PERGAMON

Quaternary Science Reviews 21 (2002) 1153–1165

QSR

Changes in North Atlantic deep-water formation associated with the Dansgaard–Oeschger temperature oscillations (60–10 ka)

Mary Elliot[a,*], Laurent Labeyrie[b], Jean-Claude Duplessy[b]

[a] *Lamont Doherty Earth Observatory, Rt. 9W Palisades, New York, NY 10964, USA*

[b] *Laboratoire des Sciences du Climat et de l'Environnement, Laboratoire mixte CNRS-CEA, Av de la Terrasse, 91198 Gif sur Yvette, France*

Received 3 May 2001; accepted 30 August 2001

Abstract

We closely compared high-resolution $\delta^{13}C$ records of benthic foraminifera *Cibicides wuellerstorfi,* a proxy for deep-water ventilation, with the Dansgaard–Oeschger temperature oscillations. Our results reveal different perturbations of deep-water formation in the North Atlantic Ocean associated with the millennial-scale climate oscillations during the last glacial period. The cooling episodes associated with the drastic Heinrich events are related to large reductions of deep-water formation and a northward migration of ^{13}C depleted southern source deep waters to 62°N in the North Atlantic Ocean. The inter-Heinrich events which are correlated to the other cold stadials, are marked by significant changes of sea surface temperature around the Rockall Plateau, variations in the flux of icebergs to the North Atlantic Ocean but are not associated with such important reduction of deep-water formation. If changes in the thermohaline circulation (THC) are associated with these millennial-scale climatic oscillation they affect only the deeper water masses, below 2000 m, of the North Atlantic Ocean.

We thus show that equivalent degree of cooling over Greenland is obtained with different perturbations of deep-water formation. Our results either question the role of the THC as the unique explanation for these millennial-scale climate oscillations, or call upon an amplifying mechanism not yet taken into account.

1. Introduction

Numerous archives such as ocean sediment deposits and ice core records have given evidence of abrupt, high-frequency climate changes, which occur every 1000–3000 yr. The most prominent feature of this mode of climate variability can be observed in the Greenland temperature record of the last glacial (10–60 ka), namely the Dansgaard–Oeschger (DO) temperature oscillations (Dansgaard et al., 1993). These atmospheric temperature oscillations have been associated with coeval decreases of sea surface temperature (SST) in the sub-polar North Atlantic Ocean and periods of increased iceberg delivery from the continental ice sheets to the open ocean e.g. (Bond et al., 1993). Layers rich in ice rafted debris (IRD), some of which known as the Heinrich events (HEs) e.g. (Heinrich, 1988; Bond and Lotti, 1995), were formed as the icebergs drifted across the North Atlantic Ocean, melted and released the rock fragments they transported. The explanation of the millennial-scale ice sheet instabilities and the consequent freshwater pulses is still a matter of controversy. These episodes of ice sheet instability could either be a response to external (Renssen et al., 2000) or an internal climatic forcing e.g. (Bond et al., 1999; Cane and Clement, 1999; Clement and Cane, 1999) or could result of internal ice sheet dynamics (MacAyeal, 1993; van Kreveld et al., 2000).

However, most of the theories, which aim to explain the mechanisms of abrupt millennial-scale climate oscillations call upon changes in the global thermohaline circulation (THC) either as the result or the trigger of the observed ice sheet instabilities. Broecker (1997) proposed a conceptual model of a "bipolar sea saw", where the THC was able to swing from one mode of operation to another and could generate the observed frequent climate oscillations (see also (Broecker and Denton, 1989; Broecker et al., 1999). Numerical models of diverse complexity have since been used to simulate the impact of a freshwater perturbation in the North Atlantic Ocean on deep-water formation, the "Achilles Heel" of the global THC e.g. (Stocker and Wright, 1991;

*Corresponding author. Tel.: +1-845-365-8768; fax: +1-845-365-8165.

E-mail address: mary@ldeo.columbia.edu (M. Elliot).

0277-3791/02/$ - see front matter
PII: S0277-3791(01)00137-8

Rahmstorf, 1994; Manabe and Stouffer, 1995; Schiller et al., 1997; Ganopolski and Rahmstorf, 2001; Rind et al., in press). These studies have illustrated that freshwater pulses into the North Atlantic Ocean are able to reduce or shut down North Atlantic deep-water (NADW) formation.

