The role of the deep ocean in North Atlantic climate change between 70 and 130 kyr ago

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THE suggestion[1] that changes in North Atlantic Deep Water (NADW) production are linked through surface heat flux to the atmospheric temperature over Greenland is supported by earlier indications[2,3] that NADW production decreased during glacial times, and by the subsequent finding[4–6] that it declined during the Younger Dryas cool period at the end of the last glaciation. Changes in North Atlantic surface temperatures have been found[7] to mirror high-frequency temperature changes recorded in Greenland ice cores over the past 80 kyr, but the connection to abyssal circulation has yet to be established, except for one or two isolated oscillations[8,9]. Here we present carbon and oxygen isotope analyses of benthic foraminifera in a high-resolution North Atlantic deep-sea sediment core for the period 70–130 kyr ago. These data allow us to reconstruct the history of NADW production, which shows a close correlation with Greenland climate variability for much of this time interval, suggesting that the climate influence of NADW variability was widespread. We see no evidence, however, for changes in NADW production during substage 5e (the Eemian interglacial period), in contrast with recent ice-core data[10] which suggest severe climate instability in Greenland during this time period. Our results may support suggestions, based on data from a second ice core, that this apparent instability is an artefact caused by ice flow[11]. Alternatively, the Eemian climate instability may have had a different origin from the subsequent climate events.

We chose piston core Knorr 31 GPC9 for a detailed time-series study. This long, large-diameter core was recovered from a depth of 4,758 m in a sediment wave field on the west flank of the Bahama Outer Ridge. The ridge is a constructional deposit, moulded by deep contour-following currents[12]; modern bottom waters at the core location contain ~15% recirculated Antarctic Bottom Water (AABW) mixed with NADW[13]. Because of its western boundary location, GPC9 is ideally positioned to monitor the chemistry of waters exported from the Atlantic basin, and its water depth makes it very sensitive to changes in the balance between NADW and AABW in the deep North Atlantic.

Oxygen isotope analyses of the benthic foraminifera *Cibicidoides* spp. and *Nuttalides umbonifera* provide a standard sequence of maxima and minima which are readily identified as the substages of isotope stage 5 (Fig. 1*a*). Eight levels have been assigned ages according to an orbitally-tuned and globally-stacked age model[14]. Average rates of sedimentation derived from this age model for stage 5 at GPC9 are 5.6 cm kyr^{-1}, although rates of sedimentation may be much higher during episodes of harsher climate[9]. Between ~135 and 125 kyr ago δ^{18}O values decreased by ~2‰ (Fig. 2*a*), reflecting both the decrease in continental ice volume which produced a high stand of sea level during isotope substage 5e and the warming of deep waters in the Atlantic Ocean[15]. The substage 5e interglaciation, which has been previously correlated to the Eemian on land[16], lasted ~10 kyr and was followed by falling sea level and two intervals of equable climate (substages 5c and 5a) interrupted by two intervals of harsher climate (substages 5d and 5b).

In general, δ^{13}C results anticorrelate with δ^{18}O (Figs 1*b*, 2*b*). During peak interglacial (substage 5e) and interstadial (substages 5a and 5c) conditions, maximum δ^{13}C is achieved due to a combination of increased terrestrial biomass[17,18] (which might account for about a third of the signal in GPC9) and increased production of NADW. Today, NADW typically has δ^{13}C values of ~1‰ (ref. 19), although no data exist from near core GPC9 where the influence of AABW would be expected to lower the values. Thus, the values of 1‰ observed during substage 5e support the interpretation of Duplessy and Shackleton[20] that NADW production was then near modern levels. Values during glacial stage 6 and leading into glacial stage 4 were as low as −0.5‰, equivalent to δ^{13}C in the Southern Ocean during the maximum of the last glaciation[18,21]. This implies that the deep western North Atlantic experienced a substantial decrease in the ratio of NADW to AABW during glacial stages 4 and 6.

