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North Atlantic thermohaline circulation during the past 20,000 years linked to high-latitude surface temperature

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During a surface cooling event 10,000 to 12,000 years ago, higher Cd/Ca and lower $^{13}C/^{12}C$ ratios are observed in benthic foraminifera shells from rapidly accumulating western North Atlantic sediments. Data from sediment cores show that marked nutrient depletion of intermediate waters occurs in association with reduced glacial North Atlantic Deep Water flux. It is proposed that cold high-latitude sea surface temperatures enhance intermediate-water formation at the expense of deep-water formation.

THE transition from latest Pleistocene glaciation to the Holocene interglacial period occurred over thousands of years[1-4]. In the North Atlantic Ocean and continental Europe, this long-term warming trend was interrupted by a brief return to glacial temperatures ~10,000-12,000 yr ago referred to as the 'Younger Dryas'[5,6]. The relative abundance of cold- and warm-water fossil fauna in oceanic sediments shows that the surface North Atlantic was also cooler then. Recent accelerator radiocarbon dating of single-species foraminifera from a high-accumulation-rate sediment core indicates a terrestrial-equivalent radiocarbon age of ~10,300 yr[7]. Oxygen isotope studies from ice cores demonstrate that deglacial warming in Greenland was also interrupted by a brief period of near-glacial temperatures[8]. Recent counting of annual $\delta^{18}O$ and dust cycles indicates that this event ended 8,770 chronological years ago[9] (equivalent to 9,050 conventional radiocarbon years). These events are synchronous within the uncertainties in the ages; a climate model study indicates that synchroneity is to be expected[10]. The sudden reversal of the warming trend is not seen in ice-volume indicators, except perhaps as a pause in melting[2].

This event is one of the shortest well-documented extreme climate reversals, occurring over $\leq$1,000-2,000 yr. The rapidity of this fluctuation can help determine the response time of other climate subsystems. For example, the Younger Dryas can be used to test the response of deep-ocean circulation to a short climate event. The duration of the Younger Dryas climate reversal is short compared to Milankovitch (orbital) cycles, implying that other mechanisms can generate large climatic changes. A number of models have been proposed to account for the Younger Dryas, but no consensus has been reached yet[6].

No convincing evidence for a deep-ocean response to Younger Dryas cooling has been obtained from high-sedimentation-rate cores. Given normal deposition at 2-3 cm kyr^{-1} (and typical biological stirring of the upper sediment layer down to 8 cm), it is unlikely that such a short-duration event could be reliably recorded in most deep-sea sediment cores. The north-east Bermuda Rise, a feature created from transport of fine-grained detritus, has high deposition rates, of >10 cm per thousand yr[11,12]. We present data from a Bermuda Rise core which document the deep-ocean response to Younger Dryas cooling.

The sediment core and its timescale

Core EN120 GGC1 is a large-diameter (10 cm inner diameter) gravity core. It was raised from 4,450 m depth (33°40′ N, 57°37′ W) on RV *Endeavor* cruise 120, 31 August 1984. The core site is near that of core KNR31 GPC5 for which previous stratigraphic work has been undertaken[11,12].

The timescale for this core is estimated by cross-correlation with features that have been radiocarbon dated in other cores from the North Atlantic basin. This correlation assumes that ice volume is the dominant factor controlling changes in benthic foraminiferal $\delta^{18}O$, with a lesser contribution from deep-ocean temperature[13]. The ice-volume signal must be globally synchronous within the mixing time of the ocean (<1,000 yr[14]); part of the temperature contribution to $\delta^{18}O$ variability is likely to be correlated with ice volume as well (W. F. Ruddiman, in preparation). The estimated timescale is based on the following assumptions. (1) The sample with most positive $\delta^{18}O$ is assigned the conventional glacial maximum age of 18,000 yr. Because benthic foraminifera are scarce in the glacial section of the core, there is some uncertainty as to the true depth of the glacial maximum (Fig. 1). The following discussion will not rely on the details of the chronology preceding deglaciation. (2) The first rapid shift of ^{18}O towards interglacial values ('Termination Ia') is very clear in this core at 145 cm, with an uncertainty of no more than a few cm. This feature has been dated at 14,000 radiocarbon years using the accelerator radiocarbon technique in another North Atlantic sediment core[15]. (3) After a brief $\delta^{18}O$ plateau, a second rapid shift to full interglacial conditions (Termination Ib) occurs which has been accelerator radiocarbon dated at 8,500 radiocarbon years[15]. This transition is less clearly defined than Termination Ia, but it occurs somewhere between 80 and 95 cm in EN120 GGC1. The Termination Ib level was chosen at 90 cm. (4) There is a regional $CaCO_3$ minimum (at 17 cm depth in this core), as well as several other $CaCO_3$ cycles in the Holocene section of this core (Fig. 1), that are identical to cycles observed in nearby core GPC5 (L. D. Keigwin and G. A. Jones, in preparation). Planktonic foraminiferal carbonate from the youngest carbonate minimum has been radiocarbon dated at 1,900 yr in KNR31 GPC5[11]. (5) The core top is assigned an age of 0 years. In view of the characteristics of the corer, similarity of the chemical data for the core top and Smith-McIntyre box-core sample, and the high sedimentation rate, this assignment appears reasonable and is unlikely to be low by more than a few hundred years.

