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Multidecadal ocean variability and NW European ice sheet surges during the last deglaciation

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[1] A multiproxy paleoceanographic record from the Atlantic margin off the British Isles reveals in unprecedented detail discharges of icebergs and meltwater in response to sea surface temperature increases across the last deglaciation. We observe the earliest signal of deglaciation as a moderate elevation of sea surface temperatures that commenced with a weakly developed thermocline and the presence of highly ventilated intermediate waters in the Rockall Trough. This warming pulse triggered a series of multidecadal ice-rafted debris peaks that culminated with a major meltwater discharge at 17,500 years before present related to ice sheet disintegration across the NW European region. The impact of meltwater caused a progressive reduction in deep water ventilation and a sea surface cooling phase that preceded the collapse of the Laurentide Ice Sheet during Heinrich event 1 by 500–1000 years. A similar sequence of rapid ocean-ice sheet interaction across the European continental margin is identified during the Bølling-Allerød to Younger Dryas transition. The strategic location of our sediment core suggests a sensitive and rapid response of ice sheets in NW Europe to transient increases in thermohaline heat transport.

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1. Introduction

[2] Abrupt shifts in North Atlantic climate records indicate a strong link between the meridional heat flux driven by thermohaline convection and glacimarine discharges from the Late Pleistocene ice sheets [*Ruddiman and McIntyre*, 1981; *Lehman and Keigwin*, 1992; *Bond et al.*, 1993]. The impact of meltwater on thermohaline circulation is often invoked as a forcing factor for rapid climate change [*Ganopolski and Rahmstorf*, 2001], although the mechanism for promoting submillennial scale instability of marine ice margins remains unclear [*Bond and Lotti*, 1995; *van Kreveld et al.*, 2000]. The massive discharges of icebergs during Heinrich events have been related to periodic collapses of the Laurentide Ice Sheet (LIS) [*MacAyeal*, 1993], but internal ice sheet dynamics cannot explain the 1–2 ka cycle of glacial fluctuations that is related to the stadial events of the Greenland summit climate record [*Bond and Lotti*, 1995]. There is increasing evidence that this millennial-scale ice-rafted debris (IRD) signal was produced by ice sheets that were much smaller than the LIS and directly influenced by ocean climate [*Fronval et al.*, 1995; *Elliot et al.*, 1998; *Knutz et al.*, 2001]. The position of the Icelandic, Fennoscandian, and British ice sheets in vicinity of the main path of the North Atlantic Drift which presently feeds the thermohaline overturn cell in the Nordic Seas [*McCartney and Talley*, 1984], provides the potential for a close coupling between oceanic heat transport and glacial mass balances. A differential response between circum-North Atlantic ice sheets, possibly linked to changes in the thermohaline circulation, has been suggested from IRD provenance studies across Heinrich events 1 and 2 [*Grousset et al.*, 2000; *Scourse et al.*, 2000]. Here we present a paleoclimatic record of the last deglaciation from the NE Atlantic margin, which provides new evidence of multidecadal scale interaction between ocean circulation and sensitive ice sheets in NW Europe.