Our most significant new finding in GPC9 is that δ^{13}C results display variability at significantly higher frequencies than seen

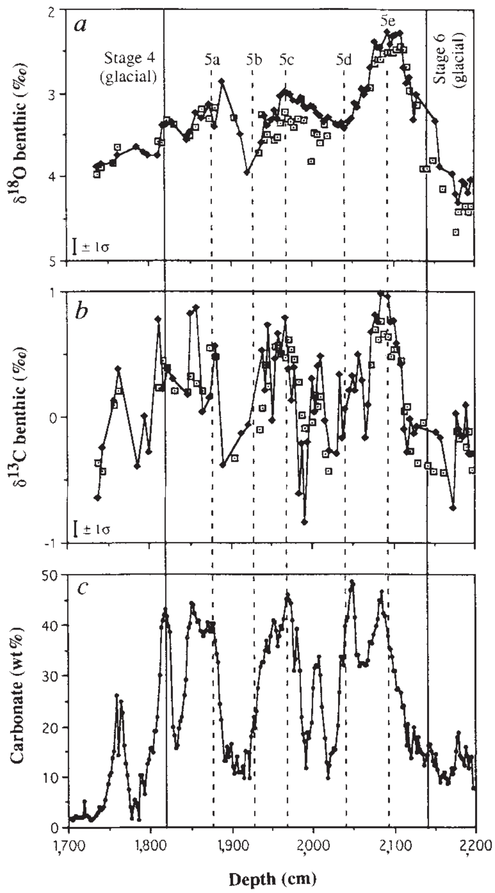

FIG. 1 Oxygen isotope (*a*), carbon isotope (*b*) and wt% $CaCO_3$ (*c*) from western North Atlantic core KNR31-GPC9 (28° 14.7′ N, 74° 26.4′ W, 4,758 m). Stable isotope ratios were measured on the benthic foraminifera *Cibicidoides* spp. (solid symbols) and on *N. umbonifera* (open symbols) with an analytical uncertainty (in standard analyses) represented by the small bar in the lower left corner of panels *a* and *b*. For δ^{13}C data, the signal-to-noise ratio is 15:1. Each displays a similar pattern of changing continental ice volume and deep ocean temperature, which is used to assign the boundaries of isotope stages (solid vertical lines) and the position of stage 5 extrema (dashed vertical lines; *a*). These foraminifera also reveal the same general pattern of high δ^{13}C during warm substages 5a, c and e, with lower values during cooler substages 5b and d (*b*). Ignoring occasional one-point peaks, note that $CaCO_3$ and δ^{13}C results bear a striking resemblance, reflecting the common origin of these signals in deep ocean circulation patterns.

in $\delta^{18}O$ (Fig. 2). The brevity of the $\delta^{13}C$ oscillations rules out forcing by the ice-sheet-driven, continental biomass effects suggested by Shackleton[17] because significant continental ice accumulation takes many millenia. Instead, the millennial-scale variability is probably dominated by abrupt and oscillating changes in the flux of NADW. Significant $\delta^{13}C$ changes occur without regard for the background state of the climate system; high production rates of NADW occur during both interstadial and stadial conditions. For example, although $\delta^{13}C$ initially declined early in glacial substage 5d as is typical at other locations[20,22], it was quickly followed by a prominent $\delta^{13}C$ maximum ~113 kyr ago, and then another minimum (Fig. 2*b*). Conversely, ~103 kyr ago in interstadial substage 5c there was a $\delta^{13}C$ minimum which indicates NADW suppression comparable to full glacial times. In all, there appear to be three maxima and minima in $\delta^{13}C$ for each precession cycle seen in $\delta^{18}O$.

