Reconstructing North Atlantic deglacial surface hydrography and its link to the Atlantic overturning circulation

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ABSTRACT

Paired Mg/Ca-δ^{18}O measurements on multiple species of planktic foraminifera are combined with published benthic isotope records from south of Iceland in order to assess the role North Atlantic freshwater input played in determining the evolution of hydrography and climate during the last deglaciation. We demonstrate that *Globigerina bulloides* and *Globorotalia inflata* are restricted to intervals when warm Atlantic waters reached the area south of Iceland, and therefore Mg/Ca-δ^{18}O data from these species monitor changes in the temperature and seawater δ^{18}O signature of the northward inflow of Atlantic water to the area. In contrast, *Neogloboquadrina pachyderma* (sinistral) calcifies within local subpolar/polar waters and new Mg/Ca-δ^{18}O analyses on this species document changes in this water mass. We observe two major surface ocean events during Heinrich Stadial 1 (~17-14.7 ka): an early freshening of the Atlantic Inflow (~17-16 ka), and a later interval (16-14.7 ka) of local surface freshening, sea-ice formation and brine rejection that was associated with a further reduction in deep ocean ventilation. Centennial-scale cold intervals during the Bølling-Allerød (BA, 14.7-12.9 ka) were likely triggered by the rerouting of North American continental run-off during ice-sheet retreat. However, the relative effects of these freshwater events on deep ventilation and climate south of Iceland appears to have been modulated by the background climate deterioration. Two freshwater events occurred during the Younger Dryas cold interval (YD, 12.9-11.7 ka), both accompanied by a reduction in deep ventilation south of Iceland: an early YD freshening of the Atlantic Inflow and local subpolar/polar waters, and a late YD ice-rafted detritus event that was possibly related to brine formation south of Iceland. Based on our reconstructions, the strengthening of the Atlantic Meridional Overturning Circulation at the onset of BA and Holocene may have been promoted by

the subsurface warming of subpolar/polar water, brine formation that drew warm saline Atlantic water northwards, and the high background salinity of the Atlantic Inflow.

Keywords: Deglaciation, North Atlantic, melt water, climate, temperature, salinity

1. INTRODUCTION

Earth's climate during the Middle and Late Pleistocene has been dominated by ~100 000 year timescale glacial-interglacial cycles, associated with ice-sheet growth and decay and sea-level change (e.g. Clark et al., 1999; Waelbroeck et al., 2002; Siddall et al., 2003; Clark et al., 2009), fluctuating global temperatures (NGRIP members, 2004; Jouzel et al., 2007) and atmospheric carbon dioxide concentrations (Lüthi et al., 2008 and references therein). A long-standing goal of palaeo-climate research has been to understand the mechanism(s) responsible for governing glacial terminations, which are typified by the transition from the Last Glacial Maximum (LGM, ~26-19 ka, Clark et al., 2009) to the present warm interglacial – the Holocene (~11-0 ka). (In this paper all ages are in ka before 2000 (b2k).)

It is widely accepted that changes in solar insolation (indexed by summer insolation at 65 ^{o}N), due to orbital variability, are the likely driver for glacial terminations. However, at least two observations can be called upon to demonstrate the complexity of glacial terminations, thereby implicating the involvement of other components of the climate system: (1) The last termination (T-I, ~19-7 ka) involved a number of abrupt centennial to millennial timescale climate jumps and reversals (e.g., NGRIP members, 2004), such as the Younger Dryas cold interval (YD, 12.9-11.7 ka), which cannot solely be attributed to a gradual change in solar insolation. (2) Recent work by Cheng et al. (2009) has demonstrated that earlier terminations do not all follow the same sequence of events, e.g. Terminations I and III (T-1, T-III) both contain abrupt warm intervals (the Bølling-Allerød, BA, and its T-III equivalent) bracketed by two cold intervals (the Younger Dryas and the Heinrich Stadial/'Mystery Interval' and their equivalents) whilst Termination II and IV only consist of a single Heinrich Stadial/'Mystery Interval' cold event. A likely cause for this complexity is the interplay between freshwater discharge from decaying ice-sheets and changes in the mode of circulation of the Atlantic meridional overturning circulation (AMOC) (Keigwin et al., 1991; Bjorck et al., 1996; Sarnthein et al., 2000; Clark et al., 2001; Carlson, 2008; Thornalley et al., 2010a; Obbink et al., 2010); although the possible involvement of changing atmospheric circulation patterns (e.g., Seager & Battisti, 2007; Eisenman et al., 2009) and low-latitude processes (e.g., Clement & Peterson, 2008) should not be neglected.

In their examination of the mechanisms of the past four terminations, Cheng et al. (2009), building upon previous studies (e.g., Clark et al., 2004; 2009), propose that rising Northern Hemisphere summer insolation triggered the initial disintegration of a massive, isostatically compensated Laurentide ice-sheet. The discharge of ice and freshwater, occurring as a Heinrich Event (Hemming, 2004), caused a reduction in the AMOC, leading to the ensuing cold Heinrich Stadial interval over the North Atlantic (note, in this study, the term 'Heinrich Event' refers to an ice-sheet collapse and discharge of ice, whereas 'Heinrich Stadial' refers to the entire cold interval – this was not continually occupied by IRD). This North Atlantic cold anomaly caused a southward shift in atmospheric circulation patterns and a build up of heat in the South Atlantic via the bipolar see-saw effect (Stocker & Johnsen, 2003; Barker et al., 2009). Consequently, ventilation and upwelling of CO_2 rich deep waters was enhanced in the Southern Ocean (Toggweiler et al. 2006; Anderson et al., 2009; Barker et al; 2009) and CO_2 was released to the atmosphere, promoting further global warming. Following these initial events, further feedbacks, such as albedo changes and alkalinity-based feedbacks between sea-level and atmospheric CO_2, likely contributed to the full glacial-interglacial transition (Cheng et al., 2009).

The complexity of T-I and T-III, in contrast to the simpler one-step termination observed for T-II and T-IV, has been related to a slower rate of insolation rise (Ruddiman et al., 1980; Oppo et al., 1997; Carlson, 2008; Cheng et al., 2009). During T-I and T-III, the disintegration of circum-North Atlantic ice-sheets and freshwater flux to the North Atlantic was sufficiently slow to allow a transient reinvigoration of the AMOC. However, with climate in an intermediate state and remnant large circum-Atlantic ice-sheets, strong NADW formation was unstable and continued melt-water input eventually caused weakening of the AMOC. The subsequent North Atlantic cooling (e.g., the YD) facilitated the final release of CO_2 to the atmosphere. A further factor determining the finer scale structure of terminations was probably the precise geometry of ice-sheet retreat, which will have controlled the rerouting of freshwater input into the North Atlantic (Clark et al., 2001; Obbink et al., 2010; Thornalley et al., 2010a).

Southern Hemisphere processes may also play an active role in determining the sequence of events during a termination, with variants on a common theme being proposed (e.g. Keeling & Stephens, 2001; Weaver et al., 2003; Rickaby & Elderfield, 2005): Depending upon their respective densities, Antarctic Intermediate Water (AAIW) may compete with North Atlantic Deep Water (NADW) or Glacial North Atlantic Intermediate Water (GNAIW) for occupation of the zone between ~1-2.5 km. When surface water at AAIW formation sites attains a density greater than that found at the deep convection sites in the North Atlantic, NADW formation will decrease, AAIW formation will increase, and vice versa. For example, Weaver et al. (2003) have proposed that an Antarctic origin for meltwater pulse-1a (mwp-1a) may have triggered a reduction in the density of AAIW causing a renewal of NADW formation leading to the onset of the BA.

To further investigate the role that freshwater input to the North Atlantic played in determining the centennial to millennial timescale structure of T-I, we examine material from sediment cores located in the northern North Atlantic on the South Iceland Rise. The northern North Atlantic is a sensitive location for ascertaining past changes in the AMOC and freshwater inputs to the high latitude deep convection sites. In this study we provide a synthesis of the palaeoceanographic changes south of Iceland during T-I, drawing upon previously published data and new paired $\delta^{18}O$-Mg/Ca measurements. We utilise the habitat preferences of different planktic foraminifera species to reconstruct the changing properties of local subpolar/polar water and inflowing warm Atlantic water. Published paired $\delta^{18}O$-Mg/Ca measurements for *Globigerina bulloides* and *Globorotalia inflata* record changes in the warm Atlantic water (Thornalley et al., 2009, 2010a), whilst new paired $\delta^{18}O$-Mg/Ca measurements on *Neogloboquadrina pachyderma* (sinistral) from two cores (located ~100 and 200 km south of Iceland) are used to examine the deglacial changes in local subpolar/polar water. These results are compared to published benthic isotope data from the South Iceland Rise depth transect (Thornalley et al., 2010b).

2. REGIONAL SETTING

The four cores used in this study are located on the South Iceland Rise beneath the surface flow of a major branch of the North Atlantic Current and the deep flow of Iceland-Scotland Overflow Water (Figure 1).

The surface circulation of the northern North Atlantic consists of a warm saline northeastward flow from Newfoundland to the Nordics Seas (the North Atlantic Current, NAC) that draws water from the cold, fresh, subpolar gyre (SPG) and warm, saline, subtropical gyre (STG). The major cold surface current of the Northeast Atlantic is the East Greenland Current (EGC) which flows south around Greenland and into the Labrador Sea. The warm NAC that enters the Nordic Seas (the Atlantic Inflow) cools, sinks and returns southwards as Iceland-Scotland and Denmark Straits Overflow Water (ISOW and DSOW). These overflows entrain adjacent subsurface water masses and after passing around the southern tip of Greenland combine with Labrador Sea Water (LSW, formed by deep convection in the Labrador Sea) to form North Atlantic Deep Water (NADW).

