Evidence from Southern Ocean sediments for the effect of North Atlantic deep-water flux on climate

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The Southern Ocean is perhaps the only region where fluctuations in the global influence of North Atlantic Deep Water (NADW) can be monitored unambiguously in single deep-sea cores. A carbon isotope record from benthic foraminifera in a Southern Ocean core reveals large and rapid changes in the flux of NADW during the last deglaciation, and an abrupt increase in the NADW production rate which immediately preceded large-scale melting of the Northern Hemisphere ice sheets. This sudden strengthening of the NADW thermohaline cell provides strong evidence for the importance of NADW in glacial-interglacial climate change.

FUNDAMENTAL aspects of the Earth's climate variability over the last deglaciation indicate the operation of strong internal feedbacks and amplifiers in the global climate system. The rapidity and amplitude of the glacial-interglacial transition[1-4], the pulsed destruction of the Northern Hemisphere ice sheets[4], and the interhemispheric synchroneity of change[5,6] are all difficult to explain by solar insolation changes alone. Two mechanisms often invoked to explain these phenomena are variations in greenhouse gas concentrations and poleward oceanic heat transport (NADW formation)[2,7-11]. The basic principles of how both these amplifiers might work are well established (see ref. 12 for a review and further references), yet rigorous geological assessment of how either was actually involved with the last deglaciation has not been possible.

Precise radiocarbon dating of the Barbados sea-level record, which is marked by two intervals of massive ice-sheet melting centred on 12.0 kyr and 9.5 kyr (ref. 4), now provides an explicit geological test for these proposed climate amplifiers. Comparable chronologies for atmospheric and deep-ocean variability over the last deglaciation are clearly required. Existing deglacial ice-core records of greenhouse gases suffer either from summer melting artefacts[13] or dating uncertainties[14] and are therefore not appropriate for comparison to the sea-level record. Similarly, no clear consensus on the timing of deep-water variability has emerged[4,12,15,16] and the evidence for deglacial deep-water fluctuations has thus far been limited to North Atlantic records[15,16] which could be overprinted by local deep-water mixing effects[16,17].

Our approach to this problem has been to focus on Southern Ocean deep-sea sediment cores. The nutrient chemistry of the Southern Ocean is strongly influenced by the influx of NADW, and the global-scale mixing of deep waters that occurs in the Antarctic circumpolar current provides a filter for local variability. Thus, Southern Ocean palaeonutrient records can, in principle, provide a suitable deep-ocean counterpart to the Barbados coral record of ice-sheet decay. Here we present a radiocarbon-dated $\delta^{13}C$ record from benthic foraminifera in a Southern Ocean deep-sea core, with resolution detailed enough to test the interactions between the Northern Hemisphere ice sheets and deep-ocean circulation over the last deglaciation.

NADW and circumpolar $\delta^{13}C$

Figure 1 illustrates why the $\delta^{13}C$ of circumpolar water is sensitive to the variations in the contribution of NADW. Circumpolar Deep Water (CPDW) represents a mixture of NADW and recirculated deep water from the Indian and Pacific. Consequently, CPDW $\delta^{13}C$ values lie between those of NADW (characterized by high $\delta^{13}C$ and low nutrients) and Indo-Pacific Deep Water (low $\delta^{13}C$ and high nutrients). Without any NADW influence, the $\delta^{13}C$ of CPDW should decrease to Indo-Pacific values, regardless of how the oceanic inventory of ^{13}C or ^{12}C might have changed. Comparison of North Atlantic, Pacific and Southern Ocean $\delta^{13}C$ records from benthic foraminifera confirmed this concept[17], highlighting a primary advantage of using CPDW $\delta^{13}C$ records for documenting the history of NADW: any uncertainties involving the mixing ratio of northern and southern sources are circumvented. On the other hand, the problem with using North Atlantic $\delta^{13}C$ records is that the measured changes may simply reflect shifts in the mixing zone between Antarctic Bottom Water (AABW) and NADW, with no relationship to net NADW flux. Thus, although it may be less sensitive to subtle changes in NADW strength because it is 'downstream' of the source, the Southern Ocean is perhaps the only region where the global influence of NADW fluctuations can be monitored unambiguously in single locations. This strategic advantage is important, because deep-sea cores will never begin to capture the three-dimensional structure of the deep ocean in the past.

