nature geoscience

ARTICLES

PUBLISHED ONLINE: 25 JULY 2010 | DOI: 10.1038/NGEO921

Extreme deepening of the Atlantic overturning circulation during deglaciation

Stephen Barker[1]*, Gregor Knorr[1,2], Maryline J. Vautravers[3†], Paula Diz[1†] and Luke C. Skinner[4]

Glacial terminations during the late Pleistocene epoch are associated with changes in insolation. They are also punctuated by millennial-scale climate shifts, characterized by a weakening and subsequent strengthening of the Atlantic meridional overturning circulation. This ubiquitous association suggests that these oscillations may be a necessary component of deglaciation. Model simulations have suggested that the period of weakened circulation during these terminal oscillations would be followed by an overshoot of the circulation on its resumption, but this phenomenon has not yet been observed. Here we use radiocarbon measurements of benthic foraminifera and carbonate preservation indices to reconstruct ventilation changes in the deep South Atlantic Ocean over the past 40,000 years. We find evidence for a particularly deep expansion of the Atlantic overturning cell directly following the weak mode associated with Heinrich Stadial 1. Our analysis of an ocean general circulation model simulation suggests that North Atlantic Deep Water export during the expansion was greater than that of interglacial conditions. We find a similar deep expansion during Dansgaard–Oeschger Interstadial Event 8, 38,000 years ago, which followed Heinrich Stadial 4. We conclude that the rise in atmospheric CO_2 concentrations and resultant warming associated with an especially weak overturning circulation are sufficient to trigger a switch to a vigorous circulation, but a full transition to interglacial conditions requires additional forcing at an orbital scale.

The present interglacial period, known as the Holocene epoch (past ∼10 kyr), is characterized by an overturning circulation within the Atlantic basin (the Atlantic meridional overturning circulation, AMOC) that plays an important role in the transfer of heat to high northern latitudes[1]. The modern AMOC involves formation of North Atlantic Deep Water (NADW) in the high-latitude North Atlantic Ocean. This deep water is exported southwards and fills the Atlantic basin with relatively well ventilated (high $\delta^{13}C$, $[CO_3^{=}]$ and $\Delta^{14}C$) waters. As NADW spreads southwards, it mixes with more poorly ventilated (relatively low $\delta^{13}C$, $[CO_3^{=}]$ and $\Delta^{14}C$) deep waters of southern origin, defining a mixing zone between NADW and underlying Antarctic Bottom Water within the South Atlantic Ocean (see Supplementary Information). During the last glacial period, the formation region of NADW is thought to have been pushed southwards and its penetration into the deep Atlantic Ocean limited by the enhanced incursion of southerly sourced deep waters[2]. Model simulations suggest that the strength of the AMOC may also have been reduced[3]. The transition from glacial to interglacial conditions therefore involved significant changes not only in continental ice volume (sea level), global temperature and atmospheric CO_2, but also in ocean circulation. Furthermore, AMOC changes themselves seem to play an instrumental role in the mechanism of glacial termination.

Terminal oscillations of the AMOC

The ultimate driver of deglacial climate change is the changing geometry of Earth's orbit around the sun[4–7]. However, additional feedbacks (such as from greenhouse gas concentrations and the ice albedo effect) are required to supplement the rather small changes in orbital forcing[7]. The last deglacial period (Termination 1, T1, ∼19–10 kyr) was punctuated by several abrupt changes in the mode of the AMOC. In particular, an extended cold period in the North Atlantic ∼18–14.6 kyr ago (Heinrich Stadial 1, HS1) is thought to have been related to a significant weakening of the AMOC (ref. 8) (note that Heinrich Event 1 occurred during HS1; the two should not be regarded as synonymous). Simultaneous warming across the Southern Ocean and the release of CO_2, partially as a result of the bipolar see-saw[9], could have provided the additional global warming necessary for termination to occur[9,10]. The association of AMOC see-saw oscillations with glacial terminations during the late Pleistocene seems ubiquitous[11], suggesting that they are a necessary component of deglacial climate change.