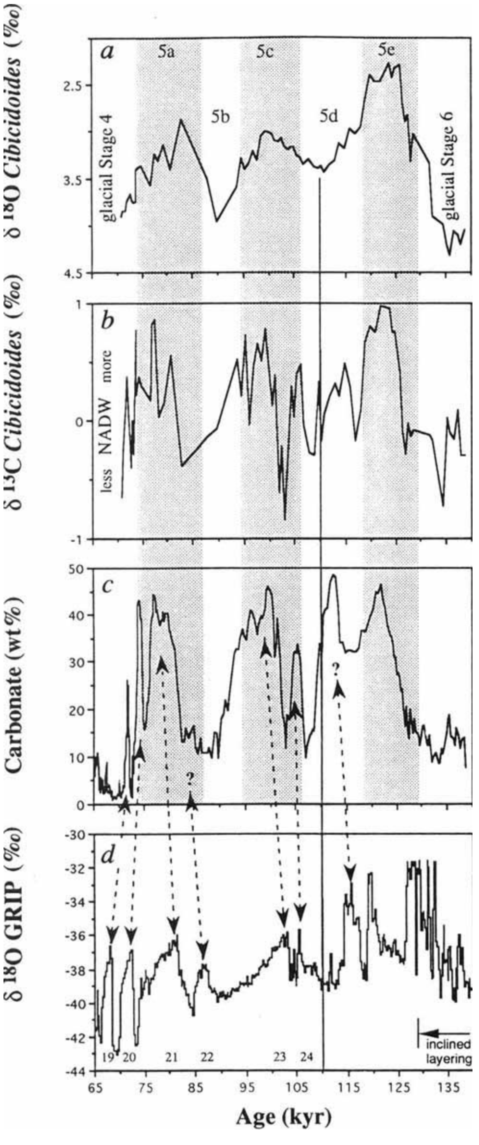

FIG. 2 *Cibicidoides* stable isotope data (*a*, *b*) and wt% $CaCO_3$ (*c*) from Fig. 1 compared to the $\delta^{18}O$ of the GRIP ice core from Summit, Greenland (*d*). GPC9 data are plotted versus age using $\delta^{18}O$ stratigraphy (Fig. 1*a*) and the Martinson *et al.*[14] chronology. Warm interglacial substages in the sediment core are shaded, and interstadials in the ice core are numbered. The ice-core chronology was discussed by Dansgaard *et al.*[23] and is pinned to the deep-sea chronology at the 110-kyr level, denoted by the solid vertical line. Both GRIP and GISP2 records correlate, and are thought to be reliable back to interstadial 23, but the deeper occurrence of inclined layering at GRIP suggests that it may have a record that is reliable as far back as ~129 kyr ago[11]. As discussed in the text, $\delta^{13}C$ variability in GPC9 probably reflects changes in the relative proportion of North Atlantic Deep Water and Antarctic Bottom Water, and $CaCO_3$ variability likewise reflects changes in thermohaline circulation. Changes in thermohaline flow and its associated influence on meridional heat flux of the surface North Atlantic are probably linked to atmospheric temperature over Greenland for events indicated by short dashed arrows.

