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Surface and deep hydrology of the Northern Atlantic Ocean during the past 150000 years

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SUMMARY

The abrupt shifts in foraminiferal $\delta^{18}O$ observed in core ODP 609 (the meltwater signature of the Heinrich events, see Bond *et al.* 1992*b*, 1993) are seen in ten North Atlantic high sedimentation rate cores; the decreasing south-west to north-east gradient is well pronounced. This confirms that the Heinrich events are associated with major surges of the Laurentide ice sheet, when it is believed approximately 10^6 km^3 of ice are liberated during each event.

A tentative reconstruction of the changes in surface and deep-water density, based upon the study of cores SU 90-39 (53° N 22° W) and SU 90-08 (43° N 30° W), is presented. To calculate the density of surface water, sea surface temperature is obtained using a foraminiferal transfer function (see CLIMAP 1981) and salinity is estimated using the foraminiferal $\delta^{18}O$ record corrected for the temperature effect on isotopic fractionation. The density of deep water is directly derived from the benthic $\delta^{18}O$ record, after corrections for the mean global changes in Ocean $\delta^{18}O$. Results indicate that the North Atlantic Ocean has been repetitively a potential area of deep-water formation during the last glacial period.

1. INTRODUCTION

Broecker & Denton (1989), among others, have proposed that the changes in Northern Atlantic surface salinity driven by the growth and decay of the large ice sheets may play a significant role in the control of deep-water circulation. Rapid changes in North Atlantic surface-water hydrology are recorded in the oxygen isotopic ratio of fossil planktic foraminifera from both the last deglaciation (Duplessy *et al.* 1986; Duplessy *et al.* 1992) and the last glacial period (Bond *et al.* 1992*b*; Bond *et al.* 1993). Most of these isotopic events are linked to catastrophic surges of the continental ice sheets (the Heinrich events, from Heinrich (1988) and Bond *et al.* (1992*b*). Rapid changes may also have occurred during the last interglacial period (Dansgaard *et al.* 1993; McManus *et al.* 1994; Keigwin *et al.* 1994). The purpose of this paper is to study the evolution of the surface-water hydrology in the North Atlantic Ocean and its implications for deep-water formation over the last glacial–interglacial cycle. Surface hydrological characteristics are reconstructed from high resolution planktic $\delta^{18}O$ and sea surface temperatures (SST) (derived from foraminiferal transfer functions following Imbrie & Kipp 1971). Deep-water density is derived from benthic foraminifera $\delta^{18}O$ (Labeyrie *et al.* 1987, 1992).

2. STRATEGY

Foraminiferal $\delta^{18}O$ records monitor the changes in surface and deep-water hydrology as well as the changes in ice volume. Oxygen isotopic fractionation between foraminiferal shells and the ambient water is a known function of the growth temperature (O'Neil *et al.* 1969; Shackleton 1974) and the water $\delta^{18}O$, which is itself a function of the water salinity as ^{18}O is fractionated against ^{16}O during evaporation: $\delta^{18}O_{water}$ changes locally by 0.5 per salinity unit for surface salinity gradients (Ostlünd *et al.* 1987), and, globally, by 1.2 per salinity unit during continental ice sheet growth and decay (Labeyrie *et al.* 1987; Shackleton 1987). Labeyrie *et al.* (1986) and Duplessy *et al.* (1991) have shown that salinity may be derived from the planktic foraminifera $\delta^{18}O$ during periods of known ice sheet volume, if SST is independently estimated from the foraminiferal species distribution (CLIMAP 1981).

3. DATA

The core location is plotted in figure 1. The location and the list of parameters which were measured are given in table 1. Three cores have been analysed isotopically for benthic foraminifera (SU 90-08,

Table 1. *Location of the cores and parameters presented here.*

core	latitude	longitude	data presented
SU 90-08	43° N	30° W	planktonic and benthic foraminifera δ^{18}O, SST, grey reflectance, ^{14}C AMS
SU 90-11	44° N	40° W	planktonic and benthic foraminifera δ^{18}O, grey reflectance
ODP 609	50° N	24° W	*N. pachyderma* (*s.*) δ^{18}O, grey reflectance
SU 90-39	53° N	22° W	planktonic foraminifera δ^{18}O, grey reflectance
V 23-81	54° N	16° W	*N. pachyderma* (*s.*) δ^{18}O
NA 87-22	56° N	15° W	planktonic and benthic foraminifera δ^{18}O, ^{14}C AMS
CH 73-110	60° N	09° W	*N. pachyderma* (*s.*) δ^{18}O
SU 90-32	60° N	22° W	*N. pachyderma* (*s.*) δ^{18}O
SU 90-16	58° N	45° W	*N. pachyderma* (*s.*) δ^{10}O
HU 75-42	62° N	54° W	*N. pachyderma* (*s.*) δ^{18}O
SU 90-44	50° N	17° W	*N. pachyderma* (*s.*) δ^{18}O