With these control points, other depth intervals are assigned ages by linear interpolation (and extrapolation beyond the glacial maximum). Preliminary results from accelerator radiocarbon dates in this core and nearby core GPC5 support the timescale used here. In particular, in GPC5, the Younger Dryas interval is bracketed by radiocarbon dates of 9,210±150 yr (*G. ruber*) at 175 cm and 11,150±310 years (*N. pachyderma*) at 202 cm (L. D. Keigwin and G. Jones, in preparation). The $CaCO_3$ cross-correlation indicates that these intervals in GPC5 are equivalent to 100 cm (interpolated age 9,500 yr) and 115 cm (interpolated age 11,000 yr) in EN120 GGC1. It is unlikely that

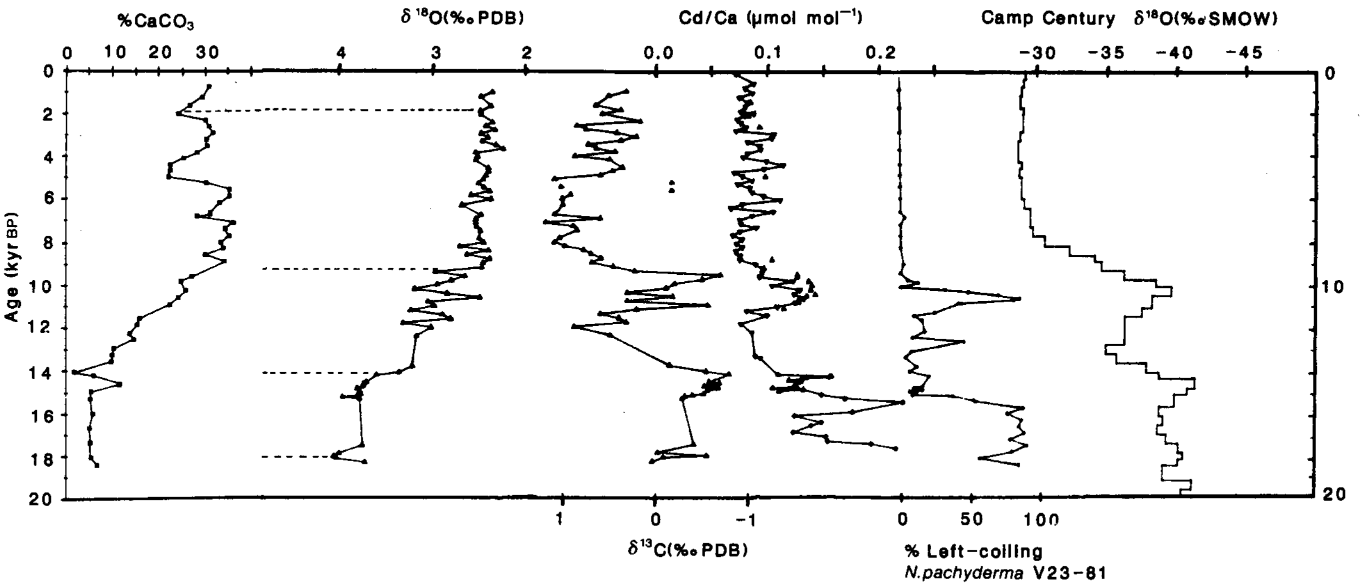

Fig. 1 Data from core EN1200 GGC1 against estimated age. The line connecting $\delta^{13}C$ points is interrupted where two outlier points occur. In the Holocene section of the core, *N. umbonifera* (inverted triangles) was used for Cd/Ca analyses; in the Younger Dryas section both *C. wuellerstorfi* (triangles) and *N. umbonifera* were used; during the deglacial transition, only *C. wuellerstorfi* was used, and in the glacial period *Uvigerina* (diamonds) was used. Although the use of multiple species might raise concerns about possible differential species fractionation, the analysis of several hundred paired samples of these species shows no evidence for differential fractionation, including intervals of this core where species overlap. In particular, both *C. wuellerstorfi* and *N. umbonifera* were analysed in the Younger Dryas section, and both species confirm the event. The line is drawn through the *N. umbonifera* data, with the *C. wuellerstorfi* data plotted as individual points. Note that the two right-hand datasets are cross-correlated from other records, not measured in our core. $\delta^{13}C$ ($\equiv 1{,}000\times ((^{13}C/^{12}C)_{sample}-(^{13}C/^{12}C)_{standard})/(^{13}C/^{12}C)_{standard}$) is measured in parts per thousand relative to the PDB standard; $\delta^{18}O$ is measured relative to standard mean ocean water (SMOW).

any of the Holocene age assignments are >500 yr in error. Each 2-cm sample represents the equivalent of 200 years of deposition. If the biologically stirred mixed layer on the sea floor is 8 cm deep (and a gravity-core foreshortening factor of 0.57 is assumed), then the resolution of the record will be better than 400 yr.