HS1 ended with an abrupt strengthening of the AMOC (ref. 8) and significant warming across the North Atlantic region at the start of the Bølling–Allerød (B–A) warm interval ∼14.6 kyr. Temperatures over Greenland rose by ∼9 °C within decades at this time[12], implying that interglacial-like conditions were attained several thousands of years earlier than the actual start of the present interglacial period. The B–A can therefore be considered as anomalously warm. Furthermore, climate models of different complexities show an AMOC overshoot during the B–A (refs 13, 14). During this event the export of deep water from the Atlantic basin may have been enhanced even with respect to expectations for the modern day[13]. Recent evidence from the North Atlantic suggests that intermediate-water export during the B–A may indeed have been enhanced relative to Holocene conditions[15]. Here we apply benthic radiocarbon analyses and multiple indices for carbonate dissolution to a sediment core retrieved from the deep South Atlantic Ocean (TNO57-21; 41.1° S, 7.8° E, 4,981 m) to investigate these predictions.

Rapid ventilation changes in the deep South Atlantic Ocean

Modern deep waters at the site of TNO57-21 are poorly ventilated with respect to the North Atlantic Ocean and reflect the

[1]School of Earth and Ocean Sciences, Cardiff University, Cardiff CF10 3YE, UK, [2]Alfred Wegener Institute, 27570 Bremerhaven, Germany, [3]British Antarctic Survey, Cambridge CB3 0ET, UK, [4]Department of Earth Sciences, University of Cambridge, Cambridge CB2 3EQ, UK. †Present addresses: Department of Earth Sciences, University of Cambridge, Cambridge CB2 3EQ, UK (M.J.V.); Department of Xeociencias Mariñas e Ordenación do Territorio, Facultade de Ciencias, University of Vigo, 36200 Vigo, Spain (P.D.). *e-mail: barkers3@cf.ac.uk.

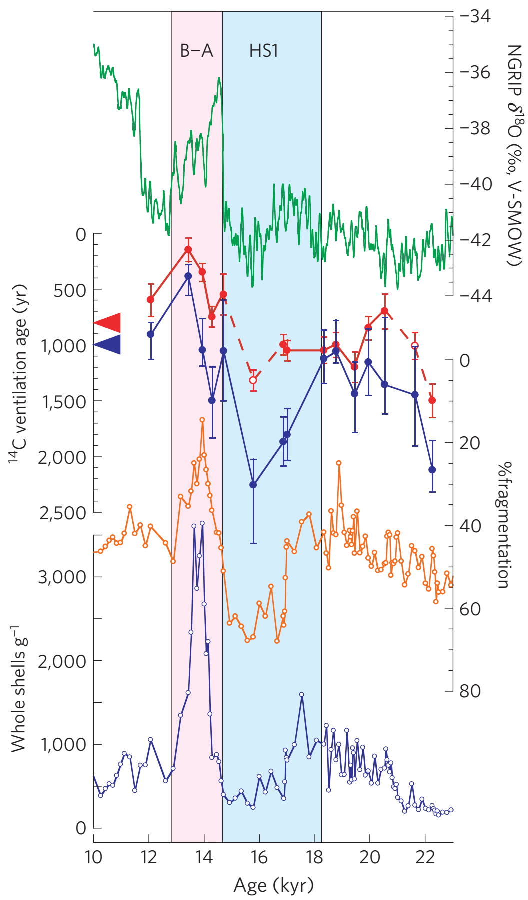

Figure 1 | Deglacial radiocarbon and carbonate preservation. Benthic ^{14}C age and $CaCO_3$ preservation in TNO57-21 compared with temperature variability over Greenland (the NGRIP ice core[32]). ^{14}C ventilation ages are plotted as benthic–planktonic age differences (red) or using the projection age method (blue) (see Supplementary Information). The error bars account for measurement errors as well as calibration errors and uncertainties in atmospheric $\Delta^{14}C$ (see Supplementary Information). The open benthic–planktonic markers represent inferred values (see Supplementary Information). The fragment counts are from ref. 9. The red and blue triangles represent modern values[33]. A significant improvement of deep-water ventilation occurs at ∼14.6 kyr. Increased carbonate dissolution occurs within HS1 (blue rectangle); enhanced preservation occurs during the B–A (pink rectangle).