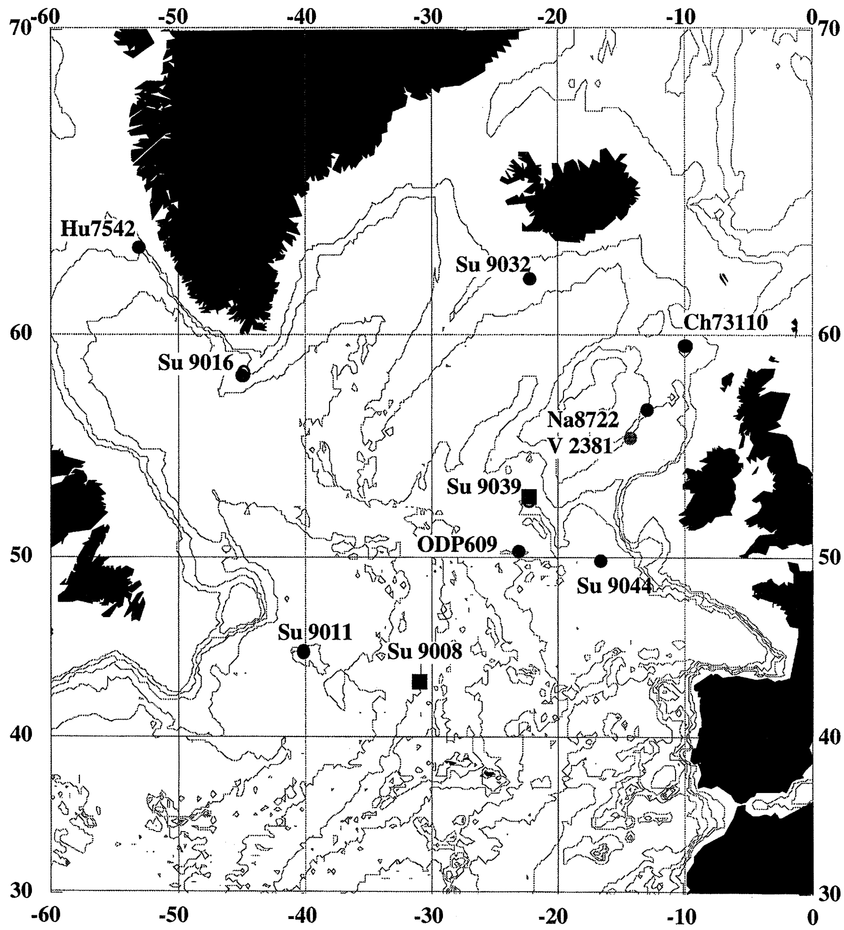

Figure 1. Location of the cores; square boxes indicate the location of the two key records for the study of vertical convection (cores SU 90-08 and SU 90-39).

SU 90-11, NA 87-22); and they constitute the basis for the chrono-stratigraphy and deep-water reconstruction. Core SU 90-39 and SU 90-08 have summer SST estimates, whereas planktic foraminifera δ^{18}O records are presented for all ten cores. The isotopic record from core HU 75-42 is a composite signal of two nearby cores, HU 75-41 (for the Holocene and deglaciation) and HU 75-42 for the glacial period (see Fillon & Duplessy 1980). Cores V 23-81 and ODP 609 have been extensively studied at very high resolution (2–3 cm) by Bond *et al.* (1992*b*, 1993) for the reconstruction of surface-water variability during the last glacial period in relation with the Heinrich events.

(*a*) *Isotopic measurements*

Foraminiferal δ^{18}O values have been measured for most cores on an automated preparation line coupled to a Finnigan MAT 251 mass spectrometer. The mean external reproducibility of powdered carbonate standards is ± 0.05‰ for oxygen and carbon. Data is reported as δ^{18}O versus PDB (Pee Dee Belemnites Standard; see Hut 1987; Coplen 1988). Respectively, planktonic foraminifera species *Globigerina bulloides* and *Neogloboquadrina pachyderma* (*s.*) were analysed in the 250–315 μm and 200–250 μm size range, the most abundant in north Atlantic sediments over the last