Deep-water variability

Benthic oxygen isotope data document the core's continuity of sedimentation and high resolution (Fig. 1). Benthic carbon isotope and cadmium data from this North Atlantic sediment core provide information on deep-water nutrient variability, which is linked to changes in relative proportions of North Atlantic and Antarctic deep-water sources[16-18]. Nutrient differences between North Atlantic and Antarctic waters result from the interaction between biological nutrient cycling and the general circulation of the ocean[14]. As various sources of deep water spread throughout the ocean, their characteristic chemical signatures remain evident on a global scale[19-21]. Because North Atlantic Deep Water derives from nutrient-depleted waters of subtropical origin (whereas Antarctic waters originate from nutrient-rich waters upwelled from great depth), North Atlantic Deep Water is enriched in ^{13}C and depleted in Cd and nutrients relative to Antarctic waters.

In addition to the effect of chemical redistribution throughout the ocean, chemical inventories change in response to imbalances between input (continental weathering) and output (sedimentation). Changes in the global oceanic inventory will influence the record at each site in the ocean[22]. Data from ocean sediment cores suggest that glacial mean $\delta^{13}C$ was $\sim -0.5‰$ relative to the modern ocean, and the oceanic Cd inventory does not appear to change significantly[22,23]. Carbon isotope variability in this core encompasses equal contributions from changes in the oceanic mean and circulation, while Cd responds mainly to changes in ocean circulation.

As expected, benthic data from this core show that glacial deep water (4,450 m) is depleted in ^{13}C and enriched in Cd relative to the present. This reflects a lower proportion of northern source water relative to admixed southern source water, as seen in previous studies on cores from 2,500-3,500 m[16,17,22,24]. The amplitude of the signal in this record is larger than that observed for shallower, lower-sedimentation-rate cores. Both $\delta^{13}C$ and Cd data imply that the most nutrient-depleted waters (hence greatest relative North Atlantic Deep Water flow) occurred in the early Holocene (7,500-9,000 yr ago). There is some hint of upper Holocene fluctuations but these are not equally supported by both $\delta^{13}C$ and Cd, and hence are not considered here.

Both $\delta^{13}C$ and Cd data show that an episode of nutrient-enriched deep water occurred between 9,000 and 11,500 years ago. Coincident with (or perhaps preceding) the initial phase of deglaciation 14,000 years ago (Termination Ia), North Atlantic Deep Water proportions rose from a glacial nadir to levels comparable to those found in the modern ocean. These high fluxes continued for ~2,000 years. Between 9,000 and 11,500 years ago, North Atlantic Deep Water ebbed to a level only slightly higher than that of the glacial maximum. During this event, reduced North Atlantic Deep Water persisted until just before the renewed warming phase (Termination Ib). This ebb of the North Atlantic Deep Water coincided (within the likely error of the timescale) with the Younger Dryas cooling event observed on the European continent, in Greenland, and in high-latitude surface waters of the North Atlantic Ocean. These data demonstrate, for the first time, that this brief cold event is associated with a response of similar import and duration in the deep ocean. It is now evident that the deep ocean can undergo dramatic changes in its circulation regime on timescales of ≤500 yr.

North Atlantic Intermediate Water

Higher nutrient concentrations in the deep North Atlantic (2,500-4,500 m) are accompanied by lower nutrient levels at intermediate depths. Down-core analyses of $\delta^{18}O$, $\delta^{13}C$ and Cd in benthic foraminifera from western North Atlantic core CHN82 Sta41 Core15PC (43°22.3′N, 28°13.9′W, 2,151 m; Table 2, Fig. 2) do not indicate any increase in nutrient content during