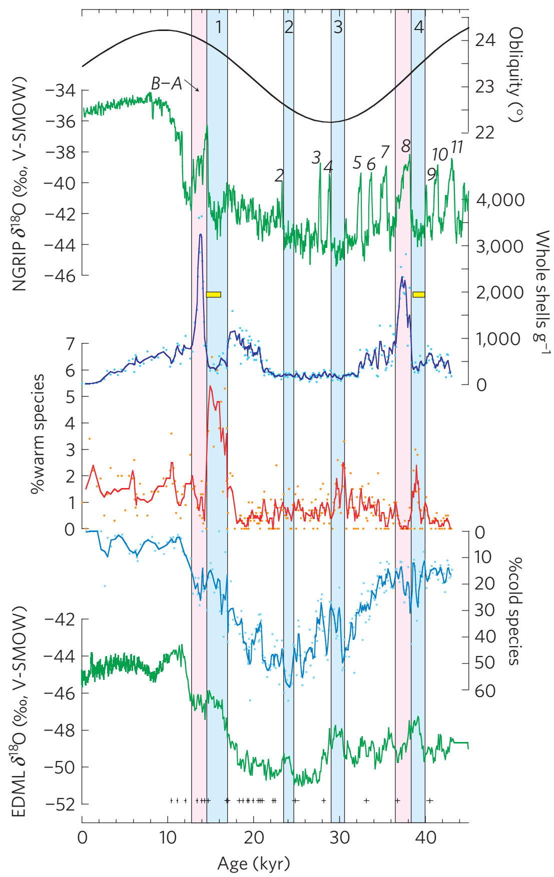

Figure 2 | The bipolar see-saw and deep-water ventilation over the past 40 kyr. Temperature variability over Greenland (NGRIP; ref. 32) and Antarctica (EDML; ref. 34) and carbonate preservation (whole shells g^{-1}) in TNO57-21 (three-point running mean). Warm planktonic foraminiferal species in TNO57-21 (see the Methods section) (ref. 9 plus this study) suggest intervals of enhanced preservation follow periods of weakened AMOC (cold in the north and warm in the south). Cold planktonic foraminiferal species reveal generally cold conditions throughout Marine Isotope Stage 2. The yellow bars represent wet intervals in northeast Brazil during intervals of weakened AMOC (ref. 29). The crosses indicate age control for TNO57-21 (ref. 9). The upper numbers denote Heinrich events. The numbers and letters in italics indicate interstadial episodes. Also shown is the obliquity of Earth's rotation axis[35].

dominance of southern-sourced deep waters at this location (see Supplementary Information). During glacial times, the influence of northern-sourced deep waters at the core site is thought to have been reduced even further[16,17]. Our benthic foraminiferal ^{14}C measurements confirm this with an increase in ^{14}C ventilation age of up to a few hundred years (that is, 'poorer' ventilation) during the Last Glacial Maximum (LGM) with respect to the modern (Fig. 1). During HS1 we observe an increase in ventilation age of up to 1,000 yr. Previous studies have suggested that the North Atlantic was dominated by poorly ventilated waters during the interval of weakened AMOC associated with HS1 (refs 18–20). The release of CO_2 from within the Southern Ocean during HS1 (refs 9,21) represented the exhalation of carbon stored during glacial times. Concomitant depletions in ^{14}C observed in the Pacific Ocean[22] and throughout the Atlantic Ocean[19,20] reflect this release (which probably contributed to the significant drop in atmospheric $\Delta^{14}C$ during the so-called Mystery Interval[23]). The observation of low-$\Delta^{14}C$ deep waters at the site of TNO57-21 (which lies on the northern margin of the Southern Ocean) suggests that the in-mixing of glacial-aged ^{14}C-depleted waters was associated with the increased penetration of southern-sourced deep waters into the Atlantic Ocean while the AMOC was in a weakened state. At the same time, we also observe intense carbonate dissolution in TNO57-21 (Fig. 1). This lends credence to the idea that northward-penetrating deep waters during HS1 were particularly poorly ventilated.

The most pertinent feature of the deglacial ^{14}C results shown in Fig. 1 is the occurrence of significantly younger-than-modern ^{14}C ventilation ages during the Bølling–Allerød warm interval. Benthic $\delta^{13}C$ measured in TNO57-21 (ref. 16) and Nd isotopes from nearby core RC11-83 (ref. 17) suggest a switch to a more northerly influence at this time, suggesting that the younger ventilation ages we observe were in response to the increased influence of NADW in the deep South Atlantic Ocean. Previous studies from the North Atlantic Ocean support the suggestion that the Atlantic Ocean was as well or even better ventilated (with respect to radiocarbon at least) during the Bølling–Allerød as compared with the modern[19,20]. We also observe a strong peak in $CaCO_3$ preservation during the