2. Material and Methods

[3] Core DAPC2 was retrieved from a contourite drift deposit located SE of Rosemary Bank, northern Rockall Trough (58°58.10′N, 09°36.75′W) at a water depth of 1709 m (Figure 1). The site is at present influenced by Norwegian Sea Overflow Water (NSOW; a precursor water mass of North Atlantic Deep Water, NADW) crossing the Wyville-Thomson Ridge [*Ellett and Roberts*, 1973; *New and Smythe-Wright*, 2001] and recirculated NADW [*McCartney*, 1992]. The chronology of DAPC2 is constrained by seven ^{14}C-AMS datings (Table 1) and a stratigraphic control point, which relates a sharp increase in foraminiferal abundances to the YD-Holocene transition of the GRIP δ^{18}O profile (Supplementary Information, available at http://www.g-cubed.org). A calendar year timescale was derived from the linear extrapolation between age control points using CALIB 4.1 [*Stuiver et al.*, 1998], which included a marine reservoir correction of 400 years. The marine ^{14}C reservoir age is known to have varied across the last deglaciation in response to ocean circulation changes and atmospheric ^{14}C production rates [*Voelker et al.*, 1998] so the timescale represents a first order approximation. Sedimentation rates average 23 cm ka^{-1} across the last deglaciation, encompassing the Younger Dryas (YD), Bølling-Allerød (B-A) and Heinrich event 1 (H-1), increasing to >70 cm ka^{-1} across the interval associated with the late glacial-early deglaciation. Similarly high sedimentation rates have previously been observed in glacial sections of contourite deposits from the Rockall Trough and are probably related to deposition from meltwater plumes emanating from the western European shelf margins [*Knutz et al.*, 2001; *Lassen et al.*, 2002]. Samples were obtained at 2 cm intervals, enabling us to recognize paleoceanographic shifts at a multidecadal temporal resolution. The high sedimentation rates and the sharpness of the proxy signals observed in DAPC2 suggest that the influence of bioturbation is negligible [*Anderson*, 2001]. Nevertheless, should a small amount of bioturbational blurring of the paleoclimatic signals have occurred, then the speed of the events recorded in DAPC2 is likely to be faster than we claim here.

[4] The relative abundance of the polar foraminifera *Neogloboquadrina pachyderma* sinistral (*Nps*) in the >125 μm fraction was counted in

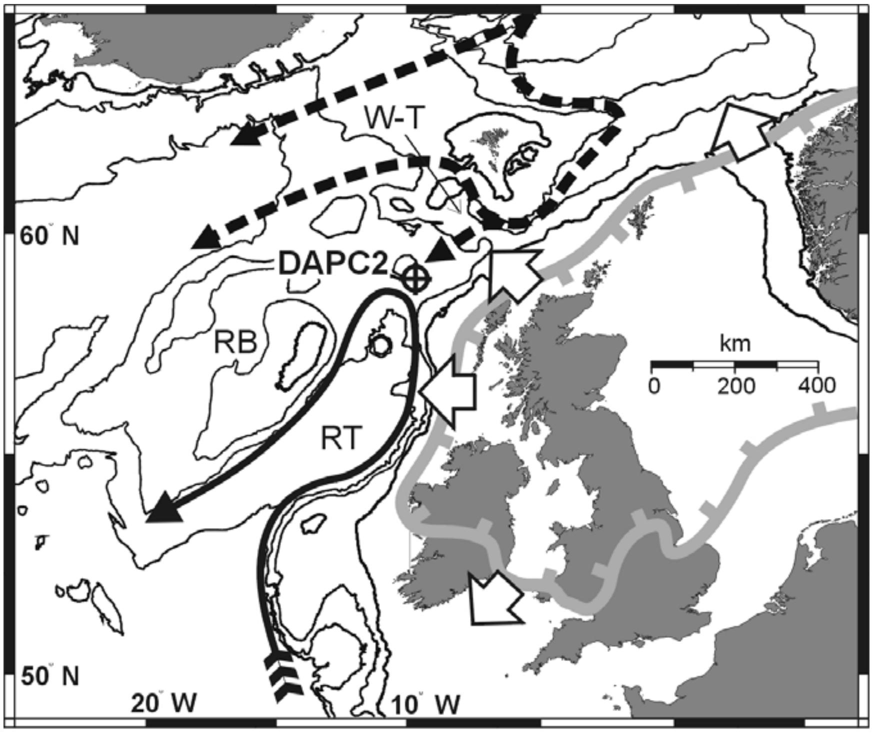

Figure 1. Location of core DAPC2, modern deep ocean circulation and the extend of the LGM ice sheet in NW Europe. Broad arrows indicate the main glacimarine outlets [*McCabe and Clark*, 1998; *Knutz et al.*, 2001]. The thin, black arrow illustrates a northerly boundary current driven by recirculated North Atlantic Deep Water, which flows along the European continental slope at depths of 2–3 km [*McCartney*, 1992]. The broken arrow represents Norwegian Sea Overflow Water, which enters the northeast Atlantic across the shallow sills between Iceland and Scotland [*Ellett and Roberts*, 1973; *New and Smythe-Wright*, 2001]. RB: Rockall Bank, RT: Rockall Trough, W-T: Wywille-Thomson Ridge. Contours represent 200 m (thick line), 1000, 2000 and 3000 m (thin lines) water depth.