Table 1 Data from sediment core EN120 GGC1

Depth (cm)	Cd species	Cd/Ca (μmol mol^{-1})	$\delta^{18}O$ (‰ PDB)	$\delta^{13}C$ (‰ PDB)
0	umb	0.072		
	wue	0.073		
5	umb	0.088		
7	umb	0.080		
9	umb	0.087	2.36	0.33
11	umb	0.075	2.48	0.52
13	umb	0.082		
	wue	0.085		
15	umb	0.080	2.37	0.66
17	umb	0.076	2.49	0.39
18	umb	0.088		
19	umb	0.080		
	wue	0.083	2.48	0.59
21	umb	0.073		
23	umb	0.076	2.24, 2.48	0.32, 0.03
25	umb	0.081	2.43	0.86
	wue	0.093		
27	umb	0.072		
	wue	0.127	2.33	0.76
29	umb	0.106	2.57, 2.39	0.49, 0.39
31	umb	0.104	2.41, 2.40	0.53, −0.08
33	umb	0.082	2.47	0.39
35	umb	0.093	2.32	0.74
37	umb	0.094	2.24	0.65
39	umb	0.082	2.54	0.45
41	umb	0.078	2.52	0.88
43	umb	0.099	2.54	0.51
45	umb	0.114		
47	umb	0.097	2.40	0.37
49	umb	0.070	2.40, 2.40	0.77, 0.19
51	umb	0.077		
	wue	0.098	2.43	0.61
53	umb	0.086	2.46	1.10
55	umb	0.073	2.52, 2.49	−0.34, 0.01
57	umb	0.084	2.45	1.03
59	umb	0.086	2.43, 2.35	−0.18, −0.14
61	umb	0.096	2.59	0.93
63	umb	0.111	2.37	1.02
65	umb	0.077		
66			2.69	1.01
67	umb	0.086		
69	umb	0.105		
71	umb	0.086	2.48	1.10
73	umb	0.075	2.54	0.61
75	umb	0.076	2.54	1.20
77	umb	0.090	2.53	0.90
79	umb	0.075		
	wue	0.106	2.49	0.86
81	umb	0.069		
83	umb	0.077	2.50	1.05
85	umb	0.072	2.45	1.10
87	umb	0.077	2.71	0.99
89	umb	0.071	2.40	0.79
91	umb	0.075	2.56, 2.69	0.38, 1.04
93	umb	0.075	2.39	0.60
95	umb	0.089	2.46	0.70
97	umb	0.095		
	wue	0.097	2.47	0.47
99	umb	0.095		
	wue	0.127	2.97	0.24
101	umb	0.093		
	wue	0.127	2.65	−0.68
103	umb	0.123		
	wue	0.137	2.80	−0.49
105	umb	0.104		
	wue	0.140	2.95	−0.19
107	umb	0.129		
	wue	0.139	3.19	−0.10
109	umb	0.124		
	wue	0.143	2.84	0.32
110	wue	0.135		
111	umb	0.133		
	wue	0.129	2.49	−0.17
113	umb	0.125		
	wue	0.128	3.05	0.32
115	umb	0.109	2.98	−0.55
116	wue	0.115		
117	umb	0.081	3.23	0.22
119	umb	0.100	2.89	0.61
121			2.80	0.41
123	umb	0.076	3.31	0.33
125			3.01	0.89
127	umb	0.086		
	wue	0.111		
129			3.17	0.50
138	umb	0.089		
139	umb	0.094		
143			3.21	−0.14
147			3.35	−0.54
149	wue	0.110		
151	wue	0.158	3.59	−0.78
152	wue	0.156		
153	wue	0.135		
155	wue	0.131		
157	wue	0.120		
159	wue	0.130		
161			3.70	−0.57
163	wue	0.125	3.73	−0.68
165	wue	0.127		
167	wue	0.105	3.74	−0.52
169	wue	0.133	3.80	−0.67
171	wue	0.111		
173			3.77	−0.55
176	Uvi	0.149		
177			3.76	−0.51
179			3.79	−0.39
181	Uvi	0.170	3.96	−0.31
183			3.77	−0.28
186	Uvi	0.222		
199	Uvi	0.177		
204	Uvi	0.125		
213	Uvi	0.149		
217	Uvi	0.140		
226	Uvi	0.124		
232	Uvi	0.153		
238	Uvi	0.155		
242	Uvi	0.194		
245			3.74	−0.41
248	Uvi	0.216		
256			3.99	−0.02
260			4.05	−0.55
264			4.02	−0.08
273			3.71	0.04

All isotope data is from *C. wuellerstorfi.* '0'-cm sample is from Smith-McIntyre surface sediment sampler; all other depths refer to distance along core liner (sediment surface in core is at 4 cm). In the species column the symbols are: wue, *C. wuellerstorfi*; umb, *N. umbonifera*; Uvi, *Uvigerina.*

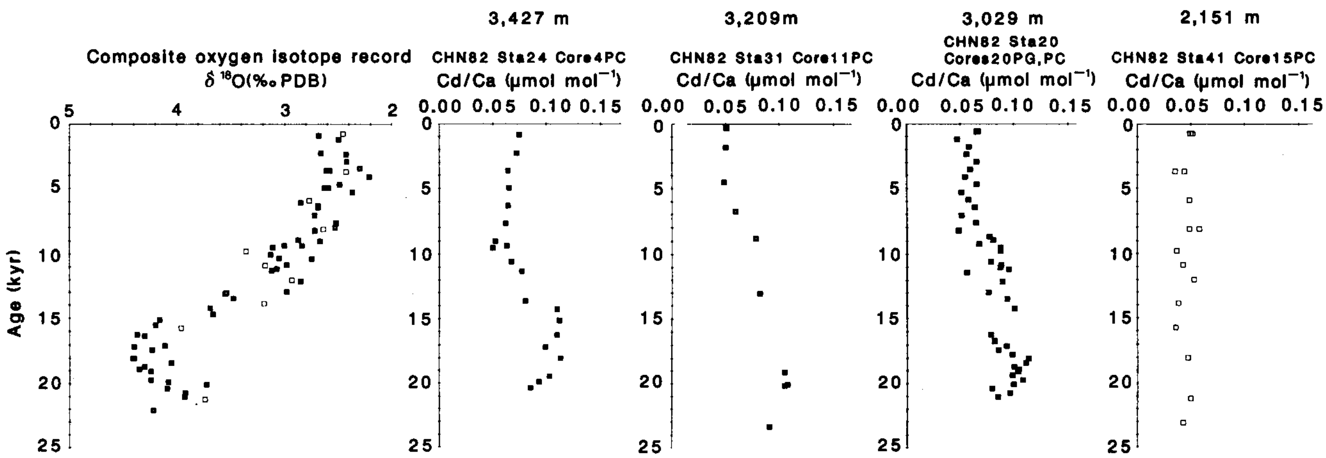

Fig. 2 Oxygen isotope and Cd data from benthic foraminifera from cores comprising a depth transect between 2,100 and 3,500 m at 42° N in the western North Atlantic ocean. The timescale (consistent with that used for EN120 GGC1) is chosen to produce the best overlap of the oxygen isotope records. The $\delta^{18}O$ plot indicates the relative success of the overlap. As the cores used here have lower sedimentation rates (varying from 1.5 to 5 cm kyr^{-1}), the resolution is poorer than for EN120 and should be considered reliable to no better than a few thousand years.