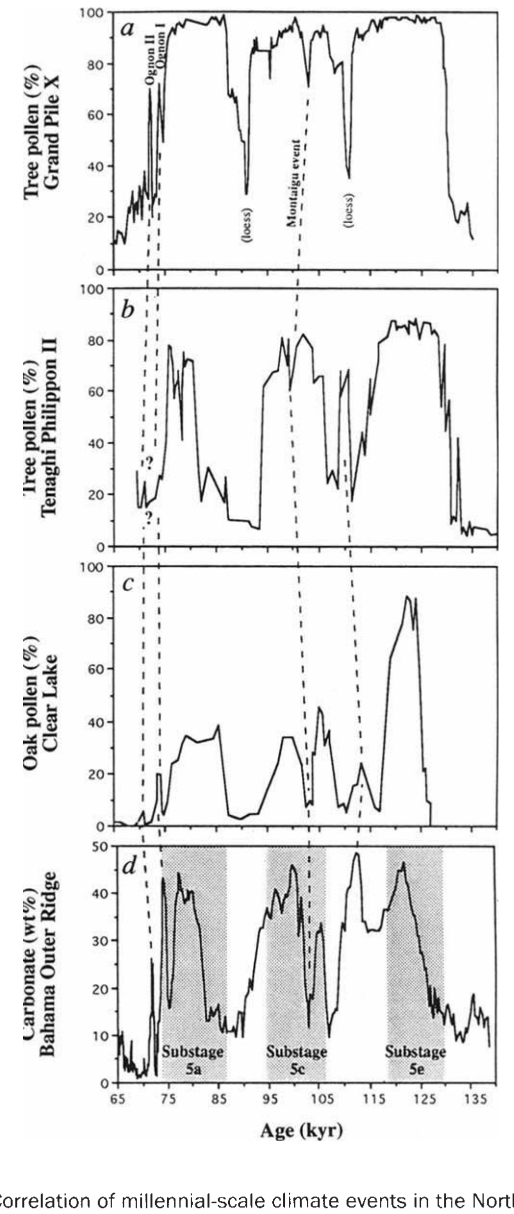

FIG. 3 Correlation of millennial-scale climate events in the North Atlantic, using wt% $CaCO_3$ in GPC9 as a proxy for deep ocean circulation change (*d*), with pollen records from Grand Pile in France[24] (*a*), from Tenaghi Philippon in Macedonia[29] (*b*), and from California[26] (*c*). We have applied the SPECMAP age model[14] to the pollen data using correlations proposed by the original authors. Shaded intervals in the GPC9 panel mark warm substages of interglacial isotope stage 5. Dashed lines correlate the Ognon I and Ognon II warm events of ~70–75 kyr ago, the cold Montaigu event which occurred ~103 kyr ago, and a warm event within stage 5d (~113 kyr ago). We propose that these and other short duration events in palaeoclimate proxy data are global, and may be related to brief changes in North Atlantic thermohaline circulation.

Carbon isotope changes in benthic foraminifera closely parallel the variability in the concentration of $CaCO_3$ (wt%) in GPC9 (Figs 1 and 2). Carbonate analyses provide a smoother, more detailed series of data than $\delta^{13}C$ because (not being subject to the availability of foraminifera) they can be performed on small bulk samples. The $\delta^{13}C$–$CaCO_3$ relationship has been observed previously for a few millennial-scale oscillations in the last 80 kyr of GPC9 as well as core KNR31-GPC5 from the northeast Bermuda Rise[9]. The large changes in sediment carbonate reflect mostly terrigenous input to the North Atlantic during glaciation, its redistribution by deep recirculating gyres (that is, the strength of thermohaline circulation), and $CaCO_3$ dissolution. As NADW (^{13}C-enriched) is less corrosive to $CaCO_3$ than AABW (^{13}C-depleted), replacement of NADW by AABW drives down percentage $CaCO_3$. Thus, both the records of wt% $CaCO_3$ and $\delta^{13}C$ in deep western North Atlantic cores respond to high-latitude thermohaline processes, accounting for their similarity.

Considering the importance of NADW production in maintaining the meridional flux of heat at the surface of the North Atlantic[1], it is appropriate to ask how our proxy data for deep ocean circulation change compare to the $\delta^{18}O$ data (an atmospheric temperature proxy) from the new Greenland ice cores. We focus on the GRIP core (Fig. 2*d*) because it appears to have a longer undisturbed record than the GISP2 core[11]. Age assignments for GRIP data are based on ice flow modelling, and correlation to marine isotope substage 5d[23] (Fig. 2*d*).

Younger than ~110 kyr, GRIP $\delta^{18}O$ results compare well with our proxy for NADW production within accepted dating uncertainties (a few thousand years) in each proxy record (Fig. 2). As noted previously for both core GPC9 and Bermuda Rise core GPC5, two maxima in $\delta^{13}C$ and wt% $CaCO_3$ near the stage 5a/4 transition are equivalent to isotopically warm events in Greenland and Antarctic ice cores, as well as European pollen records[9]. These events are clearly recorded in the GRIP core as interstadials 19 and 20. Interstadial 21 is equivalent to the peak of isotope substage 5a, although interstadial 22 is not readily identified at core GPC9. (That event may not be recorded because of a 10-kyr interval of especially low benthic foraminiferal abundance that prevented isotope analysis of all samples.) Marine isotope substage 5c is marked by a prominent double maximum in the NADW proxy data, and those data appear to correlate well with interstadials 23 and 24 in the ice core. Correlations older than substage 5d could be complicated by uncertainties in the ice-core record, but the presence of maxima in wt% $CaCO_3$ and $\delta^{13}C$ within substage 5d (~113 kyr ago) in GPC9 suggest that a strong warming may have followed peak interglacial conditions. How that event and the maximum in $\delta^{13}C$ that occurred during substage 5e at GPC9 correlate with the three oxygen isotopic maxima in the GRIP record remains inconclusive. We speculate that the warm event seen in the ice cores at ~115 kyr ago (Fig. 2*d*) may correlate with the $\delta^{13}C$ and wt% $CaCO_3$ maxima in substage 5d at GPC9. There is no evidence for deep ocean variability within substage 5e, implying that ice-core variability at that time must result from some other process.