sample splits containing >300 specimens, as a relative indicator of sea surface temperature (SST) changes. In the North Atlantic this species makes up more than 95% of the planktic faunal assemblage at summer SSTs below 5°C [*Johannessen et al.*, 1994]. The abundance of detrital components (grains per gram dry bulk sediment) was analyzed in the >250 μm fraction. The fractured quartz (FQ) component provides a positive indicator of iceberg-transported debris [*Knutz et al.*, 2001], while the content of detrital carbonate (DC) is related to deposition from distal icebergs derived from the LIS [*Andrews and Tedesco*, 1992]. The FQ flux was calculated from the FQ abundance and bulk mass accumulation rates (Supplementary Information, available at http://www.g-cubed.org). The stable isotope composition of planktonic foraminifera species *Nps* and *Globigerina bulloides* (*Gb*), and the epibenthic species *Cibicidoides wuellerstorfi* (*Cw*) was determined using a Micromass Multiprep system attached to a VG PRISM mass spectrometer. $\delta^{18}O$ and $\delta^{13}C$ values are reported relative to the Vienna Peedee belemnite (VPDB) international standard with analytical precision better than ±0.06%. The content of sortable silt (10–63 μm) was measured on a Sedigraph grain size analyzer subsequent to removal of calcium carbonate using 2 M acetic acid. In contourite drift sediments, this parameter is primarily related to relative changes in flow speed of near-bottom currents [*McCave et al.*, 1995b].

3. Rapid Ocean-Ice Sheet Responses

[5] Major events of increased supply in ice-rafted debris during the Younger Dryas and the early deglaciation are clearly recognized by the flux of FQ (Figure 2b). The ratio between DC and FQ grains provides a signal of icebergs discharged from the Hudson Bay region of the LIS and pinpoints the

