sejrup, D. K. Kristensen, and S. Johnsen (1995), Coupled response of the late glacial climatic shifts of northwest Europe reflected in Greenland ice cores: Evidence from the northern North Sea, *Geology*, *23*, 1059–1062, doi:10.1130/0091-7613(1995)023<1059:CROTLG>2.3.CO;2.

Hall, J. M., and L.-H. Chan (2004), Ba/Ca in *Neogloboquadrina pachyderma* as an indicator of deglacial meltwater discharge into the western Arctic Ocean, *Paleoceanography*, *19*, PA1017, doi:10.1029/2003PA000910.

Hanebuth, T., K. Stattegger, and P. M. Grootes (2000), Rapid flooding of the Sunda Shelf: A late glacial sea-level record, *Science*, *288*, 1033–1035, doi:10.1126/science.288.5468.1033.

Hemming, S. R. (2004), Heinrich events: Massive late Pleistocene detritus layers of the North Atlantic and their global climate imprint, *Rev. Geophys.*, *42*, RG1005, doi:10.1029/2003RG000128.

Hillaire-Marcel, C., A. de Vernal, G. Bilodeau, and A. J. Weaver (2001), Absence of deep-water formation in the Labrador Sea during the last interglacial period, *Nature*, *410*, 1073–1077, doi:10.1038/35074059.

Hughen, K. A., et al. (1998), Deglacial changes in ocean circulation from an extended radiocarbon calibration, *Nature*, *391*, 65–68, doi:10.1038/34150.

Hut, G. (1987), Consultants group meeting on stable isotope reference samples for geochemical and hydrological investigations, report to the director general, 42 pp., Int. At. Energy Agency, Vienna.

Keigwin, L. D., B. H. Corliss, E. R. M. Druffel, and E. P. Laine (1984), High resolution isotope study of the latest deglaciation based on Bermuda Rise cores, *Quat. Res.*, *22*, 383–386, doi:10.1016/0033-5894(84)90031-0.

Keigwin, L. D., G. A. Jones, S. J. Lehman, and E. A. Boyle (1991), Deglacial meltwater discharge, North Atlantic deep circulation, and abrupt climate change, *J. Geophys. Res.*, *96*(C9), 16,811–16,826, doi:10.1029/91JC01624.

Keigwin, L. D., J. P. Sachs, Y. Rosenthal, and E. A. Boyle (2005), The 8200 year B.P. event in the slope water system, western subpolar North Atlantic, *Paleoceanography*, *20*, PA2003, doi:10.1029/2004PA001074.

Kienast, M., T. J. J. Hanebuth, C. Pelejero, and S. Steinke (2003), Synchroneity of meltwater pulse 1A and the Bølling warming: New evidence from the South China Sea, *Geology*, *31*, 67–70, doi:10.1130/0091-7613(2003)031<0067:SOMPAT>2.0.CO;2.

Kim, S.-T., and J. R. O'Neil (1997), Equilibrium and nonequilibrium oxygen isotope effects in synthetic carbonates, *Geochim. Cosmochim. Acta*, *61*, 3461–3475, doi:10.1016/S0016-7037(97)00169-5.

Knorr, G., and G. Lohmann (2007), Rapid transitions in the Atlantic thermohaline circulation triggered by global warming and meltwater during the last deglaciation, *Geochem. Geophys. Geosyst.*, *8*, Q12006, doi:10.1029/2007GC001604.

Knutz, P. C., R. Zahn, and I. R. Hall (2007), Centennial-scale variability of the British Ice Sheet: Implications for climate forcing and Atlantic meridional overturning circulation during the last deglaciation, *Paleoceanography*, *22*, PA1207, doi:10.1029/2006PA001298.

LeGrande, A. N., and G. A. Schmidt (2006), Global gridded data set of the oxygen isotopic composition in seawater, *Geophys. Res. Lett.*, *33*, L12604, doi:10.1029/2006GL026011.

LeGrande, A. N., et al. (2006), Consistent simulations of multiple proxy responses to an abrupt climate change event, *Proc. Natl. Acad. Sci. U. S. A.*, *103*, 837–842, doi:10.1073/pnas.0510095103.

Lehman, S. J., and L. D. Keigwin (1992), Sudden changes in North Atlantic circulation during the last deglaciation, *Nature*, *356*, 757–762.

Liu, Z., et al. (2009), Transient simulation of last deglaciation with a new mechanism for Bølling-Allerød warming, *Science*, *325*, 310–314, doi:10.1126/science.1171041.