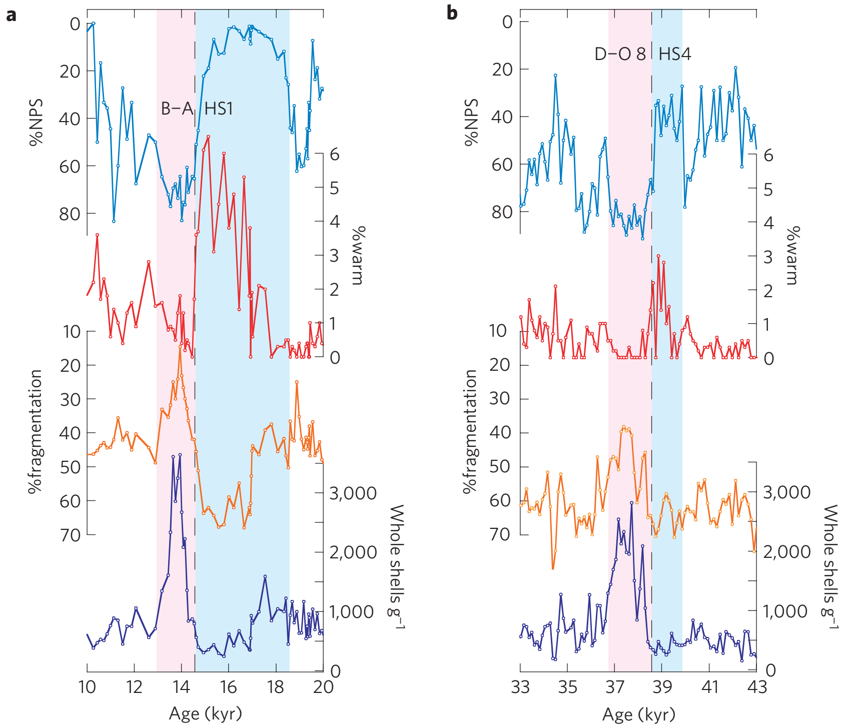

Figure 3 | Relative timing of surface- and deep-water responses. a,**b**, Co-registered records from TNO57-21 reveal the relative timing of surface- and deep-ocean changes across the B–A transition (**a**) and D–O 8 (**b**). %NPS is the percentage of left-coiling *Neogloboquadrina pachyderma* out of total *N. pachyderma*, with higher values reflecting colder surface conditions. %warm is the combined contribution of warm planktonic foraminiferal species to the entire assemblage (see the Methods section). %fragmentation is the degree of foraminiferal shell break-up (fragments/(fragments + whole shells)). The dashed line marks the shift in planktonic assemblages (that is, surface ocean). Fragment counts and warm species data in **a** are from ref. 9.

B–A (Fig. 1), suggesting a greater influence of better ventilated deep waters during the B–A than at any other time over the past ∼40 kyr (Fig. 2).

Our records also show an earlier peak in preservation associated with Dansgaard–Oeschger Interstadial Event 8 (D–O 8) ∼38 kyr ago, again following an interval of relatively increased dissolution during D–O Stadial 8 (equivalent to HS4). The abrupt onset and demise of these events suggests that they were not caused by whole-ocean changes in carbonate chemistry (see Supplementary Information), but rather that they reflect the rapid alternation between poorly ventilated and well-ventilated water masses. We argue that the preservation events we observe are due to the enhanced influence of well-ventilated northern-sourced deep waters. Our records are on their own independent age scale[9] and we can be confident that the preservation events correspond directly with Greenland interstadials (Fig. 2). In a recent study[9] we reported the increased presence of warm planktonic foraminiferal species in TNO57-21 at times of weakened AMOC as a consequence of the bipolar see-saw. Each of the preservation peaks we observe here occurs during a minimum in warm species following a maximum. This suggests that enhanced preservation occurs as the AMOC strengthens after a period of weakened circulation.

As the planktonic counts and $CaCO_3$ preservation proxies were measured in the same core material, we are able to determine unambiguously the relative timing of surface- and deep-water responses in the South Atlantic Ocean to changes in the AMOC (Fig. 3). In both observed cases, the increase in preservation (deep-water response) occurs within ∼300 years of the surface ocean shift. Maximum preservation is reached within about 1,000 yr. These results provide a critical constraint on the rapidity of signal propagation associated with abrupt changes in the AMOC. They also highlight the wholesale involvement of the Atlantic Ocean during abrupt climate changes.