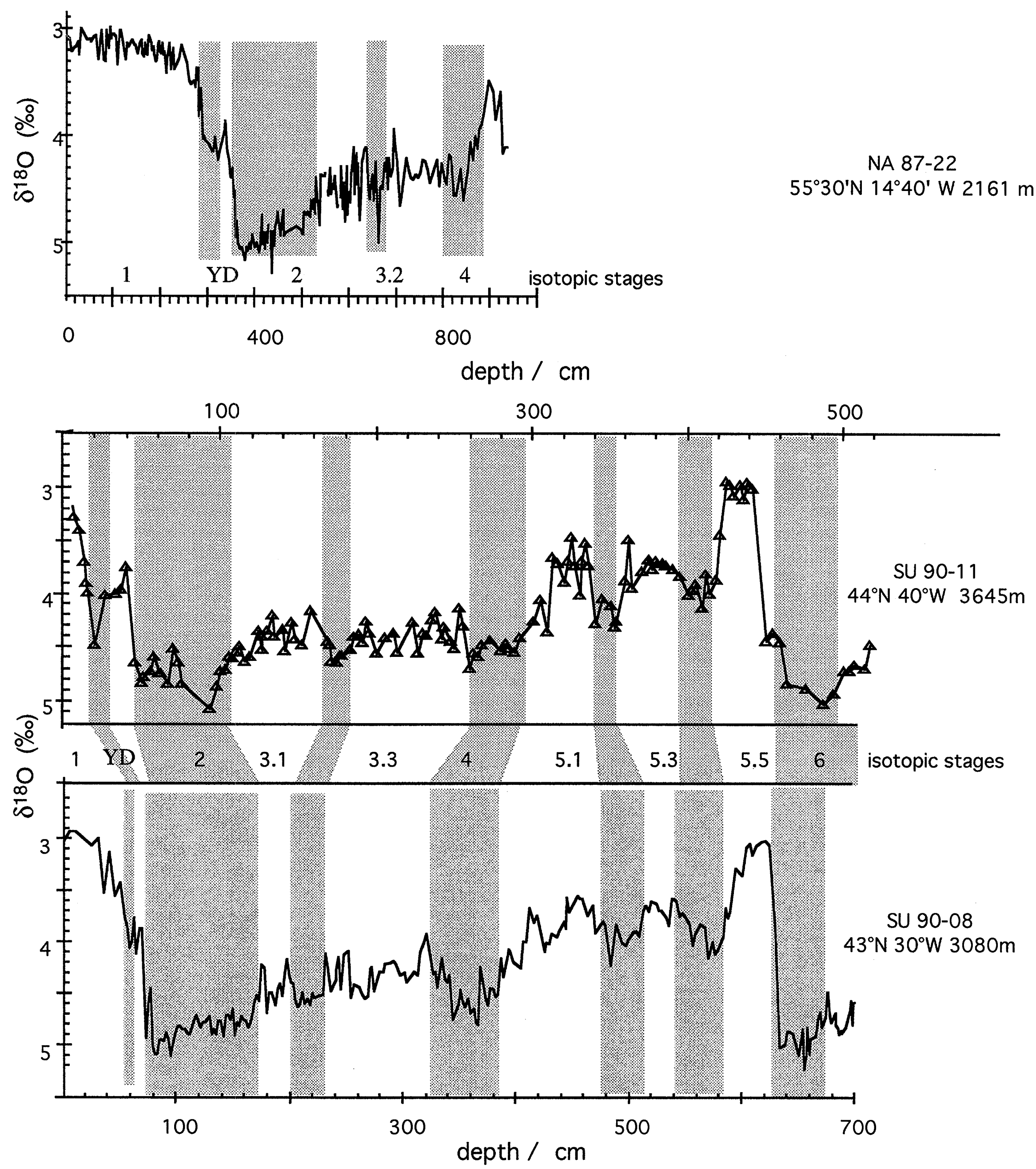

Figure 2. $\delta^{18}O$ records of the benthic foraminifera versus depth in the cores SU 90-11, SU 90-08 & NA 87-22.

climatic cycle. Wherever possible, the benthic foraminifera *Cibicides wuellerstorfi* was selected. When not available, we chose *Uvigerina peregrina*. *C. wuellerstorfi* $\delta^{18}O$ are adjusted by $+0.64‰$ for specific fractionation (Duplessy *et al.* 1984). The benthic $\delta^{18}O$ signals were analysed versus depth for cores SU 90-08, SU 90-11 and NA 87-22 in figure 2.

(*b*) *Sea surface temperature*

The SSTs were estimated on cores SU 90-08 (43° N) and SU 90-39 (53° N) using the classic Imbrie & Kipp (1971) method and a reference data base extended after CLIMAP (1981). Communality is always bigger than 0.8. The sampling resolution is not constant throughout the cores: Core SU 90-08 is analysed approximately each 2 cm (0.2–0.6 ka), except between 400–540 cm (isotopic stages 5.4 to 4), where the resolution was decreased to 10 cm (approx. equals 1–2 ka). Core SU 90-39 is only analysed at high resolution (2 cm 0.1–0.2 ka) during the last interglacial period (700–830 cm). Elsewhere, the resolution is 10 cm (1–2 ka). Estimated summer and winter SST (see figure 3) are similar to Ruddiman & McIntyre's 1984 reconstruction of nearby locations.

(*c*) *Construction of representative planktic foraminifera* $\delta^{18}O$

Drastic changes of climate during the last glacial–interglacial cycle in the northern Atlantic resulted from the displacement of the warm subtropical waters between 60° and 40° N (Ruddiman & McIntyre 1984). It is rarely possible to obtain, in this area, continuous

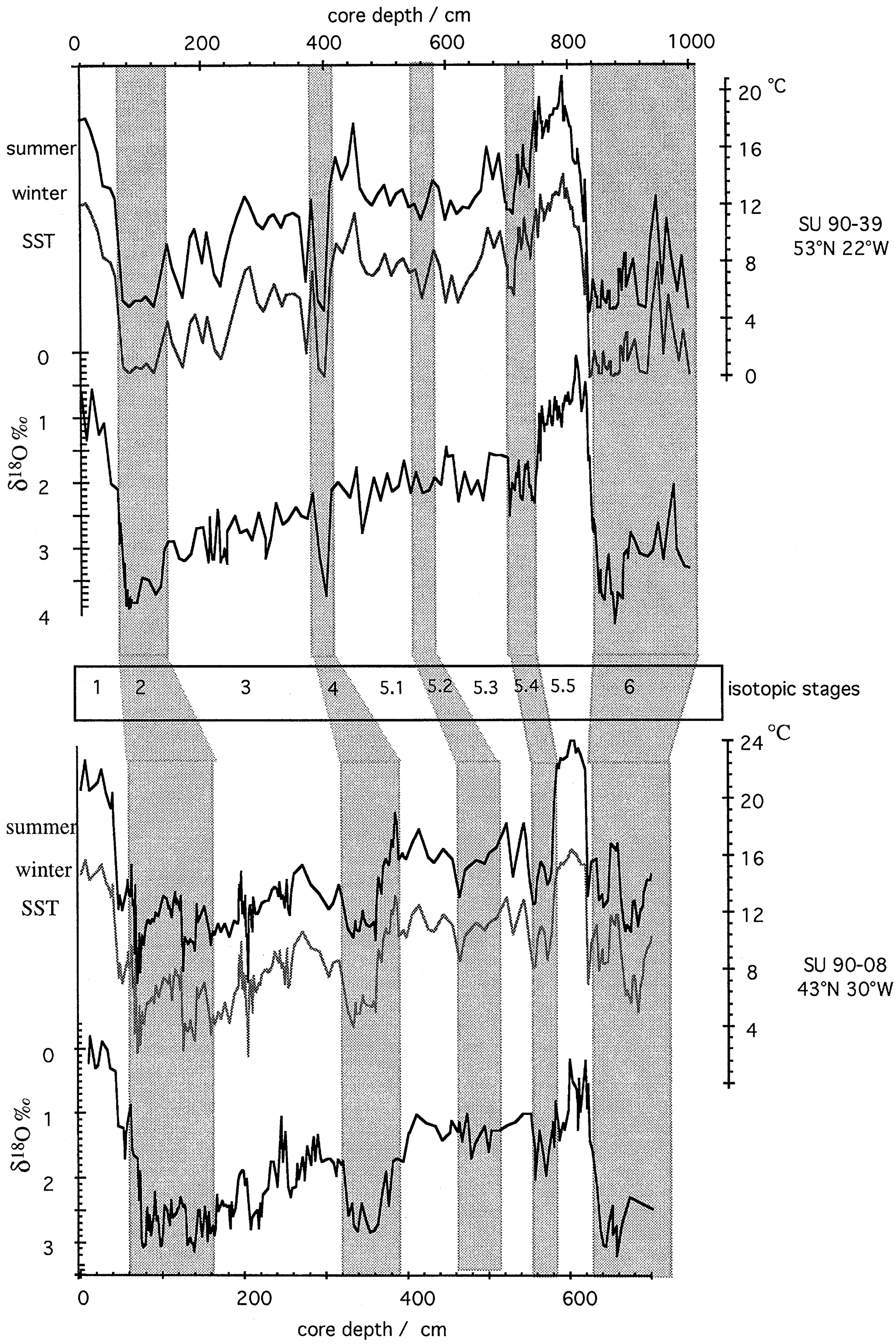

Figure 3. Results of the SST and planktonic foraminifera $\delta^{18}O$ reconstructions for cores SU 90-39 and SU 90-08 versus depth.