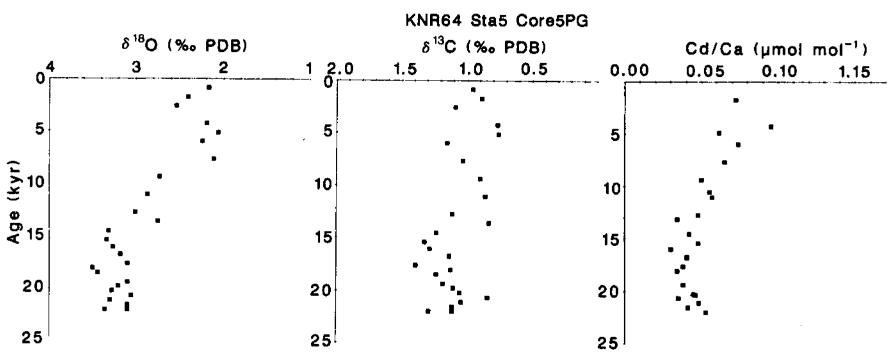

Fig. 3 Isotope and Cd data from benthic foraminifera from Carribean Sea core KNR64 Sta5 Core5PG (16°31.3′ N, 74°48.4′ W, 3047 m) plotted against estimated age. The age scale is derived from assuming that the physical top of the core (6 cm relative to the liner) is 0 kyr, that 16 cm is 8.5 kyr, that 22.5 cm is 14 kyr and that 41 cm is 18 kyr. The sedimentation rate of the core limits true resolution to ~5,000 years in the Holocene, somewhat better in the glacial. A large fraction of the total glacial/interglacial amplitude for $\delta^{18}O$ is observed, so that at least 70% of the Cd amplitude must be recorded.

glacial times. Furthermore, data from a Caribbean Sea core (KNR64 Sta5 Core5PC, 16°31.3′ N, 74°48.4′ W, 3,047 m) demonstrate that Caribbean nutrient concentrations decreased during the last glacial maximum (Table 2, Fig. 3). Reduced glacial Caribbean nutrients must be due to lower nutrient concentrations in source waters for the basin. The Caribbean derives its waters from ~1,700–1,800 m depth in the North Atlantic (the sill depth is 1,800 m, but some shallower water mixes in during inflow)[25]. Ventilation of this basin is sufficiently rapid at present to minimize nutrient enhancements due to decomposing particles from the ocean surface. Modern source waters have phosphorus contents of ~1.5 μmol kg^{-1} (ref. 19); phosphorus contents in the Caribbean are comparable, never being more than 0.2–0.4 μmol kg^{-1} higher[26]. Because the modern local nutrient input is small, significant reductions in Caribbean nutrient levels can be generated only by decreased nutrient concentrations in the advective sources. The decrease in glacial Caribbean nutrients (inferred from decreased Cd and higher $\delta^{13}C$) shows that Atlantic intermediate-depth waters were more nutrient-depleted at a time when deeper waters were more nutrient-enriched. These results can be expressed as a reconstructed profile of phosphorus in the glacial ocean (Fig. 4).

CLIMAP[27] included data showing more positive glacial Caribbean $\delta^{13}C$ in isotope stage 6 relative to stage 5e (although they did not discuss this). Cofer-Shabica and Peterson (in preparation) obtained $\delta^{13}C$ data from a long Caribbean core showing that glacial nutrient depletion occurs repeatedly during the late Quaternary (although they suggested that reductions in nutrient fluxes from the surface were responsible, which is not feasible). Higher intermediate-depth Atlantic $\delta^{13}C$ values were also reported for the glacial period preceding the last interglacial period ~140,000 yr ago[28] (although some caution should be applied in considering that data, because some of the foraminifera used are now known to be unreliable[29]). Subsequent to the present measurements and these other studies, Oppo and Fairbanks[18] made measurements on two Caribbean cores reinforcing these results. Hence it is well established that glacial Caribbean and intermediate-depth Atlantic waters were nutrient-depleted. The Cd data indicate that the nutrient content decreased by ~~a~~ factor of two, equivalent to a phosphorus decrease from 1.6 to 0.8 μmol kg^{-1}.

It has not yet been established that the intermediate-depth ocean responded in the same way during the Younger Dryas cold event as it did during the glacial maximum. This may be due to the lack of a suitable high-sedimentation-rate core which could document this event. If there is a mechanistic link between changes in the deep and intermediate ocean, then we would expect to find such an event when suitable cores are recovered.