It is possible, however, that deviations from simple mixing models for CPDW $\delta^{13}C$ can occur. For example, there is growing evidence that $\delta^{13}C$ in Southern Ocean foraminifera was lower (by ~0.2‰) than in those from the deep Pacific during the glacial periods[18], and this distribution of $\delta^{13}C$ cannot be easily explained by circulation alone. Nonconservative effects involving the ecology of benthic foraminifera[19], the varying degrees of isotopic equilibrium reached by incorporated surface waters[20] and the degradation of specific organic matter[21] are all possible. Although the first of these effects must be assessed empirically, mass balance considerations suggest that the latter two effects are small. As the glacial and interglacial amplitude of foraminiferal $\delta^{13}C$ in Southern Ocean cores is ~1‰ (as is the present Atlantic-Pacific $\delta^{13}C$ gradient in total inorganic carbon), nonconservative effects of 0.2‰ do not preclude using $\delta^{13}C$ to deduce large-scale changes in water mass circulation. A more serious challenge is the observation that Cd/Ca ratios in benthic foraminifera (another palaeonutrient proxy) from Southern Ocean cores show little glacial-interglacial change[21]. We have no definitive explanation for this apparent discrepancy, although recent data indicate that foraminiferal distribution coefficients for Cd may vary with depth by a factor of two[22]. In any case, we make the simplest interpretation that seems justified from the modern distribution of $\delta^{13}C$ in deep waters and foraminifera: a strong relative contribution of NADW is the only means for increasing CPDW $\delta^{13}C$ above Pacific (mean ocean) values.

Changes over the last deglaciation

Core RC11-83, recovered on the north flank of the Cape rise (41° 36′ S, 94° 48′ E; 4,718 m) was overlain by deep waters indistinguishable from Lower Circumpolar Deep Water. The core is characterized by sedimentation rates of roughly 25 cm kyr^{-1}; consequently, its benthic records spanning the past 15 kyr represent as yet the most detailed chronicle of deglacial isotopic change in the deep Southern Ocean. We generated stable isotope records for *Planulina wuellerstorfi*, whose shells seem to be relatively reliable recorders of the isotopic composition of the ambient water[18,19].

The deglacial timescale for this record is derived from accelerator mass spectrometric (AMS) ^{14}C dates on seven levels in the core (listed in Table 1; ages given throughout the text refer to ^{14}C kyr BP) and a polynomial fit of the resulting age-depth curve. The ages obtained from *Globigerina bulloides* have been corrected by a constant 600 yr, to account for the fact that the waters in which these foraminifera grow are not in isotopic equilibrium with the atmosphere. This constant correction is likely to be a reasonable estimate of the reservoir age (within 200 yrs) throughout deglaciation[23]. The $\delta^{18}O$ time series derived from this age model (not shown) is in excellent agreement with previously published deglacial $\delta^{18}O$ curves from the Southern Ocean[24].

The RC11-83 benthic $\delta^{13}C$ record (Figs 1 and 2) demonstrates that deglacial changes in the nutrient chemistry of the deep Southern Ocean were rapid and of considerable magnitude. At 16 kyr, the $\delta^{13}C$ values are the same as[25], or slightly lower than[18,26] analogous glacial deep Pacific values, confirming earlier findings of the reduced influence of NADW in the Southern Ocean during the last glacial maximum[17]. But an abrupt shift from low values characteristic of glacial conditions to high, essentially modern, values occurs from 12.6–12.2 kyr. Following the premise described in the previous section, this rapid rise in Southern Ocean $\delta^{13}C$ values must have been associated with the 'turn on' of NADW. As the duration of this abrupt $\delta^{13}C$ transition (400 years) is of the same order as the mixing time of the Atlantic ocean[27], the change from low to high relative NADW flux appears virtually instantaneous from a geological standpoint. The RC11-83 record also shows significant oscillations superimposed on this basic glacial–interglacial change. An excursion towards higher values occurs between 13.1 and 12.7 kyr, and a low-$\delta^{13}C$ event occurs at 11.5 kyr.