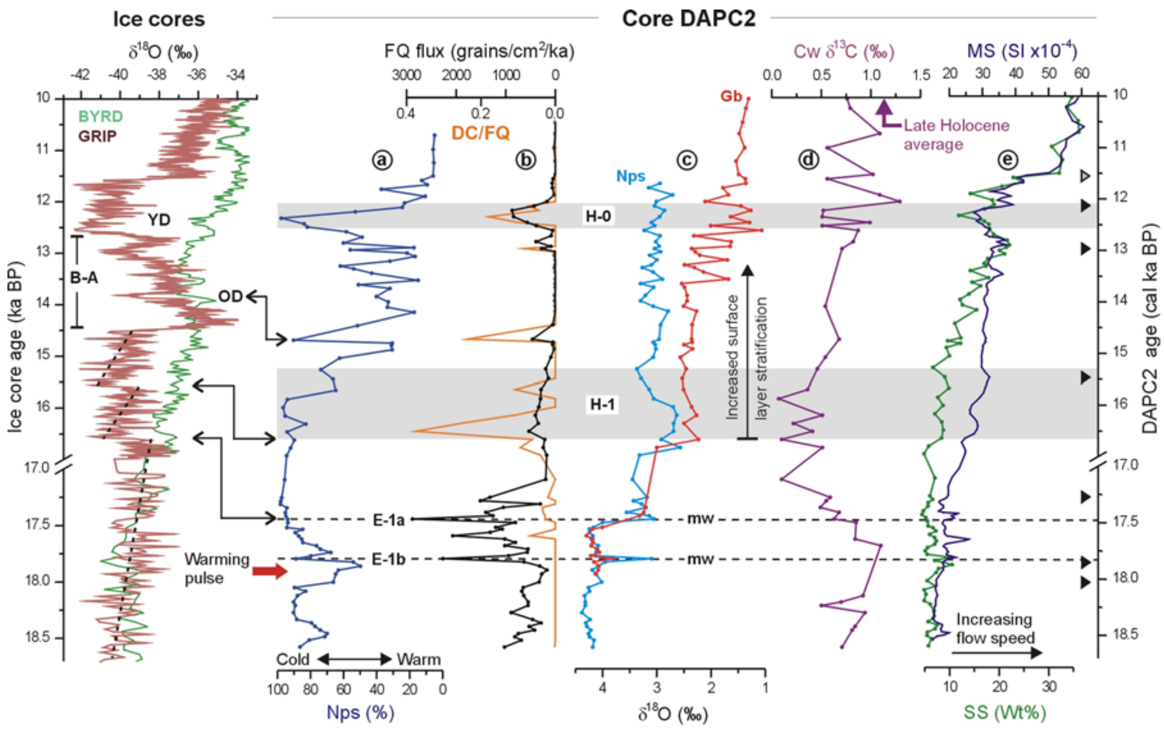

Figure 2. Paleoceanographic time series from core DAPC2. (a) Relative abundance of the polar foraminifera *Neogloboquadrina pachyderma* sinistral (*Nps*) in the >125 μm fraction. (b) Flux of ice-rafted, fractured quartz (FQ) (number of grains cm^{-2} ka^{-1}) and the ratio of pale-yellowish detrital carbonate (DC) to FQ. (c) Oxygen isotope records ($\delta^{18}O$) of planktonic foraminifera *Nps* and *Globigerina bulloides* (*Gb*). (d) Carbon isotope record ($\delta^{13}C$) of epibenthic foraminifera *Cibicidoides wuellerstorfi* (*Cw*). The arrow on the scale bar represents an average of late Holocene *Cw* $\delta^{13}C$ values measured in DAPC2. (e) Records of magnetic susceptibility (MS) and weight percentage of sortable silt (SS%, 10–63 μm terrigenous fraction). Heinrich events (H-1, H-0), European glacimarine events (E-$1_{a\text{-}b}$), Bølling-Allerød (B-A), Older Dryas (OD), and Younger Dryas (YD) are indicated. The FQ flux was calculated from the grain abundance and bulk mass accumulation rates (Supplementary Information, available at http://www.g-cubed.org). The timescale is based on linear interpolation of 7 AMS ^{14}C dates (black triangles) performed on single-species foraminifera (Table 1), and a stratigraphic control point (gray triangle), which relates a sharp increase in foraminiferal abundances to the YD-Holocene transition in the Greenland summit GRIP $\delta^{18}O$ record. Left-hand panel shows $\delta^{18}O$ records from GRIP [*Dansgaard et al.*, 1993] and west Antarctic Byrd [*Johnsen et al.*, 1972] ice cores on a common age scale [*Blunier and Brook*, 2001] allowing comparison with the atmospheric temperature changes in the northern and southern hemisphere. The late glacial warming trends in the GRIP $\delta^{18}O$ record are depicted by the hatched line. Note the change in timescale at 17 ka BP.

glacial collapse associated with H-1, which in the central North Atlantic IRD belt is dated between 16 and 17 ka before present (BP) [*Andrews*, 1998]. We also observe a DC peak within the B-A interval, which likely corresponds to the Older Dryas cooling. The sharp centennial-scale IRD peaks observed during the earliest deglaciation between 18 and 17 ka BP (Figure 2b) are outstanding as these occur more than 1500 years prior to the peak of the H-1 event. The two most prominent IRD peaks within the broad IRD increase correlate with sharp $\delta^{18}O$ depletions suggesting that these are related to regional glacimarine discharges and not just local anomalies (Figures 2b and 2c). The relative magnitude of the IRD flux between 18 and 17 ka, on average more than three times greater than during H-1, and the sharp meltwater pulses points to a nearby source on the NE Atlantic margin for these events.