Modern inter-annual variability in the temperature and salinity of the Atlantic Inflow is controlled by subpolar gyre dynamics: enhanced freshwater input to the Labrador Sea causes the SPG to retract westwards allowing warm saline STG water to dominate the Inflow (Hátún et al., 2005). Episodically, the ocean south of Iceland is also influenced by sea-ice and near surface subpolar water that is advected, via the East Icelandic Current (EIC), southwards out of the Nordic Seas, typically when atmospheric circulation is in the negative phase of the North Atlantic Oscillation, such as during the late 1960s (Blindheim & Østerhus, 2005).

During the LGM and cold stadial intervals, sea-ice and subpolar/polar waters extended south of Iceland (to at least 60 ^{o}N (Pflaumann et al., 2003)). However, there were episodic, likely subsurface, incursions of warm Atlantic water (e.g. Rasmussen & Thomsen, 2004). Freshwater input from surrounding ice sheets and extensive sea ice cover severely limited deep open ocean convection in the Nordic Seas (Labeyrie et al., 1992). Instead, there was possible brine formation in the Nordic Seas (Meland et al., 2008) and convection south of Iceland forming Glacial North Atlantic Intermediate Water (GNAIW) ventilating the North Atlantic to ~2 km depth, below which water of southern source filled the ocean basin (Oppo & Lehman 1993; Lynch Stieglitz et al., 2007).

3. MATERIALS AND METHODS

3.1 Age models

We use previously published age models, details of which can be found in Thornalley et al. (2009, 2010a, and 2010b). Because of the variable surface reservoir age of the deglacial northern North Atlantic, deglacial age models are based on tying sharp changes in the percent abundance of the subpolar/polar species, *N. pachyderma* sinistral (*Nps*), with abrupt climate transitions in the NGRIP ice core. Additional robust tie points are provided by Icelandic tephra layers found in both the marine cores and NGRIP. The combination of Holocene radiocarbon dates, and tephra and % *Nps* correlation after ~14.7 ka (the onset of the BA), provides tight age control for the cores. Yet prior to the BA, considerable uncertainty exists. LGM ^{14}C dates use an assumed reservoir age of 800 years, and an additional tie-point is provided at 17.2 ka by correlating benthic $\delta^{18}O$ between the RAPiD cores and the well-dated core SU81-18 (38^{o}N, 10^{o}W) (Waelbroeck et al., 2001), located at a latitude where changes in the surface radiocarbon reservoir age are thought to have been relatively small (Waelbroeck et al., 2001). This assumes that the decrease in benthic $\delta^{18}O$ at the onset of Heinrich Stadial 1 (HS-1, ~14.7 – 17 ka) was approximately synchronous across the Northeast Atlantic. Caution should therefore be used when comparing the timing of climate events recorded by the RAPiD cores between ~15 ka and ~19 ka with existing datasets.

3.2 Core quality

All cores were acquired on Charles Darwin cruise 159 for the RAPID programme of NERC (McCave, 2005). P denotes a piston core and K a kasten (15 cm square gravity) core.

RAPiD-10-1P (62°58.53' N, 17°35.37' W, 1237 m water depth): The proximal location to Iceland (~100 km) provides high sedimentation rates (50-60 $cm.ka^{-1}$) during the LGM, HS-1 and the YD. Lower sedimentation rates occur during the BA and Holocene (12 and 4 $cm.ka^{-1}$ respectively) (Thornalley et al., 2010b).

RAPiD-12-1K (62°05.43' N, 17°49.18' W, 1938 m water depth): High sedimentation rates (~125 $cm.ka^{-1}$) during the early Holocene (~ 8 to 11.7 ka) were caused by the input of tephra and the liberation of sediment during the deglaciation of Iceland. The onset of strong overflows at ~ 8 ka produce a foraminferal sand deposit (~22 $cm.ka^{-1}$). The core is well dated, with 13 radiocarbon dates on *G. bulloides* during the Holocene (Thornalley et al., 2009).

RAPiD-15-4P (62°17.58' N, 17°08.04' W, 2133 m water depth): The core contains an excellent deglacial sequence after ~15.8 ka with sedimentation rates of 12-42 $cm.ka^{-1}$. The effects of bioturbation are minor, with an estimated mixing interval of ~2 cm (Thornalley et al., 2010a). The core contains similar Holocene deposits to RAPiD-12-1K. Recently obtained, unpublished radiocarbon dates on *Nps* suggest that between core depths of 526 and 543 cm (~15.8 to 17.2 ka) the core contains some reworked older material. Because only the most pristine foraminifera were picked for isotope and trace metal analyses this reworked material was successfully avoided, as confirmed by the similar benthic and planktic stable isotope data and planktic Mg/Ca data from nearby cores, which show no signs of reworking (Figure 2 and 3).

RAPiD-17-5P (61°28.90' N, 19°32.16' W, 2303 m water depth): The core contains an excellent Holocene section with high sedimentation rates (~80 cm.ka^{-1}). Lower sedimentation rates occur during the deglaciation: HS-1, 7 cm.ka^{-1}; BA, 9-15 cm.ka^{-1}; YD, 15 cm.ka^{-1} (Thornalley et al., 2010b).

3.3 *N. pachyderma* sinistral stable isotope and Mg/Ca measurements

Paired δ^{18}O-Mg/Ca analysis of planktic foraminifera can be used to reconstruct past ocean temperatures and seawater δ^{18}O ($\delta^{18}O_{sw}$ – a property that shows a near linear relationship with salinity in the modern surface ocean) (Elderfield & Ganssen, 2000).

100-200 tests of the planktic foraminifera *Neogloboquadrina pachyderma* sinistral (*Nps*) were picked from the 150-250 µm fraction. Samples were crushed, homogenised and split to produce ~100-150 µg for stable isotopic analysis, the remainder being used for Mg/Ca analysis. Stable isotope measurements were performed using the Godwin Laboratory VG Prism mass spectrometer attached to a Micromass Multicarb Sample Preparation System. Measurements of δ^{18}O and δ^{13}C were determined relative to the Vienna Peedee belemnite (VPDB) standard, with an analytical precision better than 0.08 ‰ and 0.06 ‰ respectively. Mg/Ca cleaning was conducted following the methods of Barker et al. (2003), screening for contaminating ferromanganese growths, clay minerals and silicate particles. Samples were dissolved and diluted to a [Ca] of 100 ppm and analysed using a Varian Vista inductively coupled plasma optical emission spectrometer (ICP-OES). Analytical precision of Mg/Ca ratios based on replicates of foraminiferal standards is 3%.

3.4 *N. pachyderma* sinistral calibrations

To reconstruct past temperature and $\delta^{18}O_{sw}$ we assume a 9-10 % sensitivity for the Mg/Ca ratio of planktic foraminifera per degree Celsius temperature change (Elderfield & Ganssen, 2000; Anand et al., 2003; Lea, 2003), and equilibrium precipitation of inorganic calcite (Kim & O'Neil, 1997). The temperature effect was then removed from measured calcite $\delta^{18}O$ to yield $\delta^{18}O_{sw}$. These values were then corrected for the whole ocean change in $\delta^{18}O$ of ~1 ‰ that occurred during T-I by scaling to the sea-level curve of Fairbanks et al. (1989) using an LGM to late Holocene sea-level change of 120 m. This produces 'ice-volume corrected' seawater $\delta^{18}O$ ($\delta^{18}O_{sw\text{-}ivc}$).

The Mg/Ca calibrations for the published *G. bulloides* and *G. inflata* data (Thornalley et al., 2009; Thornalley et al., 2010a) use an equation of the form: Mg/Ca = $B \exp(0.1T)$ where T is temperature and the constant B is calibrated to the core tops yielding values of 0.794 and 0.675 for *G. bulloides* and *G. inflata* respectively. Alternative Mg/Ca calibrations for *G. inflata* have been developed, and differences between the calibrations may be caused by a preference for different studies to use a different morphotype (Hathorne et al., 2009). Consistent with the use of a 9-10 % sensitivity of Mg/Ca per degree Celsius temperature, we employ the Mg/Ca calibration of Elderfield & Ganssen (2000) for our *Nps* data: Mg/Ca = 0.52 exp(0.1T). The absence of *Nps* within late Holocene sediments on the South Iceland Rise prevented confirmation of the value of B. The calibration of Elderfield & Ganssen (2000) is utilised rather than the recent calibration by Kozdon et al. (2009a) because the Elderfield & Ganssen (2000) calibration has successfully been applied to a study of late Holocene sediments in the nearby Nordic Seas, with these samples being run in the same laboratory and following the same methods as this study (Nyland et al., 2006). Use of the Kozdon et al. (2009a) calibration would lower the reconstructed temperatures by approximately 0.3-1.5 °C.