monospecific $\delta^{18}O$ records from *G. bulloides* or *N. pachyderma* (*s.*), as *N. pachyderma* (*s.*) is abundant in polar waters whereas *G. bulloides* lives in sub-polar and transitional waters. Duplessy *et al.* (1991) have observed that in waters where both foraminifera develop (between 7–10 °C summer SST), *N. pachyderma* (*s.*) is 0.35‰ heavier (corresponding to a growth temperature 1.5 °C colder). This isotopic offset (probably due to the deeper habitat of *N. pachyderma* (*s.*)) appears constant for the whole 0–10 °C optimum range of that species. We therefore reconstructed representative planktic foraminifera $\delta^{18}O$ records for the two cores where SST records are available (SU 90-08, SU 90-39), using *G. bulloides* $\delta^{18}O$ during warm periods (summer

Table 2. *AMS ^{14}C dating for core SU 90-08*

N. pachyderma (*s.*) depth	AMS^{14}C age ka	age (−0.48) ka	error ±1σ
60	13.44	12.96	0.11
72	15.23	14.75	0.11
80	15.28	14.8	0.1
100	18.73	18.25	0.15
120	21.1	20.62	0.21
134	22.45	21.97	0.2
162	27.48	27	0.33
170	30.09	29.61	0.51
192	33.85	33.37	0.66
210	36.13	35.65	0.88

SST ⩾ 7 °C), and *N. pachyderma* (*s.*) δ^{18}O (corrected by −0.35°) during cold periods (summer SST ⩽ 10 °C) (see figure 3).

Our results, here again, are in agreement with results from Ruddiman & McIntyre (1979, 1984); the cooling of surface waters lags the planktic foraminifera δ^{18}O during periods of continental ice growth (transitions 5.5 -5.4, 5.3-5.2, 5.1-4 and 3-2). The surface layer of warm waters which covered the Northern Atlantic during these periods was subjected to high evaporation (shifting the planktic foraminifera δ^{18}O to heavy values). Atmospheric water vapour was available for feeding precipitation over the ice fields, which promoted faster growth of the ice sheets.

(*d*) *Chronostratigraphy*

Spatial reconstruction of past changes in surface-water hydrology are based upon sets of records built on the same chrono-stratigraphical scale. Large changes in sedimentation rates occur in the North Atlantic, both spatially and with time.

(i) *Correlation*

As a first step, all the cores have been correlated using the ANALYSERIES software (see Paillard *et al.* CFR internal report 1993). The common stratigraphic scale was built using core SU 90-08 as reference. Approximately 10–20 depth-to-depth links were used for each core, corresponding to the most significant events. The parameters considered are listed below.

1. Reference benthic foraminifera δ^{18}O records (on the three cores SU 90-08, SU 90-11, and NA 87-22, see figure 2). They are identical within sampling and analytical variability for cores in the same depth range and general area.

2. *N. pachyderma* (*s.*) δ^{18}O and δ^{13}C records (on all cores) for the glacial period and *G. bulloides* δ^{18}O for the interglacials. The δ^{13}C study will be reported elsewhere (L. Labeyrie, unpublished data).

3. *N. pachyderma* (*s.*) relative abundance has been used by Bond *et al.* (1992*b*, 1993) for detailed comparison of nearby cores V 23-81 and ODP 609.

4. Grey reflectance records, using the methodology of Bond *et al.* (1992*a*). The results will be presented elsewhere (E. Cortijo, unpublished data).

Table 3. *AMS ^{14}C dating for core NA* 87-22

depth cm	AMS^{14}C age ka	age (−0.48) ka	error ±1σ	species
95	3.52	3.04	0.09	*N. pachyderma* (*s.*)
200	8.28	7.80	0.11	*G. bulloides*
200	8.30	7.82	0.09	*N. pachyderma* (*s.*)
250	10.01	9.53	0.09	*G. bulloides*
270	10.87	10.39	0.14	*G. bulloides*
285	11,32	10.84	0.09	*N. pachyderma* (*s.*)
305	11.69	11.21	0.12	*G. bulloides*
315	12.92	12.44	0.12	*G. bulloides*
325	12.38	11.90	0.12	*G. bulloides*
345	14.58	14.10	0.13	*N. pachyderma* (*s.*)
355	15.66	15.18	0.12	*N. pachyderma* (*s.*)
370	15.80	15.32	0.14	*N. pachyderma* (*s.*)
380	16.70	16.22	0.14	*N. pachyderma* (*s.*)
400	17.59	17.11	0.14	*N. pachyderma* (*s.*)
420	18.11	17.63	0.18	*N. pachyderma* (*s.*)
440	19.31	18.83	0.27	*N. pachyderma* (*s.*)
455	20.16	19.68	0.23	*N. pachyderma* (*s.*)
485	20.77	20.29	0.21	*N. pachyderma* (*s.*)
495	21.42	20.94	0.23	*N. pachyderma* (*s.*)
510	23.78	23.30	0.27	*N. pachyderma* (*s.*)
520	23.62	23.14	0.3	*N. pachyderma* (*s.*)
520	24.73	24.25	0.31	*G. bulloides*
530	26.35	25.87	0.33	*G. bulloides*
560	28.68	28.20	0.44	*N. pachyderma* (*s.*)
585	30.40	29.92	0.52	*G. bulloides*
595	31.64	31.16	0.64	*G. bulloides*
660	43.40	42.92	2.5	*N. pachyderma* (*s.*)

(ii) *The chronological scale*

^{14}C accelerated mass spectroscopy (AMS) dating of *N. pachyderma* (*s.*) and *G. bulloides* are available for cores SU 90-08 and NA 87-22 (see tables 2 and 3). The results have been corrected for a mean ventilation age of surface waters of 0.48 ka (Bard 1988). No correction for calibration to calendar age was applied. The time scale was derived by polynomial regression between the dating of both cores after the stratigraphic correlation. Results are compatible with the ^{14}C AMS ages published by Bond *et al.* (1992*b*; 1993) for core ODP 609.