Manabe, S., and R. J. Stouffer (1997), Coupled ocean-atmosphere model response to freshwater input: Comparison to Younger Dryas event, *Paleoceanography*, *12*, 321–336, doi:10.1029/96PA03932.

McManus, J. F., R. Francois, J. M. Gherardi, L. D. Keigwin, and S. Brown-Leger (2004), Collapse and rapid resumption of Atlantic meridional circulation linked to deglacial climate changes, *Nature*, *428*, 834–837, doi:10.1038/nature02494.

Meissner, K. J., and P. U. Clark (2006), Impact of floods versus routing events on the thermohaline circulation, *Geophys. Res. Lett.*, *33*, L15704, doi:10.1029/2006GL026705.

Mortensen, A. K., M. Bigler, K. Grönvold, J. P. Steffensen, and S. J. Johnsen (2005), Volcanic ash layers from the last glacial termination in the NGRIP ice core, *J. Quat. Sci.*, *20*, 209–219, doi:10.1002/jqs.908.

North Greenland Ice Core Project (NGRIP) members (2004), High-resolution record of Northern Hemisphere climate extending into the last interglacial period, *Nature*, *431*, 147–151, doi:10.1038/nature02805.

Otto-Bliesner, B. L., and E. C. Brady (2009), The sensitivity of the climate response to the magnitude and location of freshwater forcing: Last Glacial Maximum experiments, *Quat. Sci. Rev.*, doi:10.1016/j.quascirev.2009.07.004, in press.

Peck, V. L., et al. (2006), High resolution evidence for linkages between NW European ice sheet instability and Atlantic meridional overturning circulation, *Earth Planet. Sci. Lett.*, *243*, 476–488, doi:10.1016/j.epsl.2005.12.023.

Peck, V. L., I. R. Hall, R. Zahn, and H. Elderfield (2008), Millennial-scale surface and subsurface paleothermometry from the northeast Atlantic, 55–8 ka BP, *Paleoceanography*, *23*, PA3221, doi:10.1029/2008PA001631.

Peltier, W. R. (1994), Ice-age paleotopography, *Science*, *265*, 195–201, doi:10.1126/science.265.5169.195.

Peltier, W. R. (2005), On the hemispheric origins of meltwater pulse 1A, *Quat. Sci. Rev.*, *24*, 1655–1671, doi:10.1016/j.quascirev.2004.06.023.

Poore, R. Z., L. Osterman, W. B. Curry, and R. L. Phillips (1999), Late Pleistocene and Holocene meltwater events in the western Arctic Ocean, *Geology*, *27*, 759–762, doi:10.1130/0091-7613(1999)027<0759:LPAHME>2.3.CO;2.

Rasmussen, S. O., et al. (2006), A new Greenland ice core chronology for the last glacial termination, *J. Geophys. Res.*, *111*, D06102, doi:10.1029/2005JD006079.

Rasmussen, T. L., and E. Thomsen (2004), The role of the North Atlantic drift in the millennial timescale glacial climate fluctuations, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *210*, 101–116, doi:10.1016/j.palaeo.2004.04.005.

Rasmussen, T. L., and E. Thomsen (2008), Warm Atlantic surface water inflow to the Nordic seas 34–10 calibrated ka B.P., *Paleoceanography*, *23*, PA1201, doi:10.1029/2007PA001453.

Rasmussen, T. L., E. Thomsen, L. Labeyrie, and T. C. E. van Weering (1996), Circulation changes in the Faeroe-Shetland Channel correlating with cold events during the last glacial period (58–10 ka), *Geology*, *24*, 937–940, doi:10.1130/0091-7613(1996)024<0937:CCITFS>2.3.CO;2.

Reimer, P. J., et al. (2004), INTCAL04 terrestrial radiocarbon age calibration 0–26 cal kyr BP, *Radiocarbon*, *46*, 1029–1058.

Robinson, L. F., et al. (2005), Radiocarbon variability in the western North Atlantic during the last deglaciation, *Science*, *310*, 1469–1473, doi:10.1126/science.1114832.

Robinson, S. G., M. A. Maslin, and I. N. McCave (1995), Magnetic susceptibility variations in upper Pleistocene deep-sea sediments of the NE Atlantic: Implications for ice rafting and paleocirculation at the Last Glacial Maximum, *Paleoceanography*, *10*, 221–250, doi:10.1029/94PA02683.