AMOC overshoot in a deglacial model simulation

Our records of deep-water ^{14}C ventilation (for the deglaciation) and carbonate preservation suggest that the influence of NADW in the deep South Atlantic during the B–A and D–O 8 was enhanced with respect to both mean glacial and interglacial conditions. To explore the physical changes associated with such a transition we have re-analysed a transient deglacial model simulation spanning the Last Glacial Maximum through HS1 to the B–A (ref. 13). The model employed is a three-dimensional ocean general circulation model, based on the Hamburg large-scale geostrophic ocean circulation model[24], with a horizontal resolution of 3.5° and 11 levels in the vertical. Deglaciation is implemented in the model by a linear transition from glacial to interglacial background climate conditions plus various freshwater perturbations (see Supplementary Information for a fuller description of the model design and experimental set-up). The model results show that positive feedbacks associated with a vertical temperature relaxation in the North Atlantic Ocean and the advection of salt from the tropical Atlantic Ocean and the Indian Ocean can generate an AMOC overshoot at the B–A transition from a state with effectively no NADW export at 30° S to ∼27 Sv within about 200 years (the glacial and modern control runs are characterized by NADW exports of ∼8.5 Sv and ∼14 Sv respectively). During the overshoot the deep Atlantic Ocean is dominated by a vertically expanded AMOC cell[13] and the transition is characterized by significant changes in the deep Atlantic flow field below 4,000 m (Fig. 4). During the ‘off’ mode there is no NADW export from the Atlantic basin and deep waters move northward over the site of TNO57-21, transporting poorly ventilated waters to the core site (Fig. 4a). Within ∼200 years of the AMOC amplification, the average deep Atlantic flow field has changed from northward to southward velocities (Fig. 4b), suggesting an abrupt switch from southern- to northern-sourced deep waters bathing our core site. This contrasts

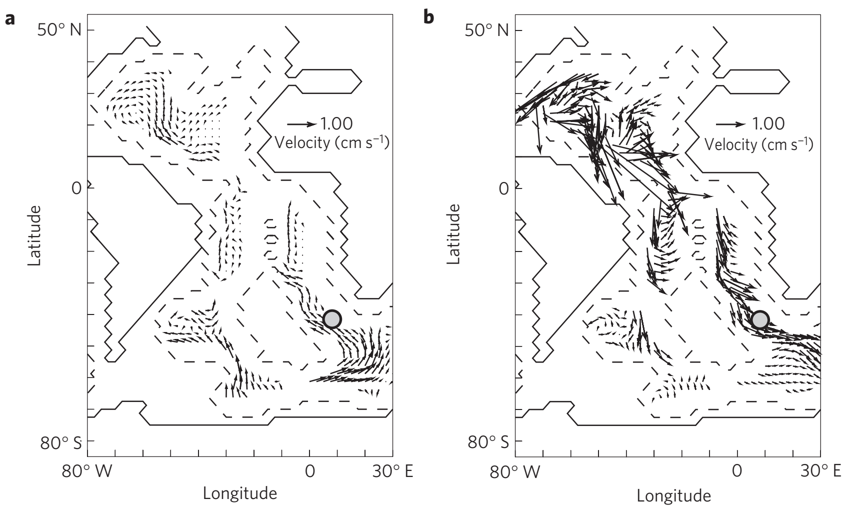

Figure 4 | Deep Atlantic flow-field changes during the HS1/B–A transition. Averaged deep-ocean (>4,000 m) flow-field velocities (10 yr mean) within the Atlantic (with simplified bathymetry) from the re-analysis of a transient deglacial model simulation[13]. **a**, Conditions of particularly weak AMOC (the 'off' mode) during HS1. Here the deep Atlantic Ocean is dominated by northward flow. **b**, Conditions 200 model years later, during the B–A. Now deep waters are dominated by southward flow. The solid line is the coast; the dashed line is the 4,000 m water-depth contour. The grey filled circle represents the location of TNO57-21.

even with the modern 'strong' mode, in which bottom velocities are also northward (not shown). Our reconstructed deep-water ^{14}C ventilation ages for the B–A are of the order of a few hundred years. The observed change in deep-ocean ventilation across the B–A transition is within the range of modelled velocities and the simulated switch from southern- to northern-sourced deep-water domination is in agreement with our proxy reconstructions. In summary, an AMOC overshoot, as described here, may be considered as a transient expansion of the NADW cell in the Atlantic basin with a consequent increase in Atlantic deep-water export.