Chronostratigraphic constraints below 30 ka are based upon the SPECMAP timescale (Martinson *et al.* 1987), with small changes introduced within isotopic stage 3, for which we used the chronostratigraphy based upon high precision tephrochronology of Mediterranean Sea ash layers (Paterne *et al.* 1986).

4. DISCUSSION

(*a*) *Spatial and temporal variability of the glacial* N. pachyderma(s.) *δ^{18}O records*

The *N. pachyderma* (*s.*) δ^{18}O records analysed age are reported for all the cores in figure 4. Large amplitude δ^{18}O negative anomalies are observed in most cores, and specially cores SU 90-08, SU 90-11 and ODP 609. Bond *et al.* (1992*b*; 1993) and Grousset *et al.* (1993) have demonstrated that the isotopic anomalies occur simultaneously with drastic surface-water cooling, and are associated with the Heinrich events of large influx

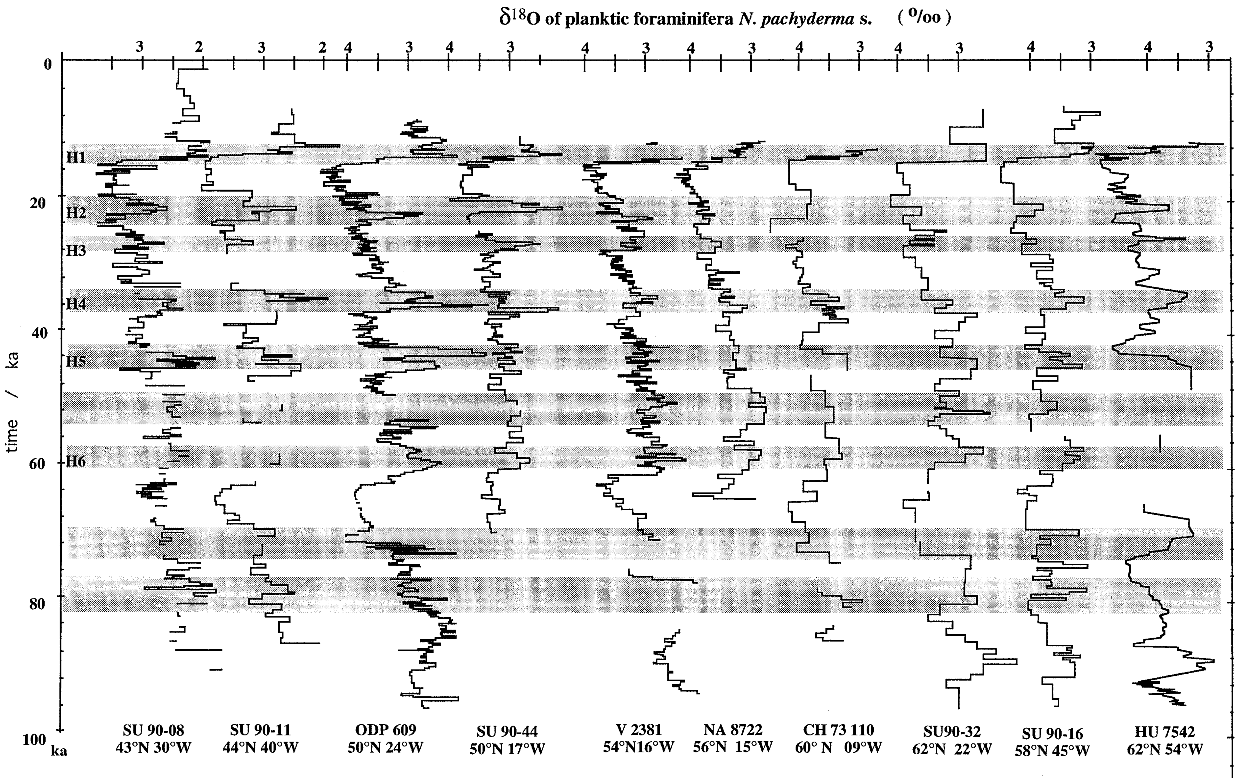

Figure 4. *Neogloboquadrina pachyderma* (*s.*) $\delta^{18}O$ records plotted against age for all the cores discussed in the text. The Heinrich events are indicated by grey shadings, according to the numbering of Bond *et al.* (1992*b*).

of ice rafted detritus. Therefore, they are the signature of large inputs of low $\delta^{18}O$ water melting from the icebergs during major ice surges of the continental ice sheets. Our results show that these large $\delta^{18}O$ anomalies extend over a significant portion of the Northern Atlantic, from core SU 90-11 (44° N 40° W and HU 75-42 (62° N 54° W) to SU 90-44 (50° N 17° W). They are also observed, with a slightly smaller amplitude (approx. equals 1‰ versus approx. equals 1.5–2‰), in core SU 90-08 (43° N 30° W) and SU 90-32 (62° N 22° W). The amplitude of these events decreases sharply to the north east of ODP 609, in the cores located around Rockall Bank (NA 87-22, 56° N 15° W) and V 23-81 (54° N 16° W).

Low $\delta^{18}O$ peaks of smaller amplitude (approx. equals 0.5 to 1‰) are also observed in all the northern cores. In Labrador Sea core HU 75-42, most of the Heinrich events are recorded, with an amplitude around 1‰. In the eastern core CH 73-110, only peaks H4 and H5 are significant. Cores SU 90-32 and SU 90-16, although at lower resolution, show most of the variability seen in the more southerly cores.