It has been widely demonstrated that planktic species do not always precipitate calcite in equilibrium. In accordance with the reconstructed temperatures south of Iceland (7-12 °C), and the culture study work of Bemis et al (1998), we use a $\delta^{18}O$ offset of -0.6‰ for *G. bulloides*. Numerous authors have also observed that post-gametogenetic processes cause *Nps* not to calcify in equilibrium with $\delta^{18}O_{sw}$, and a disequilibrium factor of ~0.6‰ has been widely used (Stangeew, 2001; Simstich et al., 2003; Nyland et al., 2006). Studies from Arctic waters suggest a higher disequilibrium of ~1‰ (Nyland et al., 2006 and references therein) whilst a study from the Icelandic continental shelf suggests a 0.25‰ offset (Smith et al., 2005) and sediment trap data from the Irminger Sea indicate no offset (Jonkers et al., 2010). Kozdon et al. (2009b) have demonstrated that this offset factor can be highly variable due to a ~ 2.1‰ difference in isotopic composition of the ontogenetic calcite and crust calcite. This study therefore urges caution when interpreting down-core $\delta^{18}O_{sw}$ records which may vary independently of temperature or salinity. The replication of down-core trends in $\delta^{18}O_{sw}$ between two cores provides some additional confidence that our reconstructed $\delta^{18}O_{sw}$ data reflect real changes in $\delta^{18}O_{sw}$. In this study we use an offset of 0.6‰, which was successfully used in the Nordic Seas study of Nyland et al. (2006).

4. RESULTS

4.1 Planktic foraminifera habitat preferences

Modern global core top studies show that *G. bulloides* and *G. inflata* become a significant component of the planktic foraminiferal assemblage when summer sea surface temperatures (SSTs) exceed ~7^{o}C (Hilbrecht, 1996). This limitation implies that during the deglaciation, the presence of *G. bulloides* and *G. inflata* will largely be restricted to intervals when relatively warm Atlantic water is advected to the south of Iceland. In contrast, *Nps* abundance increases dramatically when summer SSTs fall below ~7^{o}C (Hilbrecht, 1996), when subpolar/polar waters dominate south of Iceland. Reconstructed Mg/Ca-temperatures for the three species are shown in figure 4 and the data confirm their expected habitat preferences.

The transitional species, *G. inflata* (and to a lesser extent, the warm subpolar species, *G. bulloides*), will only be present in the sediment cores when relatively warm Atlantic water was located over the South Iceland Rise, and paired Mg/Ca-δ^{18}O measurements on these species will reconstruct the changing properties of the Atlantic Inflow. *G. inflata* $\delta^{18}O_{sw-ivc}$ will be particularly sensitive at detecting the advection of Laurentide Ice Sheet (LIS) freshwater input to the NE Atlantic via the NAC (Thornalley et al., 2010a). In contrast, measurements on *Nps* will record the changing properties of local subpolar/polar waters south of Iceland. A further consideration is the depth habitats of the species: *G. bulloides* lives in the near-surface layer (~0-50 m, Ganssen & Kroon, 2000), where it may be influenced by any thin overlying freshwater layer; *G. inflata* lives at the base of the seasonal thermocline (~100-200 m south of Iceland at present)(Ganssen & Kroon, 2000; Cleroux et al., 2007); *Nps* calcifies in the mixed layer and then sinks below the pycnocline to reproduce, where it gains its secondary calcite, resulting in an apparent calcification depth of ~100-200 m (Simstich et al., 2003). Because none of these species are true sea-surface dwellers, surface freshening will be underestimated, a phenomenon that is well documented for *Nps* (Simstich et al., 2003). Furthermore, it has been suggested that *Nps* cannot tolerate salinities <34-34.5 p.s.u and rather than reflecting strong surface freshening, extremely low *Nps* $\delta^{18}O_{sw}$ is a likely indicator of local brine formation (Hillaire-Marcel & deVernal, 2008).

The habitat preference of different planktic foraminifera species underpins modern analogue techniques, which use changes in faunal assemblage to reconstruct past SST change (e.g. Waelbroeck et al., 1998). In the North Atlantic the percent abundance of *Nps* can be used to estimate past mean SSTs: 95% abundance of *Nps* indicates an SST of <4 ^{o}C and when *Nps* abundance drops to <20% an SST of ~9^{o}C is indicated (Bé & Tolderlund, 1971). Down-core % *Nps* data is presented in figure 3.

Combining faunal assemblage data and paired Mg/Ca-δ^{18}O on different species allows a comprehensive reconstruction of the hydrography of the upper ocean. Down-core variations in % *Nps* reflect mean SST changes averaged over the bioturbated sampling intervals (typically 50-200 years), whereas paired Mg/Ca-δ^{18}O measurements record the changing properties of different water masses (the Atlantic Inflow and subpolar/polar water). The two datasets are generally consistent with one another, as can be demonstrated using simple assumptions (i.e., *G. bulloides/G. inflata* Mg/Ca-temperatures record Atlantic Inflow temperatures and *Nps* Mg/Ca-temperatures record subpolar/polar water temperatures, and the proportionate time each water mass is present at the core site is approximated by the % *Nps*). For example, during the early YD, the abundance of *Nps* is ~85%, indicating an SST of ~4.5-5^{o}C, whilst the weighted mean Mg/Ca temperature produces a temperature of 4.6^{o}C (85% x 4^{o}C + 15% x 8^{o}C). Similar calculations can be made for other time intervals, e.g. the early Holocene (>9^{o}C;10.2^{o}C) or the late Bølling (~6^{o}C; 6.5^{o}C).

4.2 Down-core *N. pachyderma* (sinistral) Mg/Ca-d^{18}O

Paired Mg/Ca-δ^{18}O measurements for *Nps* from RAPiD-10-1P and RAPiD-15-4P are presented in figure 3. Mg/Ca ratios indicate an overall warming from the LGM to the Holocene. Mg/Ca ratios increase by 0.1 mmol.mol^{-1} (equivalent to a 1.2^{o}C warming) between 17 and 16 ka, and fluctuations in Mg/Ca and δ^{18}O during the BA from the high resolution core RAPiD-15-4P are in phase with changes in % *Nps*, which correlate with abrupt multi-centennial timescale cooling and warming of Greenland air temperatures (Thornalley et al., 2010a). RAPiD-15-4P *Nps* Mg/Ca ratios then suggest cooling during the early YD and warming during the late YD, continuing into the Holocene. $\delta^{18}O_{sw-ivc}$ displays low values between 16 ka and 14.8 ka and subsequent fluctuations occur throughout the BA.

The results from the two cores are generally consistent with one another (figure 3). A notable difference occurs during Heinrich Stadial 1 (HS-1, and to a lesser extent, the BA), when RAPiD-10-1P records significantly lower $\delta^{18}O_{sw-ivc}$. Based on the low *Nps* $\delta^{18}O_{sw-ivc}$ in RAPiD-10-1P between 14.8-16 ka, there was significant surface freshening south of Iceland and it is likely that active sea-ice formation and brine rejection occurred. The more extreme $\delta^{18}O_{sw-ivc}$ in RAPiD-10-1P presumably reflects the core's proximal location to Iceland where it will be influenced by freshwater from the nearby Icelandic ice-sheet and active coastal sea-ice formation. The suggested brine formation during this interval is consistent with low benthic $\delta^{18}O_{ivc}$ (Figure 5 and Thornalley et al., 2010b).

5. DISCUSSION

5.1 The phasing of surface and deep ocean changes

In figure 5 the phasing between surface and deep ocean records is examined, with the primary aim of assessing the relationship between freshwater input and deep ocean circulation changes. Accurate phase relationships can be inferred throughout the deglacial because the derived surface temperatures and $\delta^{18}O_{sw\text{-}ivc}$ and benthic isotope data have all been obtained from RAPiD-15-4P.

5.1.1 The Last Glacial Maximum and Early Deglacial (17-21 ka)

Benthic $\delta^{18}O_{ivc}$ from the depth transect south of Iceland support the traditional view that GNAIW ventilated the North Atlantic above ~2 km, with southern source water below (Boyle and Keigwin, 1987; Oppo et al., 1993, Lynch-Stieglitz et al., 2007; Thornalley et al., 2010b).

Previous LGM reconstructions (Glacial Atlantic Ocean Mapping Project, GLAMAP, Pflaumann et al., 2003) indicate extensive winter sea-ice cover over the South Iceland Rise, with winter SSTs ~0.5-2^{o}C, and summer SSTs ~3.5-5^{o}C, consistent with our *Nps* Mg/Ca-temperatures of 3-4.5^{o}C and high % *Nps* abundance (figure 3). The relatively saline conditions indicated by our *Nps* $\delta^{18}O_{sw\text{-}ivc}$ data imply that *Nps* was likely calcifying below any near surface freshwater layer that had formed by melting winter sea-ice, analogous to their modern habitat preferences in the Nordic Seas (Simstich et al., 2003). Winter sea-ice coverage likely limited deep convection south of Iceland, although the high inferred *Nps* salinity suggests that in polynyas deep mixing forming GNAIW may have been possible.

Nps $\delta^{18}O_{sw\text{-}ivc}$ does not show any significant freshwater event between 21 ka and 17 ka that can be related to the sea-level rise and melt-water pulse that occurred at ~19 ka, likely sourced from Northern Hemisphere ice-sheets (Clark et al., 2004). An explanation for this discrepancy may either be the low sensitivity of *Nps* $\delta^{18}O_{sw\text{-}ivc}$ at detecting surface freshening, and/or, the freshwater was not advected within subpolar/polar waters to the South Iceland Rise. (Note - *G. inflata* and *G. bulloides*

$\delta^{18}O_{sw-ivc}$, that would monitor changes in Atlantic water, have not been measured due to the low abundance of these species during this interval.)