We note that the results of Liu *et al.*[14], using a fully coupled general circulation model to simulate deglaciation, do not reveal the presence of an unusually deep branch of NADW during the B–A, as suggested by our observations. However, they do imply an overshoot of the AMOC in terms of its overturning rate as well as a deepening of the AMOC cell with respect to glacial conditions. Several other model results (including both transient[13,25] and equilibrium[3] simulations) also show a first-order relationship between overturning strength and depth of the AMOC cell. Although it could be argued that our results and those of Liu *et al.*[14] are mutually exclusive, it should be noted that the modelling results of Knorr and Lohmann[13] (using an ocean model of intermediate complexity) and Liu *et al.*[14] (using a comprehensive atmosphere/ocean/general-circulation model) represent fundamentally different scenarios for explaining the transition from a weak to strong AMOC at the B–A transition. The study of Liu *et al.*[14] suggests that the B–A warming was a transient response of the AMOC to a sudden termination of freshwater discharge to the North Atlantic that is, it was a largely linear response to an abrupt forcing. In contrast, the results of Knorr and Lohmann[13] (and other studies, including that of ref. 26) suggest that a strengthening of the AMOC can be achieved through deglacial changes in the background climate such as gradual global warming[13]; that is, the B–A was a nonlinear response to a gradual forcing, associated with a strong hysteresis behaviour of the AMOC. At this time there is no consensus as to which of these mechanisms is actually at work. The relative importance of freshwater forcing versus climate background forcing on AMOC changes is a matter of ongoing research and the deglacial transition is also complicated by ongoing debate over the magnitude, timing and location of freshwater forcing that might have occurred[27]. The two scenarios put forward by Knorr and Lohmann[13] and Liu *et al.*[14] are inherently different in their forcing and the transitions within each represent two very different dynamical situations. It is open to question to what extent their different responses are a consequence of the different model complexities employed and the forcings applied.

Aside from any model verification, our results alone cannot discriminate between a southward or upward displacement of southern-sourced deep waters by the AMOC cell during the B–A and D–O 8 (they do at least imply a change in the relative densities of northern- versus southern-sourced deep waters). However, newly published results[28] from a shallower core (3.8 km water depth) situated to the southwest of our site suggest that well-ventilated waters were also present at this depth during the B–A. In combination with our study, these results suggest that southern-sourced deep waters may have been displaced southwards rather than upwards by expansion of the AMOC cell during the B–A. Future studies, including model simulations, should provide additional constraints on deep-water mass geometries to further investigate these ideas.

To terminate or not to terminate

Our records of carbonate preservation suggest the presence of better ventilated bottom waters in the South Atlantic Ocean during the B–A and D–O 8. Two key questions arise: why do we not observe similar features following HS2 or HS3 even though a see-saw response is recorded over Antarctica for both of these events? And why did the AMOC recovery during the B–A lead to glacial termination whereas that during D–O 8 did not? Of relevance to the first question is evidence from northeast Brazil that suggests a greater atmospheric impact of AMOC variability during HS1 and HS4 than during HS2 or HS3 (Fig. 2; ref. 29). The interval encompassing HS2 and HS3 was the coldest period of the last glacial cycle and witnessed the greatest extent of continental ice[7]. These extreme glacial conditions are apparent from the record of cold planktonic foraminiferal species from TNO57-21 (Fig. 2). Generally poor carbonate preservation in TNO57-21 is also apparent during this interval, potentially reflecting a generally weaker AMOC. Reconstructions of vertical nutrient ($\delta^{13}C$) distribution within the Atlantic basin also suggest that the transition between well-ventilated (northern-sourced) and poorly ventilated (southern-sourced) deep waters was relatively shallow throughout this interval[30]. In combination with very dense and salty Antarctic Bottom Water filling the abyssal glacial Atlantic Ocean[31], this could have limited the penetration of any overshoots that may have occurred. The interval was also a time of low obliquity (the tilt of Earth's rotation axis; Fig. 2). In contrast, the increased seasonality and sea-ice variability at times of relatively high obliquity may have exaggerated the oscillation between weaker and stronger modes of the AMOC, that is, from HS1 to the B–A and HS4 to D–O 8. We suggest that a combination of extreme glacial conditions and low obliquity may have dampened the ocean/atmospheric response to perturbations of the AMOC during the period containing HS2 and HS3. This dependency on background state may also provide an explanation as to why D–O variability itself was less pervasive during full glacial conditions than during the preceding intermediate glacial period.

In response to the second question, model results suggest that an AMOC overshoot may occur as a result of gradual changes in background climate[13,26], which themselves are partly the product of a weakened circulation (that is, a Heinrich stadial event). Hence, we observe preservation events after episodes of particularly weak AMOC even during a glacial period (Marine Isotope Stage 3). Additional orbital-scale forcing during deglaciation allows the preservation of interglacial-like conditions after the B–A (allowing for the transient Younger Dryas cold event) but subsequent climate deterioration during D–O 8 triggers a return to a glacial mode of circulation. The very similar characteristics of the deep-ocean changes we observe associated with D–O 8 and the B–A suggest that although an extreme deepening of the AMOC may be a necessary condition for glacial termination it is not a sufficient one.