The distribution of the isotopic anomalies follows the trends mapped by Grousset *et al.* (1993) using the magnetic susceptibility signature of the ice rafted detritus derived from the Laurentide ice sheet. East of the Rockall Bank, cores NA 87-22 and V 23-81 do not show the isotopic signature of these events, indicating that surface water was not diluted by fresh meltwater. However, in the North East core SU 90-32 shows isotopic anomalies, but without the large magnetic susceptibility peaks seen in the 40–55° N latitudinal band. The $\delta^{18}O$ anomalies derive, therefore, from either surface water carrying meltwater from southerly latitudes, or from icebergs which originated from the Fenno-Scandinavian or Icelandic ice sheets.

It seems then that the isotopic anomalies described in core ODP 609 (Bond *et al.* 1993) represent a general phenomenon over the Northern North Atlantic Ocean. The largest amplitude signal (core SU 90-11 and ODP 609) probably tracks the major path of melting icebergs and the associated surface waters spiked by the low $\delta^{18}O$ ice sheet water. When averaged over the North Atlantic Ocean between 42° N and 65° N (approximately 4×10^6 km^2), taking 0.2 km as the mixing depth, a $\delta^{18}O$ of melting iceberg -35‰ and -0.9‰ for each isotopic anomaly, about 2.10^4 km^3 of ice was melting during the Heinrich events. If the duration of the events is taken as 50 times the mixing rate of surface waters (0.5–1 ka (Bond *et al.* 1993; François & Bacon 1994)), at least 1 million km^3 of ice was stripped from the ice sheets during each of the events: this corresponds to about 2 m of mean sea level increase.

An other observation may be gained from examining the higher resolution records. They present similar and synchronous trends once the large amplitude shifts are excluded; both the shape and timing of the substructure (a saw tooth pattern of about 7–10 ka duration) are also similar. This is near the value that MacAyael (1993) and Alley & MacAyeal (1994) estimated for the pseudo-periodicity of the Laurentide ice sheet oscillations under the effect of basal geothermal heating.

(*b*) *The hydrological changes*

(i) *Deep waters*

Labeyrie *et al.* (1987; 1992) have proposed a strategy to reconstruct the mean deep water $\delta^{18}O$, temperature

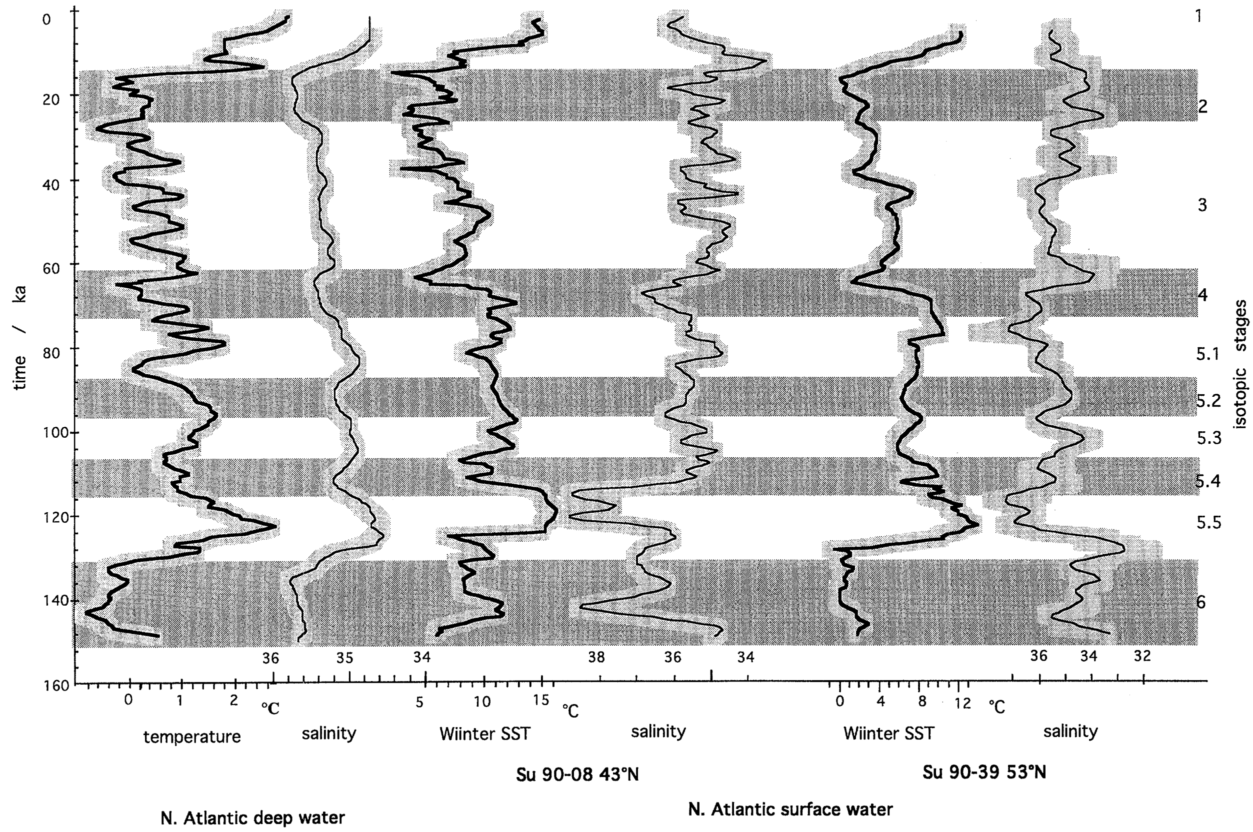

Figure 5. Calculated records of North Atlantic surface and deep temperature and salinity plotted for the last 150 ka, at the location of core SU 90-39 and SU 90-08. The salinity scale is inversed. The deep-water record is derived from the benthic $\delta^{18}O$ record of core SU 90-08. Grey shadings approximate the estimated mean uncertainty for each parameter. Isotopic stages are reported as reference.

and density. For the present work, the mean Ocean $\delta^{18}O$ record was extended to 150 ka BP using the V19–30 benthic $\delta^{18}O$ signal of Shackleton *et al.* (1983), after application of a $-0.35‰$ correction for the effect of changes in deep-water temperature during the transition between isotopic stages 6 and 5.5 (Sowers *et al.* 1993).