5.1.2 Heinrich Stadial 1 (14.7-17 ka)

IRD concentration increase in the Nordic Seas and south of Iceland between 17.6 ka and 16 ka, remaining high until the onset of the BA at 14.7 ka, suggests a significant influence of ice-bergs and associated freshwater (figure 5). It is likely that the ~1.5^{o}C warming recorded by *Nps* at ~16-17 ka was associated with this surface freshening, since upper water column stratification will have increased, thereby preventing atmospheric cooling of the subsurface water within which *Nps* gained its secondary calcite. Other studies also suggest the North Atlantic subsurface warmed during HS-1 (e.g. Shaffer et al., 2004; Clark et al., 2007; Knorr & Lohman, 2007; Liu et al., 2009; Alvarez-Solas et al., 2010), although the modelling study by Knorr & Lohman (2007) suggests the cause was increased horizontal mixing with intermediate and upper ocean subtropical water masses that warmed in response to the reduced AMOC – consistent with the warm Atlantic Inflow temperatures recorded during HS-1 south of Iceland.

Northeast (NE) Atlantic IRD data record two distinct peaks during HS-1: a detrital carbonate IRD peak (Heinrich Event 1B, H1B) at ~17 ka (characteristic of the LIS) and a later peak (Heinrich Event 1A, H1A) containing haematite-coated and volcanic grains at 15-16 ka, more characteristic of the ice-sheets surrounding the NE Atlantic (Bard et al., 2000; Peck et al., 2007). (It should be noted that NE Atlantic precursor IRD events prior to the detrital carbonate peak have also been identified (e.g. Bond & Lotti, 1995; Elliot et al., 1998; Peck et al., 2007; Hall et al., 2006).) The detrital carbonate IRD peak is accompanied by low *G. bulloides* and *G. inflata* $\delta^{18}O_{sw-ivc}$ which have been attributed to the advection of LIS-sourced freshwater to south of

Iceland during H1B (Thornalley et al., 2010a). The absence of a strong freshwater signal in the Atlantic Inflow (*G. inflata* $\delta^{18}O_{sw-ivc}$) during the second IRD peak (H1A) is consistent with a NE Atlantic source (rather than LIS source). Further support for a proximal source for H1A is provided by the timing of freshening and inferred sea-ice and brine formation south of Iceland (14.8-16 ka, figure 3).

Benthic isotope data (figure 5) can be used to examine the effect of these freshwater events on the deep ocean south of Iceland. Decreasing $\delta^{13}C$ suggests a reduction in deep ocean ventilation. Decreasing $\delta^{18}O_{sw-ivc}$ can be interpreted as an increasing influence of brine during HS-1 (Thornalley et al., 2010b). However, the inference of brine based on low benthic $\delta^{18}O$ is a matter of ongoing debate (Dokken and Jansen, 1999; Bauch and Bauch, 2001; Meland et al., 2008; Rasmussen and Thomsen, 2009a and 2009b; Thornalley et al., 2010b). Rasmussen and Thomsen (2009a) suggest that in the modern Nordic Seas, brines with a low $\delta^{18}O$ cannot reach intermediate depths because they do not attain a density great enough to sink below the subsurface saline Atlantic layer. Yet our knowledge of the thickness and salinity of this layer during HS-1 is limited. During HS-1, when deep convection decreased due to enhanced surface freshwater input, the Atlantic Inflow was almost certainly reduced and via diffusion the subsurface may have freshened. Combined with intense sea-ice and hence brine formation, low $\delta^{18}O$ brines plausibly reached greater depths than in today's Nordic Seas.

The initial decrease in ventilation south of Iceland at ~17 ka was likely triggered by freshwater discharge from the LIS during H1B and its advection to the NE Atlantic, although earlier European precursor events also began to weaken the AMOC prior to H1B (Hall et al., 2006; Menot et al., 2006; Rinterknecht et al., 2006; Clark et al., 2009). By 15.8 ka, it appears that the source of freshwater had switched from the LIS to the proximal NE Atlantic ice-sheets and deep ventilation was further reduced and intensive brine formation occurred south of Iceland.

5.1.3 The Bølling-Allerød

The onset of the BA was accompanied by an increase in deep ocean ventilation, a decrease in IRD south of Iceland and in the Nordic Seas, and a decrease in the % abundance of *Nps*, all of which suggest an enhanced inflow of warm Atlantic water to south of Iceland and the Nordic Seas, and the resumption of open ocean convection in the northern North Atlantic and strong AMOC, consistent with earlier studies (e.g. Dokken & Jansen, 1999; Sarnthein et al., 2000; Robinson et al., 2005)

Following the initial warming and resumption of strong AMOC at ~14.7 ka, the BA was characterised by a gradual deterioration in climate, with several multi-centennial cold intervals superimposed upon the longer term cooling trend. These cold intervals have been linked to rerouting of freshwater from the LIS as the ice-margin fluctuated during the deglaciation (Clark et al., 2001; Obbink et al., 2010; Thornalley et al., 2010a;). The freshwater supply to the south of Iceland increased during warm intervals as the LIS retreated and freshwater was diverted eastwards, closer to the high latitude deep convection sites. Peak freshening of the Atlantic Inflow triggered a reduction in AMOC and/or a growth in sea-ice that led to the subsequent cold interval. During this ensuing cold interval, the LIS re-advanced and freshwater was diverted southwards, allowing climate amelioration.

Low *Nps* $\delta^{18}O_{sw-ivc}$ during the warm intervals of the BA suggest enhanced melting of circum-NE Atlantic ice-sheets which caused freshening of local subpolar/polar water. Local maxima (saline) in *Nps* $\delta^{18}O_{sw-ivc}$ occur during cold intervals - the Older Dryas (OD, ~14.0 ka) and the Intra-Allerod Cold Interval (IACP, ~13.0-13.4 ka).

The impact of the surface freshwater events can again be assessed by examining benthic isotope data. Benthic $\delta^{13}C$ from south of Iceland indicate a reduction in deep ocean ventilation during the IACP, although a similar signal cannot be clearly detected during the OD: it is possible that any benthic $\delta^{13}C$ shift was masked by other effects, such as air-sea-exchange or local productivity changes.

The end of the BA, just prior to the onset of the YD, is characterised by a brief (200-300 years) recovery out of the IACP: IRD in the Nordic Seas decreased and there was better ventilation of the deep ocean south of Iceland. It should also be noted that the inferred salinity of the Atlantic Inflow had increased to its high, deglacial background level (Thornalley et al., 2010a)

5.1.4 The Younger Dryas

The involvement of freshwater in triggering the Younger Dryas has long been debated and remains unresolved. An early proposal for the cause of the Younger Dryas was the rerouting of LIS meltwater from southern to eastern outlets (e.g. Broecker et al., 1989), but conclusive palaeoceanographic evidence to support this rerouting has remained elusive, with deVernal et al. (1996) finding no evidence for increased freshwater flux through the St. Lawrence. This contrasts with the findings of a later study (Carlson et al., 2007a), although the validity of their freshwater reconstruction methods can be questioned (Peltier et al., 2008). This has led to the suggestion of an Arctic freshwater trigger for the YD (e.g. Tarasov and Peltier, 2005; Murton et al., 2010), but once again, convincing palaeoceanographic evidence of ocean freshening to support the hypothesis is so far absent (Carlson and Clark, 2008). Regardless of the initial trigger, the abrupt increase in IRD in the Nordic Seas suggests that discharge of ice-bergs there played an important role in suppressing AMOC throughout the YD.

The onset of the YD (~12.8 ka) was accompanied by a brief (~100 year) freshwater event south of Iceland recorded by *G. inflata* and *G. bulloides* δ^{18}O (Event II in Thornalley et al., 2010a), although the event was not captured by our slightly lower resolution $\delta^{18}O_{sw\text{-}ivc}$ data. These data therefore suggest there was a small freshwater flux to the NE Atlantic via the NAC. However, low *Nps* $\delta^{18}O_{sw\text{-}ivc}$ persist from the final warm interval of the BA into the early YD, indicating relatively fresh local subpolar/polar water. Since the supply of subpolar/polar waters to south of Iceland largely comes from the Nordic Seas, an Arctic freshwater event cannot be excluded. (Although this reasoning is somewhat tenuous, and the brief freshening of subpolar waters at the onset of the YD may indeed have been cause by the freshening of the Atlantic Inflow.)

Examination of the benthic isotope data elucidates the changing palaeoceanography of the South Iceland Rise during the YD. Deep ventilation reduced from 12.8 ka to 12.6 ka, possibly caused by the YD-onset freshwater event. Between 12.6 ka and ~12.1 ka there appears to have been moderate ventilation of the deep ocean south of Iceland, before a final reduction in ventilation between 12.1 ka and 11.7 ka. This later reduction in ventilation coincided with the presence of brines south of Iceland (Thornalley et al., 2010b). It is uncertain whether these brines formed locally (as possibly suggested by low *Nps* $\delta^{18}O_{sw\text{-}ivc}$) or were exported from the Nordic Seas, as suggested by Meland et al. (2008). The late YD reduction in ventilation coincided with an abrupt increase in IRD south of Iceland, and Heinrich Event 0 (H-0), recorded by cores in the Labrador Sea (e.g., Andrews and Tedesco 1992; Hillaire & deVernal, 2008). The increased presence of surface freshwater south of Iceland would also account for increasing *Nps* temperatures during the late YD (H-0), analogous to during H1 when it is proposed that stratification of the upper water column prevented atmospheric cooling of the subsurface waters inhabited by *Nps*. (Additionally, an overall warming trend in the temperature of Atlantic Inflow waters between ~13.4 ka (the onset of the IACP) and the early Holocene (~11.5 ka) may indicate that during weakened AMOC, the subtropical waters that fed the Atlantic Inflow warmed (Knorr & Lohman, 2007; Carlson et al., 2008a).