Methods

Radiocarbon measurements were made on mixed benthic foraminifera picked from the fraction >125 μm, and are paired with previously reported ^{14}C measurements on *Globigerina bulloides*[9] (Supplementary Table S1). Planktonic foraminiferal (including whole shell) counts were carried out on 1 cm intervals of sediment with a sampling frequency of 2 cm. Each sample was split to provide >300 individuals in the >150 μm fraction. Planktonic foraminiferal fragment counts were made at the same time on the same sample splits. Warm species are *Globorotalia truncatulinoides* (d), *Globigerinoides ruber*, *Globorotalia hirsuta* and *Orbulina universa*. Cold species are *Neogloboquadrina pachyderma* (s + d) and *Turborotalita quinqueloba*.

Received 21 December 2009; accepted 24 June 2010; published online 25 July 2010

References

1. Ganachaud, A. & Wunsch, C. Improved estimates of global ocean circulation, heat transport and mixing from hydrographic data. *Nature* **408,** 453–457 (2000).
2. Curry, W. B. & Oppo, D. W. Glacial water mass geometry and the distribution of $\delta^{13}C$ of ΣCO_2 in the western Atlantic Ocean. *Paleoceanography* **20,** PA1017 (2005).
3. Weber, S. L. *et al.* The modern and glacial overturning circulation in the Atlantic Ocean in PMIP coupled model simulations. *Clim. Past* **3,** 51–64 (2007).
4. Hays, J. D., Imbrie, J. & Shackleton, N. J. Variations in the Earth's orbit: Pacemaker of the Ice Ages. *Science* **194,** 1121–1132 (1976).
5. Raymo, M. E. The timing of major climate terminations. *Paleoceanography* **12,** 577–585 (1997).
6. Huybers, P. & Wunsch, C. Obliquity pacing of the late Pleistocene glacial terminations. *Nature* **434,** 491–494 (2005).
7. Clark, P. U. *et al.* The last glacial maximum. *Science* **325,** 710–714 (2009).
8. McManus, J. F., Francois, R., Gherardi, J. M., Keigwin, L. D. & Brown-Leger, S. Collapse and rapid resumption of Atlantic meridional circulation linked to deglacial climate changes. *Nature* **428,** 834–837 (2004).
9. Barker, S. *et al.* Interhemispheric Atlantic seesaw response during the last deglaciation. *Nature* **457,** 1097–1102 (2009).
10. Knorr, G. & Lohmann, G. Southern Ocean origin for the resumption of Atlantic thermohaline circulation during deglaciation. *Nature* **424,** 532–536 (2003).
11. Cheng, H. *et al.* Ice age terminations. *Science* **326,** 248–252 (2009).
12. Severinghaus, J. P. & Brook, E. J. Abrupt climate change at the end of the last glacial period inferred from trapped air in polar ice. *Science* **286,** 930–934 (1999).
13. Knorr, G. & Lohmann, G. Rapid transitions in the Atlantic thermohaline circulation triggered by global warming and meltwater during the last deglaciation. *Geochem. Geophys. Geosyst.* **8,** Q12006 (2007).
14. Liu, Z. *et al.* Transient simulation of last deglaciation with a new mechanism for Bølling–Allerød warming. *Science* **325,** 310–314 (2009).
15. Gherardi, J. M. *et al.* Glacial–interglacial circulation changes inferred from Pa-231/Th-230 sedimentary record in the North Atlantic region. *Paleoceanography* **24,** PA2204 (2009).
16. Ninnemann, U. S. & Charles, C. D. Changes in the mode of Southern Ocean circulation over the last glacial cycle revealed by foraminiferal stable isotopic variability. *Earth Planet. Sci. Lett.* **201,** 383–396 (2002).
17. Piotrowski, A. M., Goldstein, S. L., Hemming, S. R. & Fairbanks, R. G. Temporal relationships of carbon cycling and ocean circulation at glacial boundaries. *Science* **307,** 1933–1938 (2005).
18. Vidal, L. *et al.* Evidence for changes in the North Atlantic deep water linked to meltwater surges during the Heinrich events. *Earth Planet. Sci. Lett.* **146,** 13–27 (1997).
19. Robinson, L. F. *et al.* Radiocarbon variability in the western North Atlantic during the last deglaciation. *Science* **310,** 1469–1473 (2005).