Roche, D. M., H. Renssen, S. L. Weber, and H. Goosse (2007), Could meltwater pulses have been sneaked unnoticed into the deep ocean during the last glacial?, *Geophys. Res. Lett.*, *24*, L24708, doi:10.1029/2007GL032064.

Sarnthein, M., et al. (1995), Variations in Atlantic surface ocean paleoceanography, 50°–80°N: A time-slice record of the last 30,000 years, *Paleoceanography*, *10*, 1063–1094, doi:10.1029/95PA01453.

Schmidt, M. W., H. J. Spero, and D. W. Lea (2004), Links between salinity variation in the Caribbean and North Atlantic thermohaline circulation, *Nature*, *428*, 160–163, doi:10.1038/nature02346.

Siani, G., M. Paterne, E. Michel, R. Sulpizio, A. Sbrana, M. Arnold, and G. Haddad (2001), Mediterranean sea surface radiocarbon reservoir age changes since the Last Glacial Maximum, *Science*, *294*, 1917–1920, doi:10.1126/science.1063649.

Simstich, J., M. Sarnthein, and H. Erlenkeuser (2003), Paired $\delta^{18}$O signals of *Neogloboquadrina pachyderma* (s) and *Turborotalita quinqueloba* show thermal stratification structure in Nordic Seas, *Mar. Micropaleontol.*, *48*, 107–125, doi:10.1016/S0377-8398(02)00165-2.

Skinner, L. C., and N. J. Shackleton (2004), Rapid transient changes in northeast Atlantic deep water ventilation age across Termination I, *Paleoceanography*, *19*, PA2005, doi:10.1029/2003PA000983.

Spielhagen, R. F., H. Erlenkeuser, and C. Siegert (2005), History of freshwater runoff across the Laptev Sea (Arctic) during the last deglaciation, *Global Planet. Change*, *48*, 187–207, doi:10.1016/j.gloplacha.2004.12.013.

Stanford, J. D., E. J. Rohling, S. E. Hunter, A. P. Roberts, S. O. Rasmussen, E. Bard, J. McManus, and R. G. Fairbanks (2006), Timing of meltwater pulse 1A and climate responses to meltwater injections, *Paleoceanography*, *21*, PA4103, doi:10.1029/2006PA001340.

Stouffer, R. J., et al. (2006), Investigating the causes of the response of the thermohaline circulation to past and future climate changes, *J. Clim.*, *19*, 1365–1387, doi:10.1175/JCLI3689.1.

Tarasov, L., and W. R. Peltier (2005), Arctic freshwater forcing of the Younger Dryas cold reversal, *Nature*, *435*, 662–665, doi:10.1038/nature03617.

Tarasov, L., and W. R. Peltier (2006), A calibrated deglacial drainage chronology for the North American continent: Evidence of an Arctic trigger for the Younger Dryas, *Quat. Sci. Rev.*, *25*, 659–688, doi:10.1016/j.quascirev.2005.12.006.

Teller, J. T., D. W. Leverington, and J. D. Mann (2002), Freshwater outbursts to the oceans from glacial Lake Agassiz and their role in climate change during the last deglaciation, *Quat. Sci. Rev.*, *21*, 879–887, doi:10.1016/S0277-3791(01)00145-7.

Thornalley, D. J. R. (2008), Palaeoceanography of the South Iceland Rise over the past 21,000 years, Ph.D. thesis, Univ. of Cambridge, Cambridge, U. K.

Thornalley, D. J. R., H. Elderfield, and I. N. McCave (2009), Holocene oscillations in temperature and salinity of the subpolar North Atlantic, *Nature*, *457*, 711–714, doi:10.1038/nature07717.

Voelker, A. H. L., et al. (1998), Correlation of marine $^{14}$C ages from the Nordic Seas with the GISP2 isotope record: Implications for $^{14}$C calibration beyond 25 ka BP, *Radiocarbon*, *40*, 517–534.

Waelbroeck, C., et al. (2001), The timing of the last deglaciation in North Atlantic climate records, *Nature*, *412*, 724–727, doi:10.1038/35089060.

Weaver, A. J., O. A. Saenko, P. U. Clark, and J. X. Mitrovica (2003), Meltwater pulse 1A from Antarctica as a trigger of the Bølling-Allerød warm interval, *Science*, *299*, 1709–1713, doi:10.1126/science.1081002.

Yokoyama, Y., et al. (2000), The timing of the Last Glacial Maximum from observed sea-level minima, *Nature*, *406*, 713–716, doi:10.1038/35021035.

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