5.1.5 The Early Holocene

The onset of the Holocene was associated with an abrupt decrease in % *Nps* and IRD, and an increase in benthic $\delta^{13}C$ suggests better ventilation of the deep ocean. These changes were caused by the retreat of sea-ice, the resumption of open ocean convection in the Nordic Seas and strong AMOC, that brought warm NAC water to the northern North Atlantic (e.g. Dokken & Jansen, 1999; Sarnthein et al., 2000; Robinson et al., 2005). Because of weak SPG circulation, the background salinity of the Atlantic Inflow was still relatively high during the YD and early Holocene and this may have assisted the resumption of strong AMOC at the onset of the Holocene (figure 6 and Thornalley et al., 2010a).

Strong warming during the onset of the Holocene likely caused enhanced melting of the circum-NE Atlantic ice-sheets, as recorded by low *Nps* $\delta^{18}O_{sw\text{-}ivc}$ south of Iceland. Warming south of Iceland was interrupted by the Preboreal Oscillation (PBO), a ~200 year cold interval at ~11.4 ka possibly triggered by a flood from glacial Lake Agassiz into the Arctic Ocean (Fisher et al., 2002; Meissner and Clark, 2006). The PBO is recorded south of Iceland by a decrease in % *Nps* and an inferred increase in salinity of *Nps*, presumably caused by decreased melting of the circum-NE

Atlantic ice sheets. Following the PBO, the continued melting and retreat of the LIS triggered melt-water outbursts from Lake Agassiz (Teller et al., 2002), culminating in the final drainage of Lake Agassiz at ~8.4 ka (Barber et al., 1999; Teller et al., 2002 and references therein), that is detected as a ~0.5‰ decrease (freshening) in the $\delta^{18}O_{sw\text{-}ivc}$ of *G. inflata* (Atlantic Inflow water) south of Iceland. (Earlier events at 9.3 ka and 10.1 ka are possibly also detected south of Iceland.) Via a reduction in AMOC, the release of Lake Agassiz freshwater at 8.4 ka triggered the '8.2 kyr cold event' (Barber et al., 1999; Ellison et al., 2006; Kleiven et al., 2007), and the particularly low benthic $\delta^{13}C$ recorded in RAPiD-15-4P at 8.0 ka to 8.2 ka likely indicate an increased incursion of southern source waters.

Although there was an abrupt increase in the strength of the AMOC at the onset of the Holocene, the modern circulation regime was not reached until ~7-8 ka (Marchitto et al., 1998; Oppo et al., 2003; Piotrowski et al., 2004; Renssen et al., 2009; Thornalley et al., 2010b). This delay was likely a result of the remnant LIS cooling the North Atlantic region via the ice-sheet's high surface albedo, and the continued input of LIS melt water preventing deep convection in the Labrador Sea (Hillaire-Marcel et al., 2001; Carlson et al., 2007b; Carlson et al., 2008b; Renssen et al., 2009). By ~7-8 ka benthic $\delta^{13}C$ south of Iceland had stabilised and reached modern values, ISOW attained its present strength and/or depth on the South Iceland Rise (Thornalley et al., 2010b), and upper ocean circulation resembled the modern regime with strong SPG circulation influencing the Atlantic Inflow (figure 6) and deep winter mixing events homogenising the upper ocean water column (Thornalley et al., 2009).

5.2 The role of circum-North Atlantic freshwater input on the centennial-millennial structure of Termination I.

Having determined the timing of deglacial freshwater inputs and their relationships with changes in deep ocean ventilation south of Iceland (summarised in table 1), it is now possible to examine the broader role of North Atlantic freshwater input in determining the structure of T-I.

5.2.1 The impact of early deglacial freshwater events

Despite the early melt-water events (e.g. the 19 ka mwp and early NE Atlantic IRD events) that were associated with a reduction in AMOC strength and warming of the South Atlantic via the bipolar seesaw, it is not until the onset of H1B at ~17 ka that there was a reduction in the ventilation of the glacial deep ocean (~2 km depth) south of Iceland and a substantial increase in upwelling and CO_2 release from the Southern Ocean (Anderson et al., 2009). It can therefore be inferred that early freshwater inputs were sufficient to weaken AMOC, but further freshwater fluxes to the northern North Atlantic (via the Atlantic Inflow), and extensive sea-ice formation associated with widespread dispersal of freshwater across North Atlantic, were required to strongly perturb North Atlantic climate and trigger significant changes in the Southern Ocean, resulting in an abrupt increase in the release of CO_2 to the atmosphere.

The North Atlantic remained cool and the AMOC weak for at least a thousand years after the end of the deposition of detrital carbonate during H1B (~16-17 ka). Moreover, ventilation of the deep Atlantic was further reduced at ~16 ka (McManus et al., 2004; Gherardi et al., 2005; Robinson et al., 2005; Thornalley et al., 2010b). This later reduction has been attributed to freshwater associated with the second IRD peak identified in the NE Atlantic (H1A, Bard et al., 2000), sourced from proximal ice-sheets that weakened convection in the NE Atlantic (Gherardi et al., 2005), although positive feedback processes associated with expanding sea-ice cover may also have helped intensify climate deterioration and weakening of the AMOC. LIS rerouting events after HS-1, but prior to the BA, may also have contributed towards a weaker AMOC (Clark et al., 2001; Obbink et al., 2010; Thornalley et al., 2010a). Whether the prolonged and intensified reduction of the AMOC during the late HS-1 necessitated a second freshwater injection from circum-NE Atlantic ice sheets, or if it could have been maintained by sea-ice feedbacks alone, bears relevance upon what controlled the timing of the climate amelioration at the end of HS-1, i.e. was the primary control the abrupt cessation of freshwater input, as suggested by Liu et al. (2009)?

5.2.2 The onset of the Bølling-Allerød

Numerous mechanisms have been called upon to explain the sudden strengthening of the AMOC and North Atlantic warming at ~14.7 ka. For example: increased Aghulas leakage of warm saline Indian Ocean water caused by southward migration of Southern Ocean frontal systems in response to Antarctic sea-ice retreat (Knorr & Lohman, 2003), freshening of AAIW caused by mwp-1a or retreat of Antarctic sea-ice (Keeling & Stephens, 2001; Weaver et al., 2003), northward advection of salinity that had accumulated in the tropics during the preceding interval of weakened AMOC (e.g., Schmidt et al., 2004; Carlson et al., 2008a), or, an abrupt cessation or rerouting of freshwater input to the North Atlantic (Clark et al., 2001; Liu et al., 2009; Obbink et al., 2010). Several observations from the south of Iceland shed light upon possible mechanisms responsible for the onset of the BA:

1. There was a gradual increase in the background salinity of the Atlantic Inflow towards the onset of the BA, suggesting an increased salinity flux to the high latitude North Atlantic. This flux would have then increased dramatically during the abrupt resumption of strong AMOC because of the greater volume of Atlantic Water that was drawn northwards (decreasing % *Nps* south of Iceland).
2. Subsurface warming of local subpolar/polar waters, caused by surface ocean stratification limiting atmospheric cooling of subsurface layers, and mixing with warmer waters from the south, may have destabilised the water column and triggered overturning, as suggested by modelling studies (e.g. Knorr & Lohman, 2007).
3. Active sea-ice formation and the sinking of locally formed brines south of Iceland between 16 ka and 14.8 ka (in addition to brine formation in the Nordic Seas) will have drawn warm, saline Atlantic water northwards, increasing local surface ocean salinity and triggering open ocean overturning south of Iceland and in the Nordic Seas (Dokken & Jansen, 1999).
4. The abrupt strengthening of the AMOC at the onset of the BA coincides with the end of active sea-ice formation south of Iceland, and the two processes are probably coupled via a positive feedback (increasing AMOC causes sea-ice retreat that allows further penetration of the Atlantic Inflow and invigoration of the AMOC).

In order to simulate the abrupt warming of the BA, Liu et al. (2009) require the continued input of freshwater to the North Atlantic (with a flux sufficient to maintain the weakened AMOC state) until only a few centuries prior to the onset of warming. We speculate that the necessary freshwater input was via H1A (~15-16 ka) and the subsequent sea-ice formation in the NE Atlantic kept AMOC suppressed for several further centuries, in a manner analogous to the prolonged sea-ice formation in the Labrador Sea following the end of IRD deposition associated with H1B (Hillaire-Marcel & deVernal, 2008). However, once the freshwater input had ended, it was only a matter of time before the combined effects of subsurface warming and an increasing salinity flux to the high latitudes overcame the effects of enhanced sea-ice cover, leading to the reinvigoration of the AMOC, which coupled via a positive feedback, caused rapid sea-ice retreat. Increasing atmospheric CO_2 levels may also have helped melt northern North Atlantic sea-ice (Liu et al., 2009).

5.2.3 Climate deterioration throughout the Bølling-Allerød and the onset of the Younger Dryas

To understand the relationship between North Atlantic freshwater input and climate during the BA, it is instructive to consider the BA as a period of gradual climate deterioration, accompanied by a reduction in deep ocean ventilation, upon which centennial timescale cold intervals and sharp decreases in deep ocean ventilation are superimposed. The centennial timescale cold intervals of the BA (i.e., the OD at ~14.0 ka, a small event at ~13.6 ka, and the IACP at 13.4-13.0 ka) appear to have been triggered by eastern rerouting of freshwater caused by the retreat of the LIS during the preceding warm interval (Thornalley et al., 2010a). Based on the excellent correlation between Nordic Seas IRD concentration and Greenland temperature, circum-Nordic Seas ice-sheets presumably helped to amplify and/or sustain the cooling in the Nordic Seas following the initial freshwater trigger delivered by the Atlantic Inflow freshening.