Comparison with another detailed deglacial record[16] from the deep North Atlantic (Fig. 1) indicates that $\delta^{13}C$ changes in RC11-83 can be attributed directly to the varying contribution of NADW in the Southern Ocean. The striking transition from nutrient enrichment (^{13}C depletion) to nutrient depletion (^{13}C enrichment), which is dated between 12.6 and 12.2 kyr in RC11-83, is also obvious in the North Atlantic core. Another step in the same direction, but with lower amplitude, is apparent from ~10–9 kyr. In addition, the North Atlantic record shows an anomaly similar to, and perhaps correlated with, the excursion

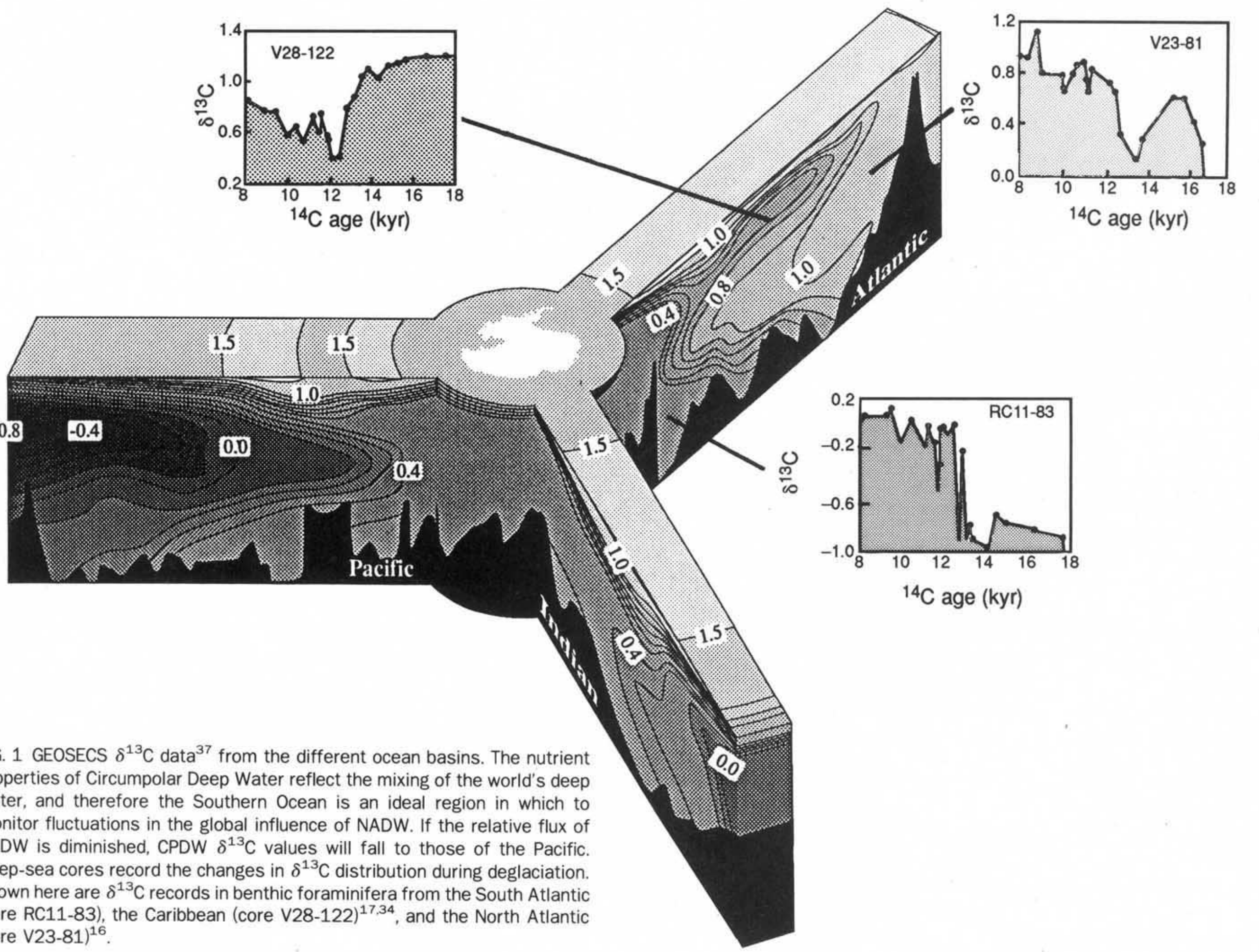

FIG. 1 GEOSECS $\delta^{13}C$ data[37] from the different ocean basins. The nutrient properties of Circumpolar Deep Water reflect the mixing of the world's deep water, and therefore the Southern Ocean is an ideal region in which to monitor fluctuations in the global influence of NADW. If the relative flux of NADW is diminished, CPDW $\delta^{13}C$ values will fall to those of the Pacific. Deep-sea cores record the changes in $\delta^{13}C$ distribution during deglaciation. Shown here are $\delta^{13}C$ records in benthic foraminifera from the South Atlantic (core RC11-83), the Caribbean (core V28-122)[17,34], and the North Atlantic (core V23-81)[16].

TABLE 1 Data for core RC11-83

Depth (cm)	Stable isotopes (Lamont-Doherty) corrected* Age (^{14}C kyr)	δ^{18}O (‰ PDB) *P. wuellerstorfi*	δ^{13}C (‰ PDB) *P. wuellerstorfi*	Accelerator radiocarbon ages (NSF–Arizona accelerator facility, Tucson) Accession number	Age *G. bulloides*
165	7.57	2.39	−0.03		
170	7.92	2.41	0.07		
175	8.27	2.64	0.13		
189	9.38	2.66	0.14	5,974	9.90 ± 0.066
189				5,974	10.08 ± 0.090
191	9.53	2.60	0.19		
197	9.97	2.54	−0.07		
205	10.48	2.75	0.10		
215	11.03	3.09	−0.09		
217	11.12	2.94	−0.11		
220	11.27	2.64	0.06		
221				5,975	11.94 ± 0.08
221				5,975	11.93 ± 0.10
225	11.49	2.71	−0.08		
228	11.61	3.02	−0.07		