Possible mechanistic link

The model we propose derives from analogy with the modern North Pacific Ocean, where active intermediate-water production occurs, but no deep water is formed. Deep North Pacific waters can be traced only to southern deep-water sources[30]. The immediate reason for this difference between the oceans is that North Pacific surface waters are relatively fresh (32.9‰). Even when cooled to their freezing point, these waters cannot attain densities as high as those of Antarctic and North Atlantic deep and bottom waters. On winter cooling, these waters become sufficiently dense only to sink to intermediate levels. For this reason, a tongue of high-oxygen, low-salinity water is observed to spread from this source[31]. The low salinity of North Pacific surface waters is maintained by the balance between advection, evaporation, precipitation and runoff. Warren[30] has shown that precipitation is similar in the North Atlantic and North Pacific, but evaporation is lower in the North Pacific due to lower surface

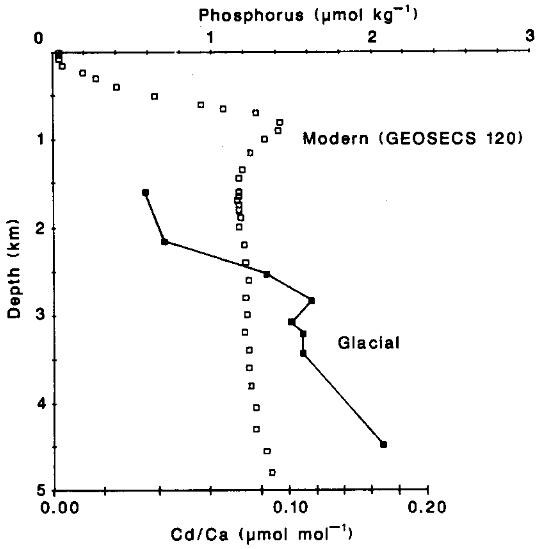

Fig. 4 Glacial depth profile of phosphorus in the modern western North Atlantic (GEOSECS station 120) compared to glacial profile. The glacial profile is based on the cores shown in Figs 1-3 as supplemented by two others whose glacial section is determined from percentage CaCO3 curves (CHN82 Sta30-1 Core9PC, 41°50.6′ N, 26°27′ W, 2,830 m, glacial sample at 30-40 cm; CHN82 Sta23 Core3PC, 41°38′ N, 27°20′ W, 2,525 m; glacial sample at 36-42 cm). The actual data for the glacial profile are Cd/Ca ratios in benthic foraminifera (bottom scale). The phosphorus scale is estimated assuming that the Cd-P relationship in the glacial ocean is similar to that of today, and assuming that the distribution coefficient for Cd/Ca in foraminifera relative to sea water is 2.9. Note the change in the Cd scale, which is based on observation of a significant 'kink' in the Cd-P relationship (ref. 23). Actual glacial P is uncertain to the extent that this relationship may have been slightly different during glacial times and to the extent that the distribution coefficient is uncertain (it is known to no better than 15%). Despite these uncertainties, the important feature is the relative vertical homogeneity in the modern North Atlantic compared to the marked depth gradient in the glacial ocean. This qualitative difference depends neither on the assumed Cd-P relationship nor on the assumed distribution coefficient.

Table 2 Data from two cores

Depth cm	Cd/Ca (μmol mol^{-1})	$\delta^{18}O$ (‰ PDB)	$\delta^{13}C$ (‰ PDB)
CHN82 Sta 41 Core15PC			
2	0.051	2.46	0.63
6	0.040	2.43	1.28
9	0.049	2.77	1.24
12	0.054	2.64	1.31
16	0.037	3.36	1.21
19	0.043	3.18	1.27
22	0.053	2.93	0.87
25		3.55	0.88
27	0.038	3.19	1.02
29	0.036	3.96	1.17
31	0.047	3.92	1.19
33		3.95	1.27
36	0.050	3.74	1.21
39	0.043		
41	0.048	3.42	1.18
KNR65 Sta5 Core5PG			
7		2.17	0.96
8	0.073	2.41	0.89
9		2.54	1.09
11	0.095	2.19	0.77
12	0.061	2.05	0.76
13	0.074	2.24	1.15
15	0.065	2.11	1.03
17	0.050	2.73	0.90
18	0.055		
19	0.057	2.87	0.86
21	0.048	3.01	1.11
22	0.035	2.75	0.83
25	0.042	3.32	1.23
29	0.048	3.34	1.32
32	0.030	3.27	1.28
35	0.041	3.18	1.13
39	0.038	3.10	1.39
41	0.034	3.50	1.12
43		3.44	1.23
47	0.038	3.10	1.18
49		3.21	1.10
51		3.28	1.05
52	0.046		
53	0.035	3.06	0.84
55		3.30	1.04
55	0.048		
57	0.041	3.10	1.11
59	0.053	3.23	1.20

Cd data from *C. kullenbergi* and *C. wuellerstorfi*. Isotope data from *C. wuellerstorfi*. The time interval is verified as isotope stages 1 and 2 by the presence of excess ^{230}Th (ref. 38) and the absence of *G. flexuosa* (ref. 39).

temperatures. Essentially, water 'distills' off the warm North Atlantic and is transported by the atmosphere to 'condense' onto the cool North Pacific[6].

Proxy climate data show that the glacial high-latitude North Atlantic was much colder than at present[32,33]. An analysis similar to Warren's[30] for the modern and glacial North Atlantic from 40 to 65 °N and from 10 °E to 70 °W indicates the consequences of that cooling (Table 3). Neglecting the effect of changes in the wind speed and relative humidity over the glacial North Atlantic, cooler glacial temperatures would tend to reduce the rate of evaporation by a factor of two. This difference is comparable to the contrast between the modern North Atlantic and Pacific. This effect is reinforced by diminished northward transport of saline water from lower latitudes. Processes tending towards lower surface salinities are countered in part by increased wind strengths which enhance evaporation[34]. Furthermore, dry catabatic winds cascading off continental ice masses would have reduced atmospheric relative humidity near the ice edge and contributed to the formation of cold saline water (J. Reid, personal communication). Hence, although it is plausible on the basis of temperature changes that the salinity of the Atlantic north of 45 °N tended to lower values during glacial times (which has already been surmised qualitatively[32]), it probably did not attain the state of the modern North Pacific.