An additional test of the link between NADW production, North Atlantic heat flux and air temperature over Greenland is provided by comparison to other climate indicators. In particular, it is useful to compare our marine record to pollen records that display previously unexplained variability at frequencies much greater than orbital ones. The events punctuating the stage 4 glacial inception in western North Atlantic cores and Greenland ice cores appear to correlate with the warm events labelled as Ognon I and Ognon II at the Grand Pile peat bog in France[24] (Fig. 3). Recently it was reported from detailed study of a second Grand Pile core that only one Ognon interstadial is clearly recognizable, and that the other event is probably due to pollen reworking[25]. A similar pair of events is also found in the Vostok ice core in Antarctica[9], and now we extend that correlation to Clear Lake, California (Fig. 3*c*), which to our knowledge contains the only continuous high-resolution pollen record reaching to the peak of the last interglaciation in North America[26]. Considering that the double events leading into stage 4 are evident in several Greenland sites, in Antarctica, in North America as well as in two high-resolution sediment cores from the North Atlantic, we conclude that these events are probably global in scope and somehow related to changes in formation rate of NADW.

Other high-frequency events from within the standard isotope stage stratigraphy are also found consistently among the most complete and best resolved palaeoclimate records. The isotopically cold event between interstadials 23 and 24 in the GRIP core, which corresponds to the $\delta^{13}C$ (Fig. 2*b*) and $CaCO_3$ minima at GPC9 at ~103 kyr ago (Figs 2*c* and 3*d*), is known from Grand Pile as the 'Montaigu event'[24,27]. That event is coherent among several French pollen sequences[27,28], and is present as well at Tenaghi Philippon in Macedonia[29] and at Clear Lake[26]. It is also possible to correlate the Montaigu event with a cooling ~103 kyr ago in Vostok deuterium data, using the new SPECMAP age model for that core[30]. Finally, we note that both Tenaghi Philippon and Clear Lake pollen reveal a warming that may correlate with the $CaCO_3$ and $\delta^{13}C$ maxima within substage 5d at GPC9. A similar warming is not evident in Grand Pile pollen sequences, perhaps because the substage 5d correlative (Melisey I) is represented by only a short section of core which is dominated by loess sedimentation[31]. We concede that these correlations are subjective and based on curve-matching, but previously there was no explanation for many of these high-frequency climate shifts. Here we provide the first evidence that these events may have been caused by changes in deep ocean circulation.

These deep ocean changes also recur more frequently than the intervals of accelerated iceberg discharge known as Heinrich Events[7], suggesting that processes other than (or in addition to) mechanical destabilization of large ice sheets modulated NADW flow. Although we cannot yet determine exactly what factors governed high-frequency variations in NADW production during stage 5, the most likely candidates are changes in meltwater flux and/or the balance between evaporation and precipitation over the high-latitude North Atlantic. The fact that we do not detect changes in NADW during substage 5e (Eemian), a time when there was less continental ice than now, suggests that ice sheets larger than today may be a critical ingredient for abrupt variations in NADW and climate. The concept of deep circulation control of North Atlantic heat flux provides a mechanism for explaining climate oscillations such as the Ognon I and II and Montaigu pollen events within the North Atlantic region, but the appearance of possibly correlative changes at more distant locations such as Antarctica and western North America probably requires additional feedbacks such as changes in water vapour content of the atmosphere[32]. □

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High-resolution climate records from the North Atlantic during the last interglacial

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THE two deep ice cores recovered by the GRIP[1] and GISP2[2] projects at Summit, Greenland, agree in detail over the past 100,000 years[3] and demonstrate dramatic climate variability in the North Atlantic region during the last glacial, before the current period of Holocene stability. This glacial climate instability has subsequently been documented in the marine sedimentary record of surface-ocean conditions in the North Atlantic[4]. Before 100 kyr ago the two ice core records are discrepant, however, casting doubt on whether the oxygen isotope fluctuations during the last interglacial (Eemian) seen in the GRIP core[1,5] represent a true climate signal. Here we present high-resolution records of foraminiferal assemblages and ice-rafted detritus from two North Atlantic cores for the interval 65 kyr to 135 kyr ago, extending the surface-ocean record back to the Eemian. The correlation between our records and the Greenland ice-core records is good throughout the period in which the two ice cores agree, suggesting a regionally coherent climate response. During the Eemian, our marine records show a more stable climate than that implied by the GRIP ice core, suggesting that localized phenomena may be responsible for the variability in the latter record during the Eemian.