231				5,976	12.04 ± 0.08
231				5,976	12.19 ± 0.12
232	11.77	3.24	−0.47		
233	11.80	3.22	−0.25		
235	11.87	3.26	0.03		
238	11.98	3.35	0.04		
245	12.20	3.38	−0.02		
248				5,977	12.79 ± 0.11
248				5,977	13.02 ± 0.09
255	12.49	3.44	0.07		
260	12.62	3.35	−0.62		
262	12.67			5,978	13.23 ± 0.11
265	12.75	3.38	−0.59		
270	12.87	3.29	−0.86		
275	13.00	3.44	−0.14		
278	13.07	3.56	−0.30		
285	13.24	3.76	−0.85		
288	13.32	3.60	−0.72		
295	13.50	3.81	−0.83		
298	13.59			5,979	14.31 ± 0.11
315	14.13	3.81	−0.91		
325	14.53	3.87	−0.65		
335	15.00	3.93	−0.70	5,980	15.62 ± 0.12
365	17.02	3.91	−0.74		
395	20.24	3.98	−0.81		

* Age adjusted by 600 yr to correct for apparent age of surface seawater[23]. $\delta^{18}O = [(^{18}O/^{16}O)_{sample}/(^{18}O/^{16}O)_{standard}] - 1$ in ‰; $\delta^{13}C = [(^{13}C/^{12}C)_{sample}/(^{13}C/^{12}C)_{standard}] - 1$.

at 11.5 kyr in RC11-83. The records also agree in the fact that the Younger Dryas interval (11–10 kyr) is not characterized by a particularly strong anomaly. Any similarities observed in both the North and South Atlantic must reflect the varying influence of NADW during deglaciation because the southward flow of nutrient-depleted NADW repressed the northward flow of nutrient-rich Southern Ocean water to the deep North Atlantic.

Where there is disagreement between the various nutrient proxy records for the Atlantic, we argue that the Southern Ocean would yield the more comprehensive measure of NADW flux variability. This issue is important for the interval from 16 to 13.5 kyr, because the North Atlantic record shows a deep minimum centred on 13.5 kyr, suggesting that the Last Glacial Maximum was not the time of lowest NADW production. The RC11-83 record, in contrast, shows more uniformly low values until 13 kyr, suggesting minimal NADW flux throughout the glacial period. Resolving such differences of interpretation is obviously crucial for an accurate evaluation of NADW's influence in global climate.