[6] The most proximal source for iceberg discharges at this location was the NW sector of the British Ice Sheet, which during the early deglaciation advanced onto the shelf margin west of Scotland [*McCabe and Clark*, 1998; *Knutz et al.*, 2001] (~200 km from the DAPC2 core site). We cannot rule out that other ice sheets along the NE

Atlantic margins contributed to the earliest deglacial IRD events, in particularly the Fennoscandian Ice Sheet, which drained into the Norwegian Channel to the NE of the British Isles [*Sejrup et al.*, 1994]. However, the low abundance of DC precludes the Hudson Bay region of the LIS as the main source of icebergs. Low DC layers observed between H-1 and H-2 in the Labrador Sea region may reflect a more frequent response from parts of the LIS other than the Hudson Bay ice stream [*Stoner et al.*, 1996], but it is difficult to associate these horizons, apparently of mixed downslope and hemipelagic origin, with the large IRD fluxes of the DAPC2 core between 18 and 17 ka BP. This conclusion leads us to term the glacimarine discharges observed during the earliest deglaciation as European events, E-$1_{a\text{-}b}$ (Figure 2). We suggest that the E-1 events in DAPC2 are analogues to the European 'precursors' of Heinrich events recently demonstrated on the NE Atlantic margin [*Grousset et al.*, 2000; *Scourse et al.*, 2000].

[7] The E-$1_{a\text{-}b}$ events are immediately preceded by a rapid warming pulse in SST, indicated by a sharp reduction in the abundance of *Nps* (Figure 2a). This warming signal starts abruptly at ~18 ka with an initial change lasting some 50 years. The warming is terminated at ~17.8 ka, and SST rapidly cools and returns to its initial late glacial level within several decades. The cooling is coincident with the first glacimarine E-1_b event, marked by an abrupt negative *Nps* δ^{18}O excursion of ~1% (Figures 2a–2c). The E-1_b peak suggests a rapid glacial response to sea surface warming, possibly related to onset of a fast-flow regime of the British Ice Sheet, driven by high meltwater production along its western margins [*McCabe and Clark*, 1998]. The E-1_a event at ~17.5 ka BP is characterized by an abrupt ~1% decrease in planktonic (*Nps* and *Gb*) δ^{18}O occurring over a period of several decades. The negative step change in δ^{18}O associated with the E-1_a meltwater pulse suggests a more widespread deglaciation in the NE Atlantic sector, possibly involving the collapse of the Barents Sea ice shelf, which has previously been estimated at ~17 ka [*Bischof*, 1994; *Hebbeln et al.*, 1994]. The multidecadal resolution expressed by the DAPC2 record allows the E-1 glacimarine events to be clearly discerned from the H-1 event, which is evident as a prolonged negative *Nps* δ^{18}O anomaly between 17 and 16 ka BP (Figure 2c).

[8] At the end of the B-A warm interval a small increase in IRD at 13 ka BP is followed by a strong IRD peak between 12.5 and 12.0 ka BP associated with the H-0 event. The H-0 and the precursor event coincide with a series of negative *Gb* δ^{18}O anomalies observed across the B-A to YD transition. However, the first two of the δ^{18}O anomalies, between 13.5 and 13.0 ka BP, clearly precede the IRD peaks (Figure 2c) and are therefore unlikely to represent a glacimarine meltwater signal. From the proximity of the DAPC2 core to the European continental margin the negative δ^{18}O anomalies are more likely to represent glacifluvial discharges produced from terrestrial ice-margins. The paleoclimatic sequence across the Allerød-YD transition supports the rapid response of NW European ice sheets to SST warming similar to that of the earliest deglaciation (Figures 2a and 2b).

4. Ocean Circulation Changes

[9] We use information derived from the stable isotope composition of planktonic and benthic foraminifera species to infer changes in North Atlantic circulation and the impact of European meltwater pulses on thermohaline circulation across the last deglaciation (Figures 2c and 2d). The offset in δ^{18}O of foraminiferal species representing surface water (*Gb*) and thermocline water (*Nps*) provides an indication of the vertical hydrographic structure (temperature, salinity) of the surface ocean layer and its potential for convection [*Hillaire-Marcel and Bilodeau*, 2000]. Benthic δ^{13}C, in turn, serves as a paleoceanographic indicator for the relative contributions of nutrient-depleted NADW (high δ^{13}C) and nutrient-enriched water originating from the southern hemisphere (low δ^{13}C) [*Kroopnick*, 1985]. Prior to the E-1_a meltwater peak at 17.5 ka BP *Nps* and *Gb* δ^{18}O signals are virtually identical, which suggests that the initial warming pulse at ~18 ka BP was associated with a well-mixed surface-subsurface layer that is indicative of a weakly developed thermocline and decreased vertical stability of the