What then determines the severity of the cold intervals and the ventilation of the deep ocean south of Iceland during the BA? The magnitude of freshening of the Atlantic Inflow is not the only consideration because the most prominent freshening of the Atlantic Inflow during the BA, at ~14.2 ka, does not result in the coldest conditions and poorest deep ocean ventilation of the BA – these events instead occur during the IACP (13.0-13.4 ka). Knorr & Lohmann (2007) have hypothesised that the extreme warmth of the Bølling was terminated by mwp-1a (or as previously reported in Thornalley et al. (2010a), the eastward rerouting of the LIS contribution to mwp-1a at ~14.2 ka), and the subsequent climate system shift towards cooler conditions may have preconditioned the AMOC to be more sensitive to subsequent freshwater inputs. Evidence from close to the LIS eastern outlets suggests background freshwater discharge increased through these outlets during the BA (with rerouting events superimposed), contributing to a gradually weakening AMOC (Obbink et al. 2010). (Note - this background increase in freshwater discharge was apparently not advected, via the NAC, as far north as the South Iceland Rise because no similar trend is observed through the BA in our *G. inflata* $\delta^{18}O_{sw\text{-}ivc}$.)

Alternatively, Liu et al. (2009) have argued that the extreme warmth of the Bølling was a transient event associated with an AMOC overshoot caused by basin-wide salinity adjustments and the preceding North Atlantic subsurface warming, and subsequent cooling may be an intrinsic relaxation of the climate system. Climate cooling throughout the BA may also have facilitated increased ice-berg activity in the Nordic Seas during centennial-scale cold intervals triggered by freshwater events in the Atlantic Inflow. Considering Southern Ocean processes, the increasing sensitivity of the AMOC to North Atlantic freshwater input through the BA may also have been caused by a gradual increase in the salinity of AAIW, so that the density of AAIW approached that of NADW, and any small freshwater input to the North Atlantic caused NADW to be less dense than AAIW, resulting in a strong reduction in NADW formation (Keeling & Stephens, 2001).

If the YD was indeed triggered by a freshwater input to the North Atlantic, the preconditioning and gradual deterioration of the background climate throughout the BA may explain why only a small freshwater input is detected south of Iceland at the onset of the YD, sea-level shows little change (Bard et al., 2010), yet there is an ensuing ~1200 year cold interval (Thornalley et al., 2010a). Of course, alternative mechanisms for triggering the YD have been hypothesised (e.g., Tarasov & Peltier, 2005; Carlson et al., 2007a; Seager & Battisti, 2007; Clement & Peterson, 2008; Eisenman et al., 2009).

6. CONCLUSIONS

Using multi-species planktic paired Mg/Ca-δ^{18}O measurements, the hydrography of the Atlantic Inflow and local subpolar/polar water south of Iceland throughout the last deglaciation has been reconstructed and compared to changes in deep ocean circulation to elucidate the role freshwater input played in the climate evolution of Termination I.

1. The strong reduction of the AMOC during the onset of HS-1 was presumably related to the most extreme deglacial freshening of the Atlantic Inflow, caused by a collapse of the LIS at ~16-17 ka. There was a further decrease in deep ocean ventilation south of Iceland, coinciding with surface freshening and possible sea-ice formation south of Iceland, that was likely caused by a later collapse of a circum-NE Atlantic ice sheet (~15-16 ka).

2. Centennial-timescale cold intervals during the BA were caused by fluctuations in the ice-margin of the LIS, which rerouted freshwater from southern to eastern North American outlets. Iceberg discharge from circum-Nordic Seas ice-sheets may have amplified or sustained cooling.
3. The gradual climate deterioration throughout the BA and into the YD, combined with the absence of a strong freshwater signal south of Iceland at the onset of the YD, suggests the YD may have been an intrinsic feature of the deglaciation and only a relatively weak trigger was required once background climate was preconditioned.
4. During the late HS-1 and YD, subsurface warming of local subpolar/polar waters, brine formation in the northern North Atlantic, and a salty Atlantic Inflow may all have helped promote the resumption of strong AMOC at the onset of the BA and Holocene. The strengthening of AMOC during these transitions is coupled with the retreat of sea-ice and a sharp increase in the predominance of warm, saline Atlantic Inflow waters south of Iceland.
5. Freshwater input from the remnant LIS into the Labrador Sea likely caused weak subpolar gyre circulation, resulting in a relatively warm and saline Atlantic Inflow south of Iceland until ~8 ka. ISOW only attained its modern strength and/or depth over the South Iceland Rise, and its characteristic, stable, well-ventilated $\delta^{13}C$ signature by ~7-8 ka, possibly related to the upper ocean reorganisation.

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9. FIGURE LEGENDS, TABLES

Fig. 1 Modern northern North Atlantic ocean circulation and the location of the study area (black star). Grey arrows indicate deep ocean currents; solid black arrows, warm surface currents; dashed black arrows, cold surface currents. SPG, subpolar gyre; NAC, North Atlantic Current; EGC, East Greenland Current; EIC, East Icelandic Current.

Fig. 2. Oxygen isotope data from cores RAPiD-15-4P (grey) and RAPiD-17-5P (black). (**a**) *G. bulloides* (**b**) *C. wuellerstorfi* (Thornalley et al., 2010b).

Fig. 3. Paired δ^{18}O-Mg/Ca data for *N. pachyderma* (s) from RAPiD-10-1P (grey) and RAPiD-15-4P (black). (**a**) % abundance of *Nps*. (**b**) *Nps* δ^{18}O, Thornalley et al., 2010b (**c**) *Nps* Mg/Ca ratios (**d**) 'Ice-volume corrected' seawater δ^{18}O.

Fig. 4. Bar chart illustrating the range of reconstructed Mg/Ca-temperatures for the different planktic species analysed.

Fig. 5. Comparison of surface and deep ocean records. Light grey bars, Greenland cold intervals; dark grey bars, Heinrich events H-0 and H1. (**a**) Records of ice-rafted debris (IRD) from south of Iceland (black line, 150-250 µm fraction, excluding volcanic grains, from RAPiD-17-5P, Thornalley et al., 2010b) and the Nordic Seas (grey line, ratio of lithics:(lithics+foraminifera), from core MD99-2284, Bakke et al., 2009). (**b**) Derived temperatures and 'ice-volume corrected' seawater δ^{18}O from planktic paired Mg/Ca-δ^{18}O data. Data from 7-11.4 ka is from RAPiD-12-1K (Thornalley et al., 2009), data from 11.4-18 ka is from RAPiD-15-4P (Thornalley et al., 2010a). *G. bulloides*, open symbols, *G. inflata*, filled diamonds. (**c**) Derived temperatures and 'ice-volume corrected' seawater δ^{18}O from *Nps* paired Mg/Ca-δ^{18}O data. All data is from RAPiD-15-4P. (**d**) Benthic δ^{13}C and 'ice-volume corrected'

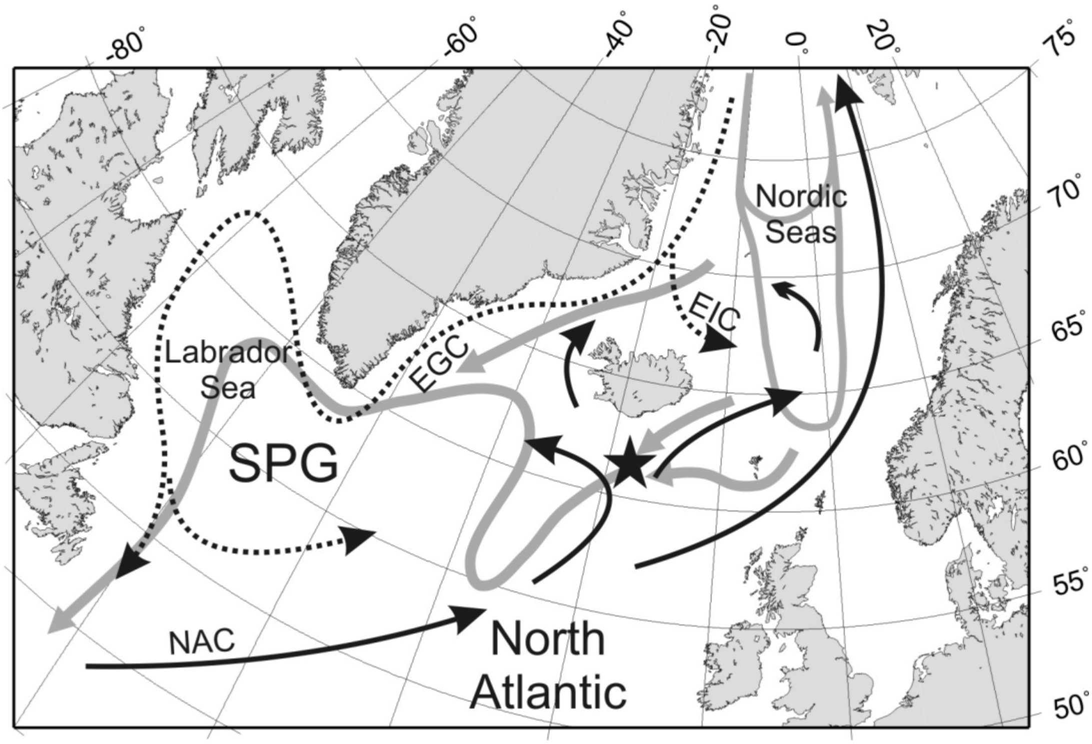
δ^{18}O for *C. wuellerstorfi* (>212 μm) from RAPiD-15-4P (2133 m), Thornalley et al., 2010b. (**e**) NGRIP δ^{18}O (NGRIP members, 2004).