Another source of evidence for a profound, rapid deglacial change in Atlantic thermohaline circulation is provided by δ^{13}C records from Caribbean deep-sea cores, which monitor mid-depth North Atlantic water (1,800 m)[28]. In these records, however, the sense of change is opposite to that of the deep Atlantic[15,17]. Although relative nutrient depletion prevailed during glacial periods, a shift to more nutrient-rich conditions occurred at 12.6 kyr, precisely the time of deep-ocean change (Figs.1 and 2). The inverse mid-depth and deep Atlantic signals could both reflect the same phenomenon if relatively nutrient-rich water from the Southern Ocean were drawn northwards across the Equator at mid-depths by the removal of surface and thermocline water to form southward-flowing NADW. This possibility is completely compatible with Boyle and Keigwin's original explanation for the Caribbean signal, that NADW resided at shallower depths during glacial periods[16], and with the glacial Atlantic reconstruction of Duplessy *et al.*[26]. In any case, the chemistry of the Atlantic over a large depth range was apparently affected by the abrupt deglacial changes in the NADW cell.

Figure 2 shows that the 'turn on' of NADW deduced from the convergence of the Atlantic nutrient records was nearly synchronous with, or immediately preceded, the first significant pulse of Northern Hemisphere ice-sheet melting measured by the Barbados sea-level curve. The ocean circulation changes also seem to be concomitant with the first signs of warmth in European continental pollen assemblages from lake sediments[29].

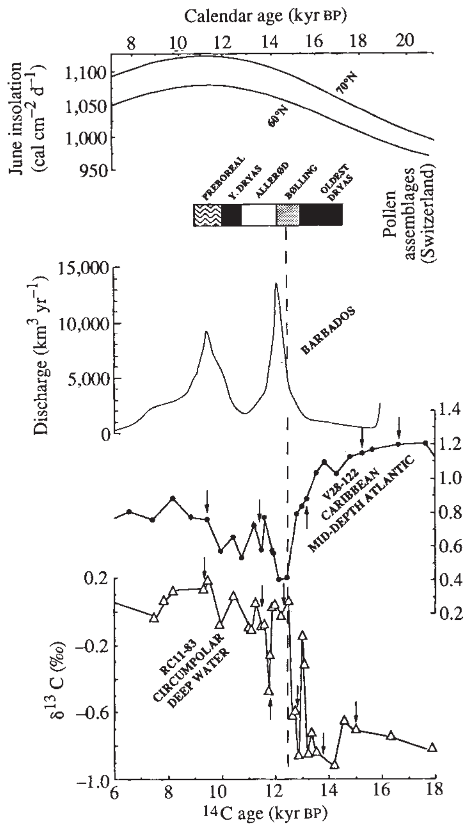

FIG. 2 Convergence of deep Southern Ocean and Caribbean[17,34] δ^{13}C records, with AMS ^{14}C dated levels indicated by arrows, marks the 'turn on' of NADW (dashed line). The timing of this event can be compared with the ice-sheet meltwater discharge curve (calculated as the first derivative of the Barbados sea-level curve[4,38]), the change in pollen assemblages in Swiss lake sediments (schematic zones taken from ref. 29), and June insolation at 60° N and 70° N (ref. 38), provided that a common timescale is established. The top scale is calendar years, and the bottom scale is ^{14}C years. Although the ocean records and lake sediment records are dated by ^{14}C, they can be compared to records dated in calendar years (insolation) by using the calibration in ref. 30.

The records of the different climate variables, correlated on the basis of detailed AMS ^{14}C dating, may be placed in the context of solar insolation changes through the Barbados coral Th/U calibration of the radiocarbon timescale[30].

Discussion

Although there is no simple astronomical explanation for the shape of the glacial meltwater curve[4], coincidence between the abrupt ocean circulation changes and the pulses of ice-sheet melting strongly suggests that NADW variability was an important factor in the shift to a deglaciated state. The Southern Ocean isotopic record might be taken as evidence for the suggestion that changes in the NADW cell promoted a rapid, global reorganization of the climate system to an interglacial mode of operation[12]. But although the renewal of NADW may seem a simple and chronologically consistent explanation for the meltwater pulses, the timing alone does not constitute proof of this hypothesis.