The global seawater $\delta^{18}O$ signal may be directly translated in terms of mean global salinity, as salinity is linearly correlated with continental ice volume: 1.04 salinity unit corresponds to 120 m sea level change and 1.2‰ change in mean Ocean $\delta^{18}O$ (Labeyrie *et al.* 1987; Shackleton 1987). The signal obtained by subtraction of this Ocean $\delta^{18}O$ record from the benthic foraminifera $\delta^{18}O$ record of core SU 90-08 represents the changes in oxygen isotopic fractionation, i.e. in deep-water temperature, at the core location. In such reconstruction we neglect the secondary effects which could derive from spatial heterogeneity in deep-water salinity. We have reported in figure 5 the derived records for North Atlantic deep-water temperature and salinity (at 2500–3000 m).

(ii) *Surface water*

The reconstruction of the changes in surface-water hydrology for cores SU 90-39 (53° N) and SU 90-08 (43° N) was obtained from their $\delta^{18}O$ and SST records. Sea surface salinity was estimated following Duplessy *et al.* (1991). The reconstructed records of surface salinity and temperature are plotted in figure 5. The general structure of the salinity record is similar to the one described in Duplessy *et al.* (1992) for the Last Deglaciation; warm periods are associated with higher salinity (surface waters of sub-tropical origin) and cold periods with lower salinity surface waters (of polar origin). It is specially true for the Last Interglacial (about 120 ka BP) in core SU 90-08, and for the end of warm stage 5.1 (around 70 ka BP) in core SU 90-39. The salinity records (as well as the SST records) are however significantly different in both cores. Salinity of the lower latitude core SU 90-08 is more or less constant around 35.5 for most of the last 110 ka, and about 38 for stage 5.5. Salinity is about 0.5 units lower at the location of the 53 °N core, SU 90-39. This core does not show a large increase in salinity during 5.5; the latitudinal gradient thus increased considerably during that period. The large negative spike at the end of glacial stage 6 in core SU 90-39 and at the end of glacial stage 2 in core SU 90-08 correspond to other well defined differences.

The Last Interglacial, sampled for both cores at better than 1 ka resolution, presents a special interest, because of the large variability observed in the GRIP ice core during that period (Dansgaard *et al.* 1993). McManus *et al.* (1994) have shown that the ice signal does not seem compatible with the Northern Atlantic Oceanic record for that period. Our records however do indicate that around mid 5.5 salinity decreased by about 1 unit in both cores. This is compatible with a