Fig. 6. Proxy records indicating changes in the strength of the subpolar gyre (SPG) – increasing SPG strength downwards. (**a**) 'Ice-volume corrected' seawater δ^{18}O from RAPID-12-1K (0-11.4 ka) and RAPiD-15-4P (11.4-18 ka), inferred salinity increases upwards. (**b**) Labrador Sea salinity, based on dinocyst assemblage data from core P-013 (Solignac et al., 2004) – salinity increases downwards. (**c**) Linear detrended temperature anomalies based on planktic foraminiferal assemblages, from the eastern subtropical Atlantic, ODP site 658C (deMenocal et al., 2000).

Table 1. Synthesis of deglacial palaeoceanography south of Iceland. (f.w., freshwater; SSW, Southern Source Water).

<table>
<tr><th rowspan="2"></th><th colspan="5">Events south of Iceland</th><th rowspan="2">Controls elsewhere</th></tr>
<tr><th>% *Nps* - (sub)polar water</th><th colspan="2">Atlantic Inflow</th><th>Local subpolar/polar water</th><th>Deep ocean (~2 km)</th></tr>
<tr><td>**Mid-Late Holo.** (0-9 ka)</td><td><1%</td><td colspan="2">Oscillating *TS* changes – SPG control
Millennial surface f.w. events</td><td>-</td><td>Modern, stable ISOW $\delta^{13}C$ and flow strength/depth by ~7-8 ka</td><td>Oscillating SPG strength.
Atmospheric circulation changes.</td></tr>
<tr><td>**8.2 ka event**</td><td><1%</td><td rowspan="7">High background salinity caused by weak SPG</td><td>0.5 psu f.w. event at 8.4-8.2 ka</td><td>-</td><td>Poor ventilation ($\delta^{13}C$ ~0.5‰)</td><td>Drainage of Lake Agassiz</td></tr>
<tr><td>**Preboreal** (~9-11.7 ka)</td><td>60% to <1%
(50% in PBO)</td><td>Warm and salty, gradually decreasing.</td><td>Warm.
f.w. input from local melting
Salty during cold interval</td><td>Variable $\delta^{13}C$ – preformed or entrained signal, or increased SSW.
Brines ~10.5-11.7 ka</td><td>f.w. event at 11.3 ka – LIS source via Arctic (?)</td></tr>
<tr><td>**YD** (11.7-12.9 ka)</td><td>85-90%</td><td>Warm and salty, 100-yr f.w. event at onset</td><td>Warming through YD
f.w. or brine formation with high IRD at 12.1-11.7 ka</td><td>Poor ventilation at 12.8-12.6 ka and 12.1-11.7 ka.
Brines at 12.1-11.7 ka.</td><td>LIS f.w. event at onset (?)
H-0 ~12.1-11.7 ka.
Abundant IRD in Nordic Seas.</td></tr>
<tr><td>**BA** (12.9-14.7 ka)</td><td>Fluctuating 20-80%</td><td>f.w. events caused by LIS rerouting, trigger cold intervals</td><td>Warmer and fresher (enhanced melting) during climate warm intervals</td><td>Well ventilated, decreasing through BA.
Poorer ventilation in cold intervals.</td><td>Open ocean convection
Nordic Seas IRD in cold intervals
Fluctuating LIS margin</td></tr>
<tr><td>**HS-1** (14.7-17 ka)</td><td>85-95%</td><td>Warm.
f.w. from LIS 17.5-16 ka</td><td>Sub-surface warming to $5^{o}C$.
Active sea-ice formation from 16-14.8 ka.</td><td>Decreasing ventilation.
Increasing brine influence</td><td>H1B (~17 ka) – LIS IRD
H1A (~16 ka) – NE Atlantic IRD
Heat/salinity retention in (sub)tropics</td></tr>
<tr><td>**LGM and early deglacial** (17-21 ka)</td><td>85-95%</td><td>-</td><td>Cold (3-4.5 ^{o}C). No signal of f.w. events or sea-ice formation</td><td>GNAIW and GAABW, boundary at 2-2.5 km.</td><td>Nordic Seas sea-ice and ice-bergs exported to south of Iceland.
19 ka melt-water pulse</td></tr>
</table>

Fig. 1



-80°
-60°
-40°
-20°
0°
20°
75°
70°
65°
60°
55°
50°
Nordic
Seas
EIC
EGC
Labrador
Sea
SPG
NAC
North
Atlantic

Fig.2

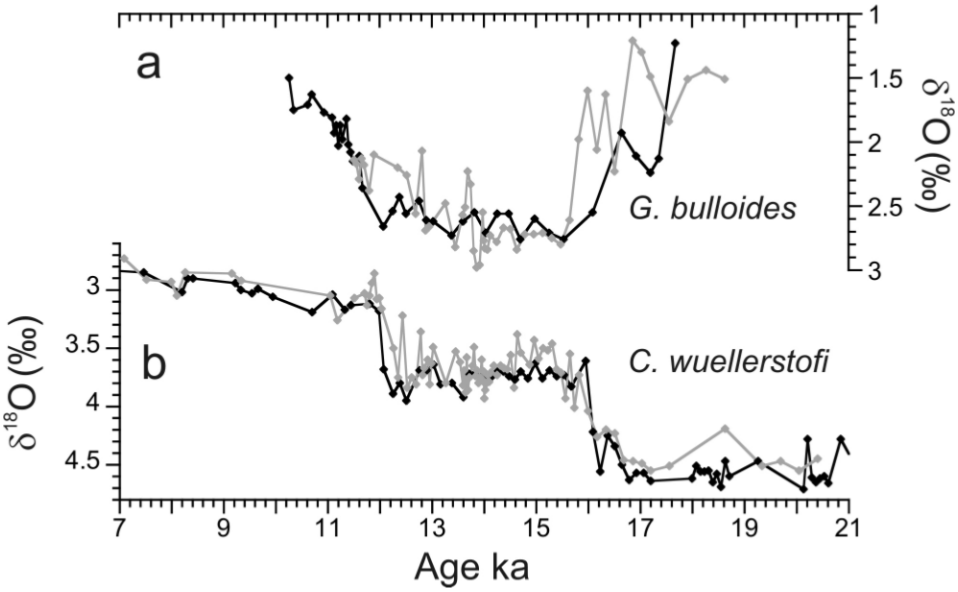

a
b
G. bulloides
C. wuellerstofi
δ18O (‰)
δ18O (‰)
1
1.5
2
2.5
3
3
3.5
4
4.5
7
9
11
13
15
17
19
21
Age ka