The questions to be asked are, first, what might have caused the NADW flux changes, and second, whether the heat transport associated with renewed NADW could have provided the trigger for ice-sheet melting. Broecker *et al.* suggest that when Northern Hemisphere ice sheets grew to their maximum extent, the strength of the NADW cell was poised to oscillate, because the salt balance of the North Atlantic involved a precarious interplay of glacial meltwater, ocean advection and water vapour loss[31]. The excursions present in the RC11-83 $\delta^{13}C$ record from 12.6 to 11.5 kyr might conform to the model of ref. 31: NADW 'turns on', causing ice to melt, which in turn causes NADW production to slow. But the oscillation from 13.1 to 12.6 kyr in the RC11-83 record, which may be considered a 'false start' towards prolonged deep-water formation, is not well resolved in the meltwater curve, and for the salt oscillator model to apply, another salt pathway (such as sea-ice growth in the Norwegian-Greenland or Labrador Seas, or oceanic advection) must have played a stronger part at this time. Therefore, at face value, the RC11-83 record allows for some independence between the deep ocean and the ice sheets.

The immediate effect of North Atlantic heat transports on ice-sheet melting is not entirely obvious. General circulation models clearly demonstrate that continental Europe would be the main beneficiary of any increased oceanic heat drawn poleward by NADW formation[32]. Thus, the correlation between changes in European pollen assemblage and NADW flux changes measured in the Southern Ocean is not surprising (although the Younger Dryas interval, where the Southern Ocean record shows little change, is an example of how the coupling between the two was not always exact). Yet deep-sea core evidence[33,34] and the mapping of meltwater plumes[35] suggest that the Laurentide (North American) ice sheet was the primary contributor to the first main meltwater pulse. If NADW formation brought warmer water to the Labrador Sea and Baffin Bay, melting along the northeast margin of the Laurentide could be explained, but the southern margin melting poses a problem for this mechanism and perhaps should be a subject for further modelling.

Of course, the 'turn on' of NADW could indirectly affect the global heat budget by altering atmospheric CO_2 concentrations. The massive redistribution of nutrients at 12.6 kyr, as evidenced by the Atlantic records, must have affected the ocean's capacity to hold CO_2, though the deglacial rate of CO_2 increase is still a matter for debate. The abrupt shift in NADW production, with a consequent increase in heat transport, may actually be responsible for the first deglacial rise in the CO_2 record in the Greenland ice cores, which has been reinterpreted as an artefact of summer melting[13]. New results from the Byrd core (Antarctica)[13] suggest that the CO_2 response to ocean circulation and nutrient changes was gradual and thus could not have been the immediate cause of the meltwater pulses. Until an unequivocal CO_2 record is produced, we prefer to emphasize the possible direct consequences of NADW formation on global climate change.

The Southern Ocean $\delta^{13}C$ record therefore demonstrates that large, rapid fluctuations in the production of NADW were felt outside the North Atlantic basin over the last deglaciation. The detailed dating of this record refines previous estimates[12] of 14 kyr for the re-initiation time of NADW to 12.6–12.2 kyr, an interval virtually coincident with the first melting episode of the Northern Hemisphere ice sheets[4]. The record thus provides strong evidence that NADW could have been the primary amplifier of glacial–interglacial climate change in high latitudes. Enigmatic low- to mid-latitude climate phenomena, such as the synchronous shifts of mountain snowlines and floral assemblages in both hemispheres, are left to be explained by other means, possibly CO_2-induced changes in plant survival or local climate patterns[36]. □

Received 23 July; accepted 17 December 1991.