upper water column (Figure 2c). This interpretation is supported by the high benthic $\delta^{13}C$ values (>0.8%) prior to 17.5 ka BP (Figure 2d), which imply the presence of a nutrient-deficient and highly ventilated water mass, presumably Glacial North Atlantic Intermediate Water (GNAIW) that is known to have extended to depths of ~2000 m [*Duplessy et al.*, 1988; *Oppo and Lehman*, 1993; *Sarnthein et al.*, 1994]. The decrease in benthic $\delta^{13}C$ directly following the E-1_b event suggests that high meltwater fluxes from western Europe caused a progressive reduction in deep water ventilation. This reflects a slow-down in the formation of nutrient-depleted GNAIW, which in the deepest parts of the Rockall Trough became replaced by nutrient-enriched southern source water [*Oppo and Lehman*, 1993; *McCave et al.*, 1995a]. The general increase in *Gb-Nps* $\delta^{18}O$ gradients after ~16 ka BP points to increased stratification and the shoaling of the thermocline during the B-A.

[10] To constrain our palaeoceanograhic interpretation, we have extracted sedimentological properties from DAPC2 that provide a physical indication of changes in thermohaline circulation (Figure 2e). The sortable silt percentage (SS%) is used as a proxy of near-bottom flow, for which greater abundance suggests faster relative flow speeds [*McCave et al.*, 1995b]. The strong influence of bottom current sediment sorting in core DAPC2 is supported by the positive correlation between SS% and magnetic susceptibility, which in other records from the NE Atlantic margin has been linked to the intensity of NSOW [*Rasmussen et al.*, 1996]. The low SS% values prior to ~17 ka suggest that bottom currents were too weak to produce a measurable sorting effect on the 10–63 μm silt concentrations (Figure 2e). The gradual increase in SS% from ~17 ka indicates a progressive strengthening of flow speeds up to a local maximum during the late Allerød warming. Combined with an increase in benthic $\delta^{13}C$, this points to an increased vigor of deep water recirculation with a gradually increasing contribution from a northern deep water source, and a corresponding weakening of the southern hemisphere water contribution (Figure 2d). We note that the benthic $\delta^{13}C$ levels during the B-A interval represent a mixed water mass rather than a pure northern source end-member (Holocene $\delta^{13}C$ values average 1.14% compared with an average of 0.57% during the B-A interval). A marked reduction in SS% across the YD cooling and an abrupt decrease in $\delta^{13}C$ at 12.5 ka BP indicate a transient decrease in deep ventilation suggestive of a close linkage between meltwater injection and convective slow-down in the North Atlantic (Figures 2d and 2e). Flow speed then increased rapidly in two discrete steps. The first increase occurred immediately after the YD meltwater peak around 12.2 ka (Figure 2c), while the second marks the YD-Holocene transition placed at 11.5 ka according to the GRIP $\delta^{18}O$ record. By analogy with the sharpness of the YD termination in the GRIP ice core (Figure 2) we infer that the transition from the convective slow-down during the YD to a modern circulation regime with full-scale NSOW influence occurred in as little as 60 years.

5. Discussion and Conclusions

[11] Comparison of our records with the $\delta^{18}O$ signal from the GRIP ice core [*Dansgaard et al.*, 1993] reveals a broad correlation between transient shifts in NE Atlantic ocean circulation and the evolution of regional climate during the last deglaciation (Figure 2). The warming pulse observed prior to the E-1 events in DAPC2 is not explicitly revealed in the GRIP ice core but appears to be embedded into the broad temperature increase observed between 18.5 and 16.5 ka. According to our correlation the early deglacial warming trend in the GRIP $\delta^{18}O$ record was aborted by the E-1_a meltwater peak. Renewed increase in Greenland air temperatures occurred subsequent to the E-1_a event, but again climatic amelioration was disrupted, this time by the H-1 meltwater pulse. The age discrepancies of 1.3–1.6 ka (including the 400 year correction used in the ^{14}C age calibration) that arise from the correlation between the late glacial $\delta^{18}O$ transitions in the GRIP record and the E-1 and H-1 meltwater pulses in DAPC2 are likely to represent the effects of increased marine ^{14}C reservoir ages [*Voelker et al.*, 1998; *Waelbroeck et al.*, 2001] on our converted, calendar-year timescale. Comparison with the Byrd ice core [*Johnsen et al.*, 1972; *Blunier and Brook*, 2001] (Figure 2) shows that atmospheric temper-