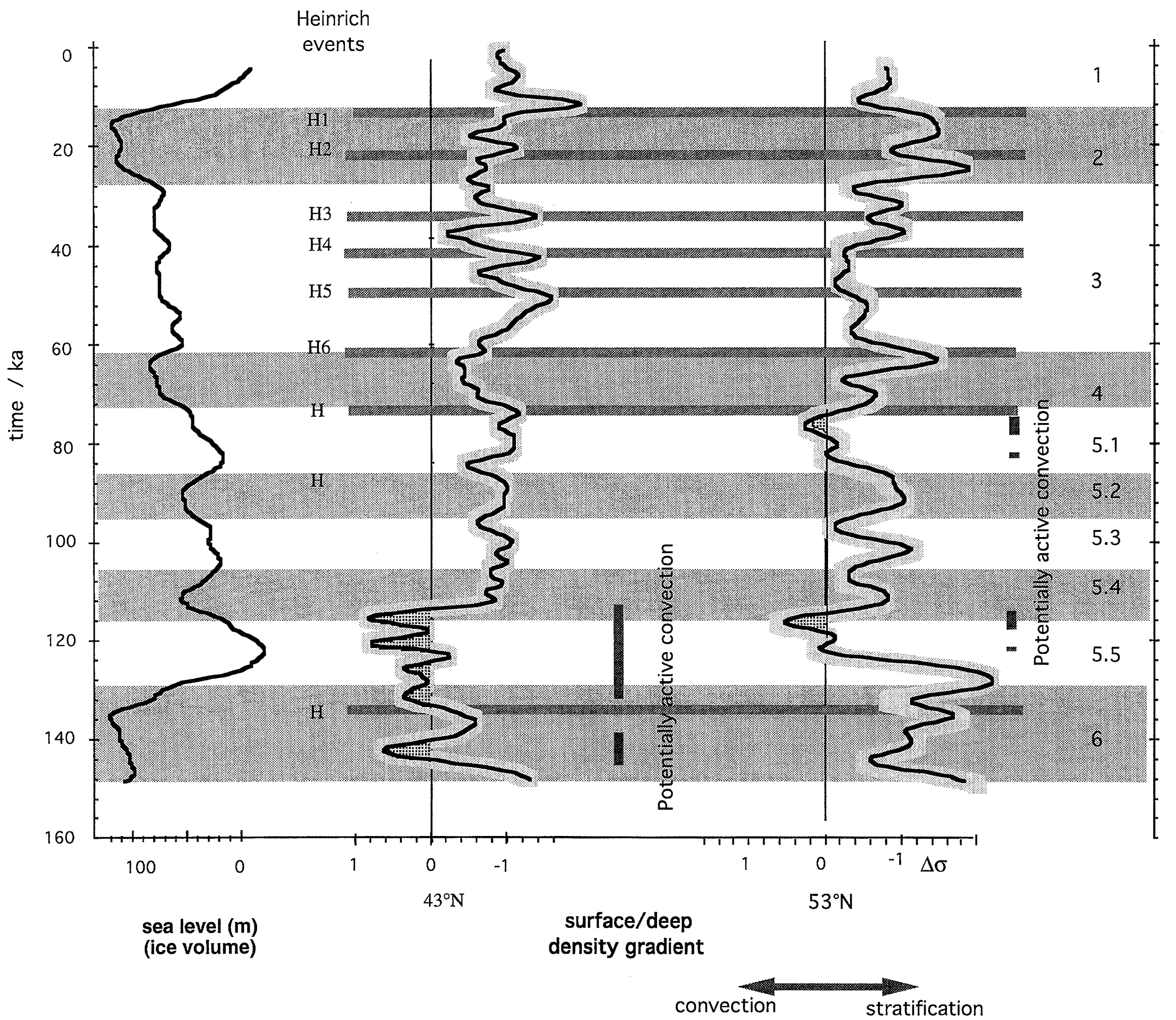

Figure 6. Calculated surface-deep density gradients at the location of core SU 90-39 (53° N) and SU 90-08 (43° N). The sea level curve is reported as reference. Shaded areas correspond to periods of potential excess density in surface waters. The H1–6 lines correspond to recognized Heinrich layers in the cores (after Bond *et al.* 1992*b*). Other events, identified as analogues to Heinrich layers during the cold events of stage 5 (McManus *et al.* 1994) are reported as H.

slowing down of the salt conveyor belt to the northern Atlantic, supporting the results of Cortijo *et al.* (1995). These authors have evidence for a drastic cooling and decrease in salinity in Norwegian sea during that period, which is more muted in the Northeastern Atlantic.

Just before the last interglacial, core SU 90-39 shows a large decrease in salinity. This meltwater event is supported by the lead of the planktic foraminifera $\delta^{18}O$ versus SST changes during the transition between isotopic stages 6 and 5 (see figure 3).

(*c*) *Potential areas and periods for deep-water convection*

The temperature and salinity may be used to calculate water density (σ_t) using the relation set out by Cox *et al.* (1970). Calculations were made for surface water in winter; the season in which deep-water convection usually occurs. Results are shown as changes in density gradient between winter surface and deep waters over the last 150 ka, at 43° N (core SU 90-08) and 53° N (Core SU 90–39) in figure 6. The sea level curve derived from the benthic foraminifera $\delta^{18}O$ record is reported as a reference on the same scale.

There are some uncertainties associated with the calculations (approximately ± 0.5 density units). We must, therefore, consider for now only the general trends. Vertical grey lines in figure 6 are periods where the calculated density for surface water exceeds the density of deep waters, i.e. potential periods of deep convection. It is evident that surface water never presented the large excess density given by the reconstruction during these periods. Planktonic foraminifera develop during the summer months mostly, especially in polar waters. The reconstruction indicates the periods during which surface water, with winter cooling, reaches deep-water density, at the origin of deep-water convection.

No general trends become apparent from these reconstructions. The area around 53° N may have been near sources of deep convection for part of stage 5.3 (100 ka), the cooling from mid 5.1 to 4 (75–65 ka),

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY B BIOLOGICAL SCIENCES

part of stage 3 (50 to 40 ka) and during the last deglaciation (12–13 ka). Apparently during LGM this area was sufficiently warm, taking into account its salinity, to have low density surface waters. This contrasts to the more northern surface waters (around 55° N 30° W) which were potential sources of deep-water convection during that period according to Labeyrie *et al.* (1992).

At 43° N, surface water density was generally lower (because of the warmer temperature), and the only large potential excess in surface density occurred during isotopic stages 6 and 5.5 (approximately 140 ka and 130–120 ka). The 5.5 excess density event presents an interesting problem. Deep-water formation was active during at least the first part of that period in Norwegian Sea (Cortijo *et al.* 1994). Our results point to a possible second source of deep-water convection driven by an excess salinity of the sub-tropical surface waters. It is impossible, however, from the available data to give its precise location and hydrological characteristics.

After the Last Interglacial, salinity may have been for limited periods sufficiently high to permit deep-water convection. This is apparently the case inbetween the ice surges (Heinrich events), and during short periods within the Last Interglacial. Analysis of other cores, from different areas, will be necessary to define a precise pattern in the shifts of possible areas of deep-water convection. But the results do indicate that during the last 150 ka, North Atlantic was frequently an area of deep-water convection. During most of the Glacial, the modern system of subtropical well-ventilated and salty Gulf Stream water flowing to Norwegian sea was replaced by more local productions of deep or intermediate waters, depending on the surface distribution of temperatures and salinity, therefore also on the direct coupling between atmosphere, ice sheets and the Ocean.

5. CONCLUSIONS

This study represents one approach to integration of the variation that exists in the planktic $\delta^{18}O$ records during the last climatic cycle over the range of the whole Northern Atlantic Ocean; it also discusses the possible shifts of conditions of deep-water formation. A special effort has been made to monitor the rapid variability of the surface water during the Heinrich events of the last glacial period, using the $\delta^{18}O$ of *N. pachyderma* (*s.*).

Our results demonstrate that during the Heinrich events, most of the surface of the Northern Atlantic was covered by a layer of low salinity water due to the large contribution of meltwater from icebergs. The track of the maximum anomaly confirms that the Laurentide ice sheet was one of the major sources of these icebergs. In agreement with the study by Maslin *et al.* (1995), our results indicate that deep-water convection was probably drastically reduced or stopped in the Northern Atlantic during these events and that in between, deep-water convection may have been rapidly re-established. Interglacial stage 5.5 presents another interesting period, with major changes in the latitudinal gradients in salinity. Our results confirm the results of Cortijo *et al.* (1995) that the initiation of the cooling of the Last Glacial, which occured in Norwegian Sea around 120 ka BP, was recorded also in the Northern Atlantic by a large decrease in surface salinity. Simultaneously, at 43° N, salinity increases drastically. This indicates a major slowing down of the surface thermohaline transport to the Northern Atlantic during that period.

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The complete data set will be deposited at time of publication in the World paleo-data centre in Boulder, in the European paleo-Data base in construction (N. Shackleton, Cambridge) and is available from L. L. This is CFR contribution 1691.

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