This decrease in salinity would have made it more difficult to form very dense waters in the North Atlantic. It would have favoured more production of intermediate waters. The palaeochemical evidence shows that the glacial North Atlantic did not fully attain the conditions of the modern North Pacific; some deep water must have been forming to maintain the Atlantic at lower nutrient levels than the Pacific[16,24]. The North Atlantic probably shifted towards a state more like the North Pacific, with less deep-water production and more intermediate-water production. This model predicts that glacial North Atlantic intermediate waters would have been colder and fresher (and perhaps relatively more depleted in ^{18}O).

An alternative model

Another intermediate-water source for the modern Atlantic is the outflow of nutrient-depleted saline water from the Mediterranean. Zahn, Sarnthein and Erlenkeuser[35] studied the carbon isotope composition of sediments from intermediate depths in the eastern Atlantic from Gibraltar southwards. They found evidence for enhanced nutrient depletion of these mid-depth waters during glacial times. This evidence was interpreted as indicating an increased flux of nutrient-depleted Mediterranean outflow water. Oppo and Fairbanks[18] followed up this hypothesis to argue that the glacial intermediate-water nutrient depletion discussed above was due to this increased flux of outflow water from the Mediterranean. This enhanced Mediterranean outflow water/intermediate water hypothesis predicts that glacial intermediate waters were warmer and more saline, in contrast to the predictions of the glacial North Atlantic Intermediate Water hypothesis presented above. Observations that could establish the salinity or other characteristic properties of these water masses would be of help in distinguishing between these hypotheses.

Current evidence is not sufficient to disprove the Mediterranean outflow water hypothesis, but an unusual conjunction of circumstances is required by this model. Water from the Mediterranean contributes a very small fraction (<10%) of mid-depth water in the modern Atlantic. All other sources of intermediate water would have to wane before Mediterranean outflow could become the dominant source and determinant of intermediate-water nutrients. In view of the favourable conditions for North Atlantic Intermediate Water formation discussed above, a substantial reduction in all other sources seems unlikely. Furthermore, the degree to which Mediterranean water outflow could increase is constrained by the relationship between evaporation, density contrast and hydraulic controls, which suggests that there should not be major changes in the rate of outflow (ref. 36; R. C. Thunnell, personal communication). It is even possible that the mid-depth nutrient depletion observed in the eastern Atlantic[35] is simply due to impingement of glacial North Atlantic Intermediate Water on the continental margin, and that no enhancement of Mediterranean outflow is necessary.

Conclusions and speculations

A link between the ocean surface and deep waters is now established for one of the most rapid extreme climate events. On timescales of several hundreds to several hundred thousand years, colder surface temperatures in the high-latitude North Atlantic are associated with reduced fluxes of nutrient-depleted Northern source water into the deep Atlantic. During the last glacial maximum, this association is accompanied by intermediate-water nutrient depletion. We propose that lower glacial North Atlantic surface temperatures reduce evaporation and hence decrease high-latitude surface salinities. The lower potential density of these surface waters when cooled leads to the formation of glacial North Atlantic Intermediate Water at the expense of reduced formation of glacial North Atlantic Deep Water.

This hypothesis might be taken to imply that the surface ocean controls a passive deep ocean. To the extent that intermediate- and deep-water events are coupled, this implication is intended. But it is not clear whether surface ocean temperatures control deep-water formation, or the reverse: warm surface temperatures in the modern North Atlantic may be due in part to the requirement for northward flow of warm surface water to compensate southward flow of cold deep water. The North Atlantic Current is wind driven and hence controlled by the atmospheric circulation. But on a global scale, atmospheric wind patterns are also affected in turn by the distribution of ocean surface temperatures. Held[37] suggests that orographic controls on atmospheric circulation (the relative positions of the Rockies and Himalayas) may account in part for differences in wind stress over the North Atlantic and North Pacific. If this is so, then the primary cause of warm North Atlantic surface temperatures is the topography of the continents coupled to atmospheric dynamics, with the ocean playing a relatively passive role. The presence of a large ice mass on the North American continent has a dramatic orographic effect on atmospheric circulation patterns. Modelling studies[34] suggest that these changes could account for the southward movement of the polar front during the last glacial maximum. The response of the planetary wave system to changes in position and elevation of the wasting glacial ice mass could possibly have led to the Younger Dryas cooling. Other plausible causes have been suggested[6,7], so this remains speculative. In any event, such considerations illustrate how the long-term progression of glaciation might be linked to both orbitally induced insolation redistribution and the dynamics of the fluid and solid Earth.