Fig. 3

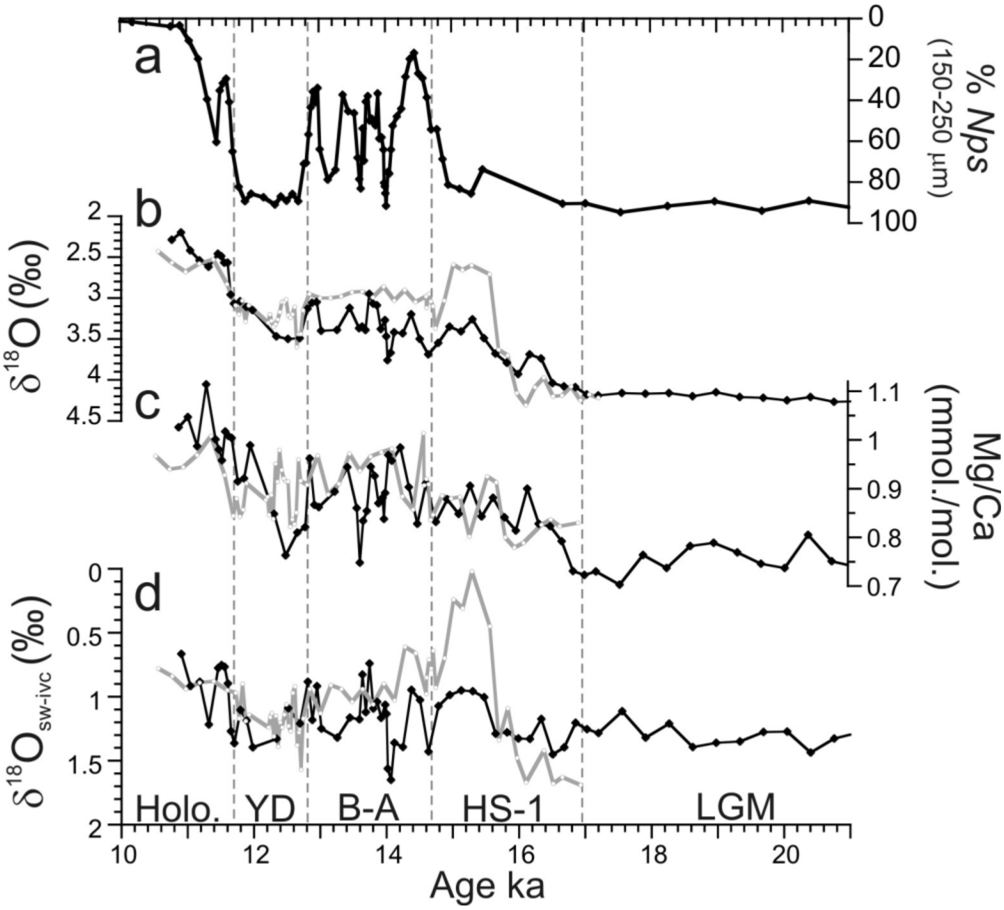
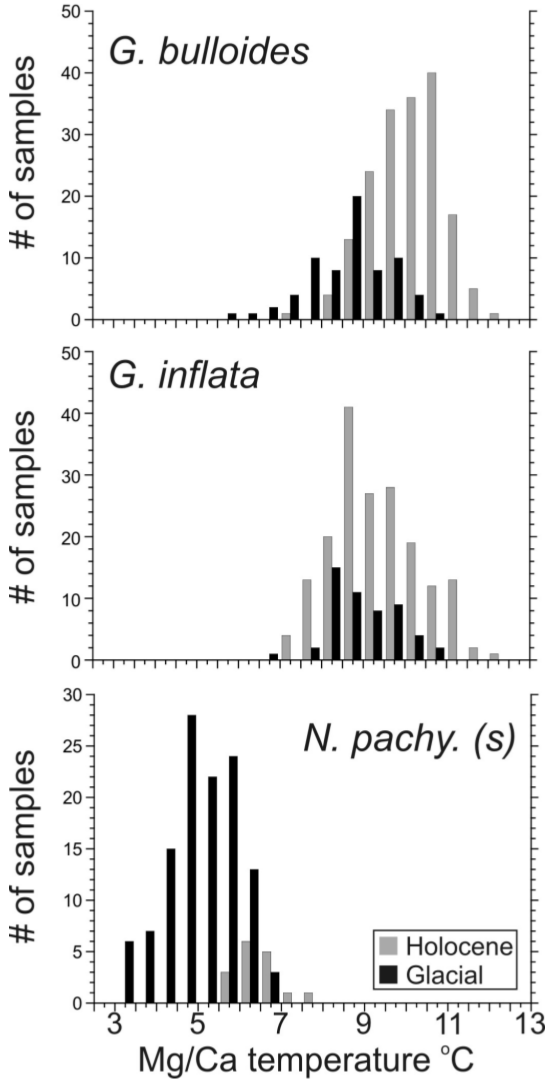
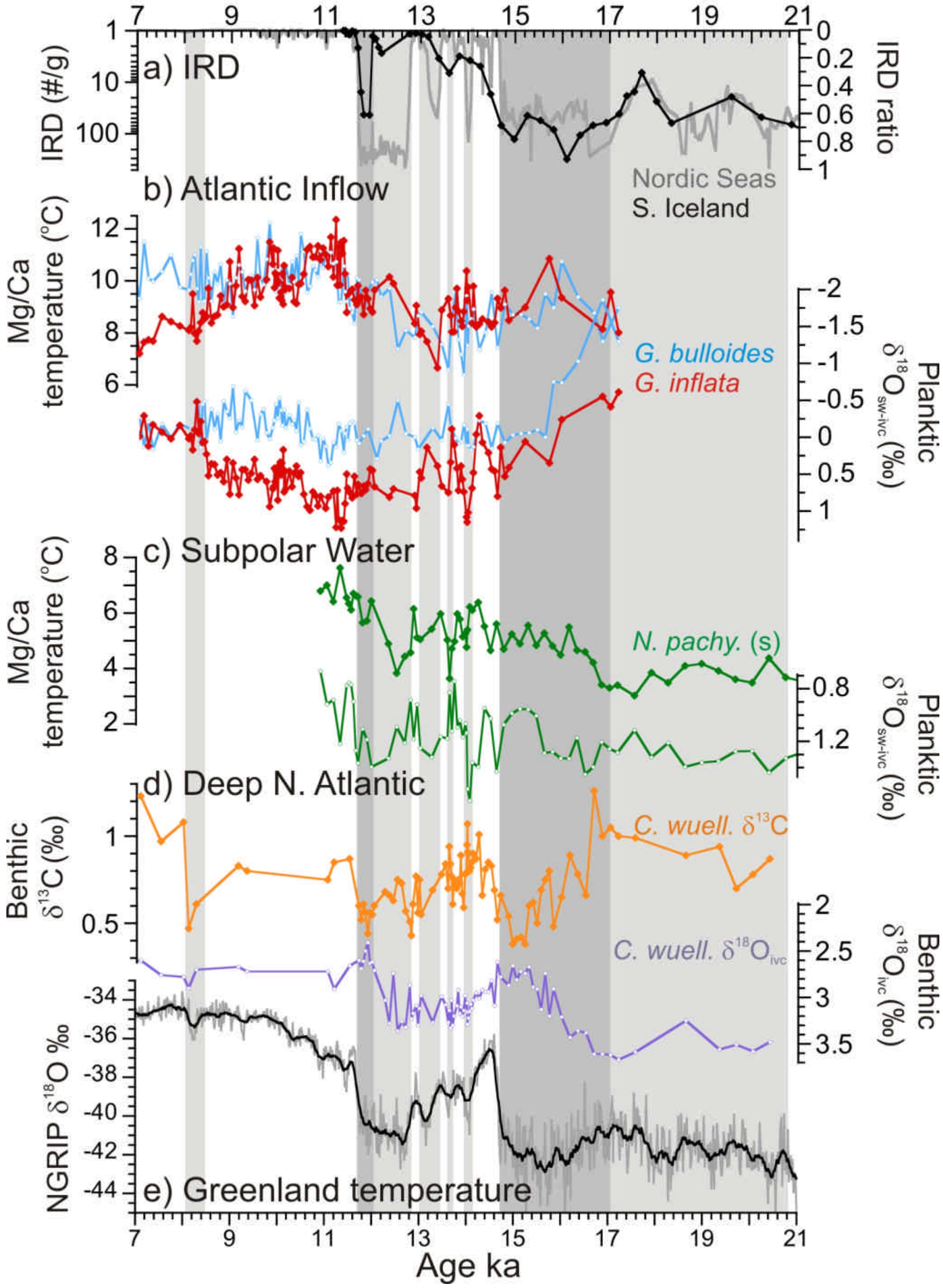

a
b
c
d
% Nps
(150-250 μm)
δ18O (‰)
Mg/Ca
(mmol./mol.)
δ18Osw-ivc (‰)
Holo.
YD
B-A
HS-1
LGM
Age ka
0
20
40
60
80
100
2
2.5
3
3.5
4
4.5
1.1
1
0.9
0.8
0.7
0
0.5
1
1.5
2
10
12
14
16
18
20

Fig. 4


G. bulloides
G. inflata
N. pachy. (s)
# of samples
Holocene
Glacial
Mg/Ca temperature °C

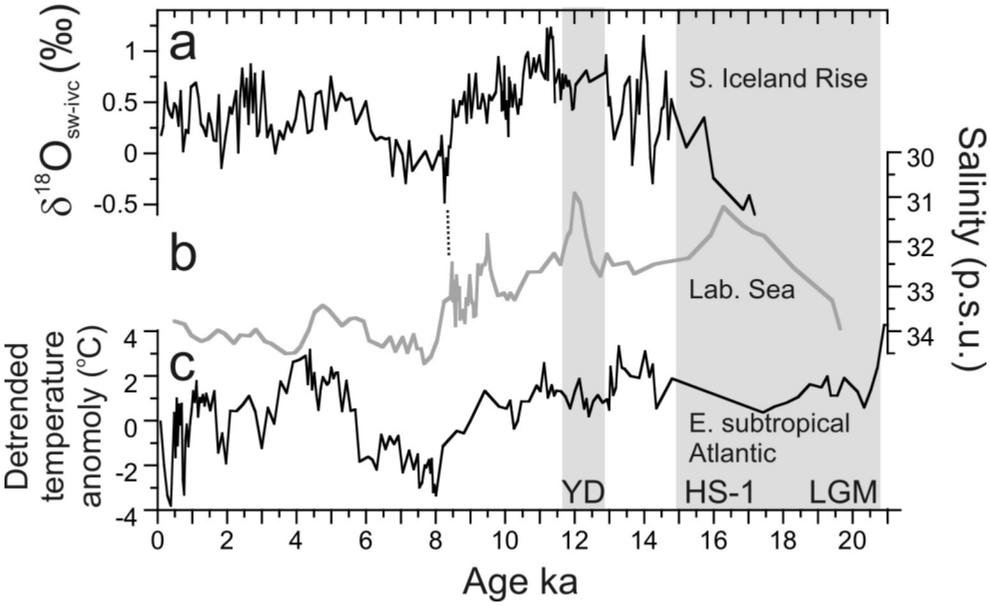
1 **Fig. 5**



a) IRD
IRD (#/g)
IRD ratio
Nordic Seas
S. Iceland
b) Atlantic Inflow
Mg/Ca temperature (°C)
G. bulloides
G. inflata
Planktic δ18Osw-ivc (‰)
c) Subpolar Water
N. pachy. (s)
d) Deep N. Atlantic
Benthic δ13C (‰)
C. wuell. δ13C
C. wuell. δ18Oivc
Benthic δ18Oivc (‰)
NGRIP δ18O ‰
e) Greenland temperature
Age ka

2

Fig. 6



a
b
c
S. Iceland Rise
Lab. Sea
E. subtropical
Atlantic
YD
HS-1
LGM
δ18Osw-ivc (‰)
Salinity (p.s.u.)
Detrended
temperature
anomoly (°C)
Age ka