Table 1. Radiocarbon Dating Results From Core DAPC2

Laboratory Number	Material	Depth, cm	^{14}C Age,[a] yr BP	Error Age, ±1σ yr BP	Calendar Age[b]
AAR-5209	*G. bulloides*	42	8,190	±90	9.067
AAR-5210	*N. pachyderma* (*s*)	78	10,430	±90	12.193
AAR-6304	*N. pachyderma* (*s*)	96	10,980	±110	12.938
AAR-6305	*N. pachyderma* (*s*)	138	13,010	±130	15.458
AAR-5211	*N. pachyderma* (*s*)	160	14,500	±110	17.262
AAR-6306	*N. pachyderma* (*s*)	199	15,000	±170	17.835
AAR-5212	*N. pachyderma* (*s*)	209	15,160	±140	18.014

[a] Including marine reservoir correction of 400 years.
[b] Obtained from CALIB 4.1 [*Stuiver et al.*, 1998].

atures over Antarctica continued to increase over this early deglacial period. We surmise that the Byrd temperature record, from its isolated location in Antarctica, is more representative of the mean global evolution of atmospheric temperature. Deviations from this trend, as seen in the GRIP ice core record, are likely a response to the forcing that ocean-ice sheet interaction, notably thermohaline heat transport and meltwater production exerted on North Atlantic climates [*Grootes et al.*, 2001].

[12] Our findings suggest that the initial cooling observed prior to H-1 in many North Atlantic records [*Bond et al.*, 1993; *Labeyrie et al.*, 1999; *Bard et al.*, 2000] is likely to be the effect of meltwater discharge from European ice sheets. However, the mechanism that links the E-1 and H-1 event remains enigmatic. One possibility is that surging of ice sheets in NW Europe acted as a trigger mechanism for H-1 through a sea level rise that over a period of centuries destabilized the marine-based margins of the LIS [*Andrews*, 1998; *Grousset et al.*, 2000]. This hypothesis is supported by the step-like decrease in δ^{18}O at 17.3 ka BP which marks a major deglaciation of the entire European and Arctic region [*Jones and Keigwin*, 1988; *Hebbeln et al.*, 1994; *Sarnthein et al.*, 1995]. Alternatively, the time lag between E-1 and H-1 may reflect a slower response of the LIS to a common climatic forcing manifest by the late glacial warming pulse [*Scourse et al.*, 2000]. A phase of climatic amelioration with a maximum around 18–17 ka BP has previously been documented in the NE Atlantic [*Lagerklint and Wright*, 1999; *Zaragosi et al.*, 2001] and Scandinavia [*Vorren et al.*, 1988] which points to a regional incursion of temperate water masses along the European continental margin. The corollary is that the warming might represent the initial developing stage of a Dansgaard-Oeschger event that was disrupted by glacimarine discharges before it could trigger full-scale thermohaline convection in the Nordic Seas, and consequently produce a sharp temperature increase in the Greenland climate record.

[13] In summary, the records from DAPC2 suggest that a major ice sheet collapse in NW Europe was triggered by a brief warming pulse that punctuated the late glacial-early deglacial interval. The ensuing sequence of ocean-ice sheet interaction in the NE Atlantic appears to have been critical for the glacimarine discharges that subsequently emerged from the Hudson Bay region during H-1. Our results confirm the high sensitivity of the glacial North Atlantic to even minor changes in freshwater fluxes that are demonstrated in ocean-climate models [*Ganopolski and Rahmstorf*, 2001].

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