Table 3 Evaporation in the modern and glacial North Atlantic

	February	August	Annual average
Modern	26.6	12.7	19.7
Glacial	15.1	8.3	11.7

The calculations were carried out as described by Warren[6]. The block size was 5 degrees of latitude by 10 degrees of longitude. The results are expressed as the areal average of the product of wind speed times the specific humidity difference between air and saturation at the sea surface temperature (units: mm s^{-1}). To convert these values to evaporation rates in mm s^{-1}, it is necessary to multiply by the density of air (1.25×10^{-3} g cm^{-3}) and by the exchange coefficient for moisture, which has been estimated in the range $(1.2\text{-}2.1) \times 10^{-3}$ (ref. 6). The area of open water is slightly less during the glacial period, but it is not clear exactly how sea-ice cover should be treated for the evaporation-precipitation cycle.

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1. Broecker, W. S. in *Milankovitch and Climate* (eds Berger, A. *et al.*) 687-698 (Reidel, Dordrecht, 1984).
2. Duplessy, J.-C., Delibrias, G., Turon, G., Pujol, C. & Duprat, J. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* **35**, 121-144 (1981).
3. Mix, A. C. & Ruddiman, W. F. *Quat. Sci. Rev.* **4**, 59-108 (1985).
4. Bloom, A. L. in *Late Cenozoic Glacial Ages* (ed. Turekian, K. K.) 355-380 (Yale University Press, 1971).
5. Lowe, J. J., Gray, J. M. & Robinson, J. E. *Late Glacial of Northwest Europe* (Pergamon, Oxford, 1980).
6. Broecker, W. S., Peteet, D. M. & Rind, D. *Nature* **315**, 21-26 (1985).
7. Broecker, W. S. *et al. Palaeoceanography* (in the press).
8. Dansgaard, W. *Am. geophys. Un. Monogr.* **29**, 288-298 (1984).
9. Hammer, C. U., Clausen, H. B. & Tauber, H. *Radiocarbon* **28**, 284-291 (1986).
10. Rind, D., Peteet, D., Broecker, W., McIntyre, A. & Ruddiman, W. *Clim. Dyn.* **1**, 3-33 (1986).
11. Keigwin, L. D., Corliss, B. H., Druffel, E. R. M. & Laine, E. P. *Quat. Res.* **22**, 383-386 (1984).
12. Laine, E. P. & Hollister, C. D. *Mar. Geol.* **39**, 277-300 (1981).
13. Shackleton, N. J. & Chappell, J. *Nature* **324**, 137-140 (1986).
14. Broecker, W. S. & Peng, T. H. *Tracers in the Sea* (Eldigio, Palisades, 1982).
15. Duplessy, J.-C. *et al. Nature* **320**, 350-352 (1986).
16. Boyle, E. A. & Keigwin, L. D. *Earth planet. Sci. Lett.* **76**, 135-150 (1985).
17. Boyle, E. A. & Keigwin, L. D. *Science* **218**, 784-787 (1982).
18. Oppo, D. & Fairbanks, R. G. *Earth planet. Sci. Lett.* (in the press).
19. Bainbridge, A. E. *GEOSECS Atlantic Ocean Atlas* (US Government Printing Office, Washington, 1980).
20. Craig, H., Broecker, W. S. & Spencer, D. W. *GEOSECS Pacific Ocean Atlas* (US Government Printing Office, Washington, 1982).
21. Weiss, R., Broecker, W. S., Craig, H. & Spencer, D. W. *GEOSECS Indian Ocean Atlas* (US Government Printing Office, Washington, 1982).
22. Boyle, E. A. *Geochim. cosmochim. Acta* **50**, 265-276 (1985).
23. Boyle, E. A. *Palaeoceanography* (in the press).
24. Boyle, E. A. *Am. geophys. Un. Monogr.* **29**, 360-368 (1984).
25. Stalcup, M. C., Metcalf, W. G. & Johnson, R. G. *J. mar. Res. Suppl.* **33**, 15-35 (1975).
26. Richards, F. A. *J. mar. Res.* **17**, 449-465 (1958).
27. CLIMAP project members *Quat. Res.* **21**, 123-124 (1984).
28. Duplessy, J.-C. & Shackleton, N. J. *Nature* **316**, 500-507 (1985).
29. Zahn, R., Winn, K. & Sarnthein, M. *Palaeoceanography* **1**, 27-42 (1986).
30. Warren, B. A. *J. mar. Res.* **41**, 327-347 (1983).
31. Reid, J. L. *Johns Hopkins Oceanographic Studies no. 2* (Johns Hopkins Press, Baltimore, 1965).
32. McIntyre, A. *et al. Mem. geol. Soc. Am.* **145**, 43-76 (1976).
33. CLIMAP project members *Geol. Soc. Am. Map and Chart Ser.* No. 36 (1981).
34. Manabe, S. & Broccoli, A. J. *J. geophys. Res.* **90**, 2167-2190 (1985).
35. Zahn, R., Sarnthein, M. & Erlenkeuser, M. *Palaeoceanography* (in the press).
36. Bryden, H. & Stommel, H. *Oceanol. Acta* **7**, 289-296 (1984).
37. Held, I. M. in *Circulation in the Atmosphere* (eds Hoskins, B. & Pierce, B.) (Academic, New York, 1982).
38. Ku, T.-L., Bischoff, J. L. & Boersma, A. *Deep-Sea Res.* **19**, 233-247 (1972).
39. Erickson, D. B. in *The Sea* Vol. 1 (ed. Hill, M. N.) (Interscience, New York, 1963).