We have examined two cores (DSDP site 609, V29-191) from the eastern North Atlantic (Fig. 1) at a depth spacing corresponding to approximately 200 years throughout marine isotope stage 5 (MIS 5). This stage was broadly defined as an interglacial period in the original oxygen isotope stratigraphy[6] and subsequently partitioned into substages 5a to 5e, with 5e alone correlative with the last interglacial (Eemian) in terrestrial records[7]. The two cores bear evidence of repeated encroachment of polar surface waters. While advances across a 20° swath of latitude have previously been documented on orbital timescales[8,9], water-mass migrations evident in this study are correlative with the higher frequency changes occurring on the Greenland ice sheet, beginning ~110,000 years ago[1,2]. Since at least this time, cooling on the ice cap and at the sea surface have been linked.

Two indicators of oceanographic conditions were utilized for this study: the planktonic foraminiferal assemblage and the abundance of ice-rafted detritus (IRD), measured relative to coarse biogenic particles and to total sediment. Time series of these proxies are remarkably similar (Fig. 2). This similarity of

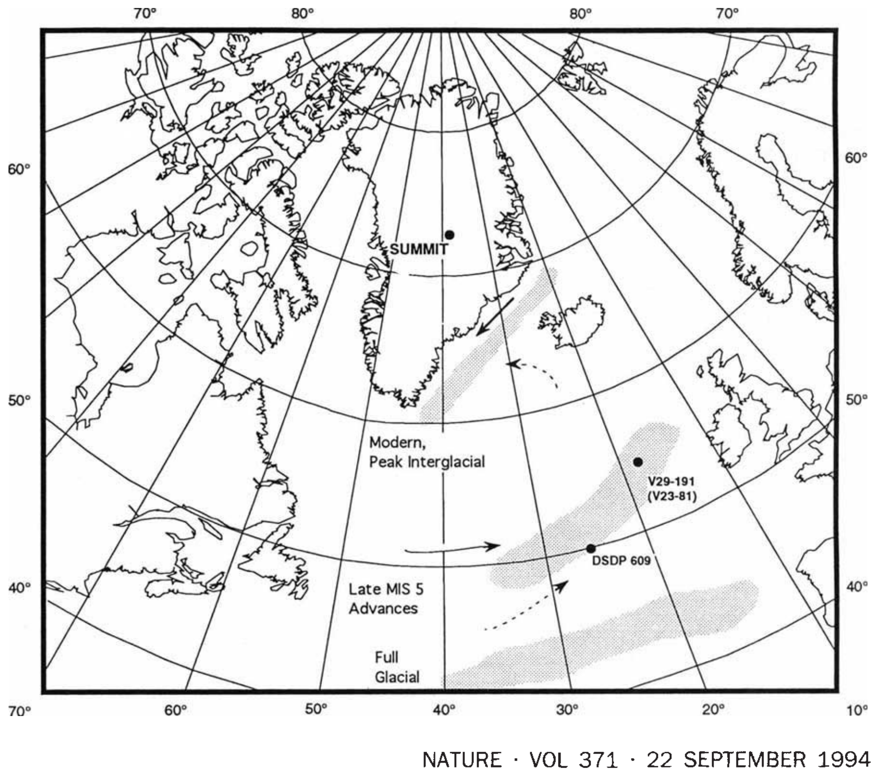

FIG. 1 Study area in North Atlantic. The Greenland location is the ice-core drilling site at the summit of the ice sheet. Ocean locations are sites of sediment cores used in this study. DSDP site 609 was drilled in a local basin on the eastern flank of the Mid-Atlantic Ridge. V29-191 was cored on an abyssal drift deposit along Feni ridge. V23-81 is a nearby core used for a previous comparison with the glacial portion of the Greenland record. Shaded areas represent the approximate location of polar water boundaries. Solid arrows indicate cold, fresh Arctic/sub-Arctic flows. Dashed arrows indicate warm, saline subtropical/subpolar flows.