1. Broecker, W. S. & van Donk, J. *Rev. Geophys. Space Phys.* **8**, 167–198 (1970).
2. Broecker, W. S. in *Milankovich and Climate* (eds. Berger, A. *et al.*) 687–698 (Reidel, Dordrecht, 1984).
3. Broecker, W. S., Peteet, D. M. & Rind, D. *Nature* **315**, 21–25 (1985).
4. Fairbanks, R. G. *Nature* **342**, 637–642 (1989).
5. Mercer, J. H. in *Climate Processes and Climate Sensitivity, AGU Geophys. Monogr. Ser.* Vol. 29 (eds Hansen, J. E. & Takahashi, T.) 307–313 (1984).
6. Nelson, C. S., Hendy, C. H., Jarrett, G. R. & Cuthbertson, A. M. *Nature* **318**, 361–363 (1985).
7. Oeschger, H. *et al.* in *Climate Processes and Climate Sensitivity, AGU Geophys. Monogr. Ser.* Vol. 29 (Hansen, J. E. & Takahashi, T.) 299–306. (1984).
8. Lorius, C., Jouzel, J., Raynaud, D., Hansen, J. & Le Treut, H. *Nature* **347**: 139–145 (1990).
9. Weyl, P. *Met. Monogr.* **8**, 37–62 (1968).
10. Rooth, C. *Prog. Oceanogr.* **11**, 131–149 (1982).
11. Manabe, S. & Stouffer, R. J. *J. Clim.* **1**, 841–849 (1988).
12. Broecker, W. S. & Denton, G. H. *Geochim. cosmochim. Acta* **53**, 2465–2501 (1989).
13. Staffelbach, T., Stauffer, B., Sigg, A. & Oeschger, H. *Tellus* **43**, 91–96 (1991).
14. Neftel, A., Oeschger, H., Staffelbach, T. & Stauffer, B. *Nature* **331**, 609–611 (1988).
15. Boyle, E. A. & Keigwin, L. *Nature* **330**, 35–40 (1987).
16. Jansen, E. & Veum, T. *Nature* **343**, 612–616 (1990).
17. Oppo, D. W. & Fairbanks, R. G. *Earth planet. Sci. Lett.* **86**, 1–15 (1987).
18. Curry, W. B., Duplessy, J. C., Labeyrie, L. D. & Shackleton, N. J. *Paleoceanography* **3**, 317–342 (1988).
19. Belanger, P. E., Curry, W. B. & Matthews, R. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* **33**, 205–220 (1981).
20. Charles, C. D. & Fairbanks, R. G. in *The Geologic History of the Polar Oceans: Arctic vs Antarctic* (eds Bleil, U. & Thiede, J.) 519–538 (1990).
21. Boyle, E. A. *Eos* **71**, 1357–1358 (1990).
22. Boyle, E. A. & Rosenthal, Y. *Eos* **72**, 265 (1991).
23. Bard, E. *Paleoceanography* **3**, 636–645 (1988).
24. Labracherie, M. *et al. Paleoceanography* **4**, 629–638 (1989).
25. Zahn, R., Pedersen, T. F., Bornhold, B. D. & Mix, A. C. *Paleoceanography* **6**, 543–560 (1991).
26. Duplessy, J.-C. *et al. Paleoceanography* **3**, 343–360 (1988).
27. Broecker, W. S. *J. geophys. Res.* **84**, 3218–3226 (1979).
28. Wüst, G., *Stratification and Circulation of the Antillean-Caribbean Basins* (Columbia University Press, New York, 1964).
29. Ammann, B. & Lotter, A. F. *Boreas* **18**, 109–126
30. Bard, E., Hamelin, B., Fairbanks, R. G. & Zindler, A. *Nature* **345**, 405–410 (1990).
31. Broecker, W. S., Bond, G., Klas, M., Bonani, G. & Wolfli, W. *Paleoceanography* **5**, 469–478 (1990).
32. Rind, D., Peteet, D., Broecker, W. S., McIntyre, A. & Ruddiman, W. *Clim. Dynam.* **1**, 3–33 (1986).
33. Leventer, A., Williams, D. F. & Kennett, J. P. *Earth planet. Sci. Lett.* **59**, 11–17 (1982).
34. Broecker, W. S. *et al. Radiocarbon* **32**, 119–133 (1990).
35. Fairbanks, R. G., Charles, C. D. & Wright, J. W. in *Four Decades of Radiocarbon: An Interdisciplinary Approach* (eds Long A. & Kra, R.) 473–500 (Springer, 1992).
36. Van De Water, P. *Geol. Soc. Am. Abstr. Prog.*, 352 (1991)
37. Kroopnick, P. *Deep Sea Res.* **32**, 57–84 (1985).
38. Fairbanks, R. G. *Paleoceanography* **6**, 937–948 (1990).

ACKNOWLEDGEMENTS. We thank T. M. Jull and D. Donohue at the Arizona Accelerator Facility for collaboration, and W. Broecker, R. Mortlock, D. Peteet, N. Slowey and J. Wright for discussions. This research was funded by NSF.