20. Skinner, L. C. & Shackleton, N. J. Rapid transient changes in northeast Atlantic deep water ventilation age across Termination I. *Paleoceanography* **19,** PA2005 (2004).
21. Anderson, R. F. *et al.* Wind-driven upwelling in the southern ocean and the deglacial rise in atmospheric CO_2. *Science* **323,** 1443–1448 (2009).
22. Marchitto, T. M., Lehman, S. J., Ortiz, J. D., Fluckiger, J. & van Geen, A. Marine radiocarbon evidence for the mechanism of deglacial atmospheric CO_2 rise. *Science* **316,** 1456–1459 (2007).
23. Broecker, W. S. & Barker, S. A 190‰ drop in atmospheric $\Delta^{14}C$ during the 'mystery interval' (17.5 to 14.5 kyrs). *Earth Planet. Sci. Lett.* **256,** 90–99 (2007).
24. Maier-Reimer, E., Mikolajewicz, U. & Hasselmann, K. Mean circulation of the Hamburg LSG OGCM and its sensitivity to the thermohaline surface forcing. *J. Phys. Oceanogr.* **23,** 731–757 (1993).
25. Ganopolski, A. & Rahmstorf, S. Rapid changes of glacial climate simulated in a coupled climate model. *Nature* **409,** 153–158 (2001).
26. Ganopolski, A. & Roche, D. M. On the nature of lead-lag relationships during glacial–interglacial climate transitions. *Quat. Sci. Rev.* **28,** 3361–3378 (2009).
27. Thornalley, D. J. R., McCave, I. N. & Elderfield, H. Freshwater input and abrupt deglacial climate change in the North Atlantic. *Paleoceanography* **25,** PA1201 (2010).
28. Skinner, L. C., Fallon, S., Waelbroeck, C., Michel, E. & Barker, S. Ventilation of the deep Southern Ocean and deglacial CO_2 rise. *Science* **328,** 1147–1151 (2010).
29. Wang, X. F. *et al.* Wet periods in northeastern Brazil over the past 210 kyr linked to distant climate anomalies. *Nature* **432,** 740–743 (2004).
30. Curry, W. B., Marchitto, T. M., McManus, J. F., Oppo, D. W. & Laarkamp, K. L. in *Mechanisms of Global Climate Change at Millennial Time Scales* (eds Clark, P. U., Webb, R. S. & Keigwin, L. D.) (Am. Geophys. Union, 1999).
31. Adkins, J. F., McIntyre, K. & Schrag, D. P. The salinity, temperature, and $\delta^{18}O$ of the glacial deep ocean. *Science* **298,** 1769–1773 (2002).
32. NGRIP members, High-resolution record of Northern Hemisphere climate extending into the last interglacial period. *Nature* **431**, 147–151 (2004).
33. Key, R. M. *et al.* A global ocean carbon climatology: Results from Global Data Analysis Project (GLODAP). *Glob. Biogeochem. Cycles* **18,** GB4031 (2004).
34. EPICA, One-to-one coupling of glacial climate variability in Greenland and Antarctica. *Nature* **444**, 195–198 (2006).
35. Berger, A. & Loutre, M. F. Insolation values for the climate of the last 10 million years. *Quat. Sci. Rev.* **10,** 297–317 (1991).

Acknowledgements

We thank T. Van de Flierdt, D. Thornalley, P. Sexton and W. Broecker for useful discussions and H. Medley for help in the laboratory. Sample material used in this project was provided by the Lamont-Doherty Earth Observatory Deep-Sea Sample Repository. We thank R. Lotty and G. Lozefski for their help with sampling. Support for the collection and curating facilities of the core collection is provided by the National Science Foundation through Grant OCE00-02380 and the Office of Naval Research through Grant N00014-02-1-0073. The work was supported by a National Science Foundation (US) grant (OCE -0435703) and Natural Environment Research Council (UK) grants NE/F002734/1 and NE/G004021/1.

Author contributions

S.B. designed and managed the project, G.K. carried out model analysis, S.B. and G.K. developed interpretation with help from all authors. M.J.V. carried out foraminiferal counts, P.D. picked and prepared benthic foraminifera for ^{14}C dating. All authors contributed to writing the manuscript.

Additional information

The authors declare no competing financial interests. Supplementary information accompanies this paper on www.nature.com/naturegeoscience. Reprints and permissions information is available online at http://npg.nature.com/reprintsandpermissions. Correspondence and requests for materials should be addressed to S.B.