The aim of this paper is to determine precisely if similar modifications of the THC are associated with each cooling, stadial event, recorded in the Greenland ice core. Previous studies of benthic foraminiferal paleo-nutrient proxies ($\delta^{13}C$, Cd/Ca, Zn/Ca) records have given evidence of millennial-scale variations. Changes of NADW formation have been documented during the HE (Vidal et al., 1997; Zahn et al., 1997) and on millennial-scale (Oppo and Lehman, 1995; Curry et al., 1999; Keigwin and Boyle, 1999). Fluctuations of deep-water circulation on millennial-scale have also been observed in the South Atlantic (Charles et al., 1996; Kanfoush et al., 2000) and in the North Pacific (Lund and Mix, 1998). Other proxies have also been used such as light reflectance (Chapman and Shackleton, 2000), grain size distribution (Bianchi and McCave, 1999), ^{14}C concentration (Hughen et al., 2000), and magnetic susceptibility (MS) (Rasmussen et al., 1996a; Kissel et al., 1999) (see Boyle (2000) for a review). However, most of these studies lacked the temporal constrain to precisely test the timing of the observed changes in deep-water formation with the Greenland temperature record. Keigwin and Boyle (1999) were the first to give evidence of changes in deep-water formation closely associated with the DO oscillations. Their interpretation is based on benthic $\delta^{13}C$ records obtained from a sediment core retrieved at 4500 m water depth close to the Bermuda Rise. Comparison with the DO cycles was done on the basis of a correlation of the carbonate content and the temperature record from Greenland ice cores.

We present two high-resolution records of benthic foraminifera $\delta^{13}C$ records, which we have been able to precisely compare to the DO cycles. We have focused our study on analyses of benthic foraminifera *C. wuellerstorfi* as this species best reflects changes in the chemistry of bottom waters (Duplessy et al., 1988; Sarnthein et al., 1994). Two sediment-cores from around 2000 m water depth in the North Atlantic Ocean have been used for this study. Sediment core results are closely compared with the Greenland ice core records by correlating MS and SST variations with the ice core temperature oscillations.

2. Data and methods

2.1. Site description and data

High-resolution records of $\delta^{13}C$ and $\delta^{18}O$ *C. wuellerstorfi*, IRD concentrations, SST estimates and MS have

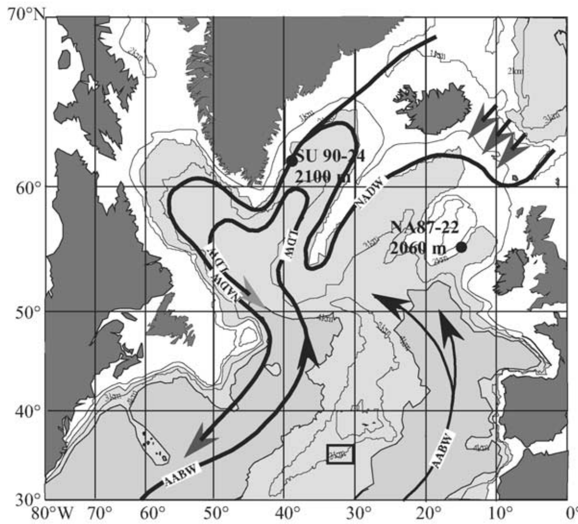

Fig. 1. Simplified map showing modern major deep water masses in the North Atlantic Ocean from (Dickson and Brown, 1994) and the position of the studied cores. The gray arrows correspond to the North Atlantic deep-water masses (NADW). The black arrows correspond to Antarctic deep waters (AADW) and the lower deep water (LDW) derived from the AABW.

been derived from two mid-depth sediment cores in the North Atlantic Ocean. Both cores presented here are well suited to monitor changes of NADW circulation at intermediate depth (Fig. 1).

Core SU 90-24 was retrieved off the coast of east Greenland (62°04′N; 37°02′W; 2100 m water depth). Modern deep-water masses at the core location correspond to the western branch of NADW. This well-ventilated deep water is formed by winter convection in the Greenland–Norwegian Seas and penetrates the North Atlantic Ocean via the Denmark Strait. This core is located beneath the path of the East Greenland Current, one of the major sources of freshwater to the North Atlantic Ocean in the present day configuration (Aagaard and Carmack, 1989). Core SU 90-24 has an average sedimentation rate of 16 cm/ka during the glacial, sampling was performed every 2 cm corresponding to an average interpolated temporal resolution of 125 yr. IRD and MS results have been presented in (Elliot et al., 1998; Elliot et al., in press). The abundance of benthic foraminifera *C. wuellerstorfi* was variable in this core and there are areas when sediments were devoid of this specific species, particularly between 300 and 400 cm (Fig. 2).

Core Na 87-22 (Cortijo et al., 1997; Vidal et al., 1997) is located on the eastern banks of the Rockall Plateau (55°29′N; 14°41′W; 2161 m water depth), on the flow path of upper North Atlantic deep water (UNADW). This core is located beneath the path of the North Atlantic current, responsible for bringing warm salty surface waters into the Nordic seas. Core Na 87-22 has an average sedimentation rate of 12 cm/ka during the glacial, average sampling interval for IRD is 5 cm

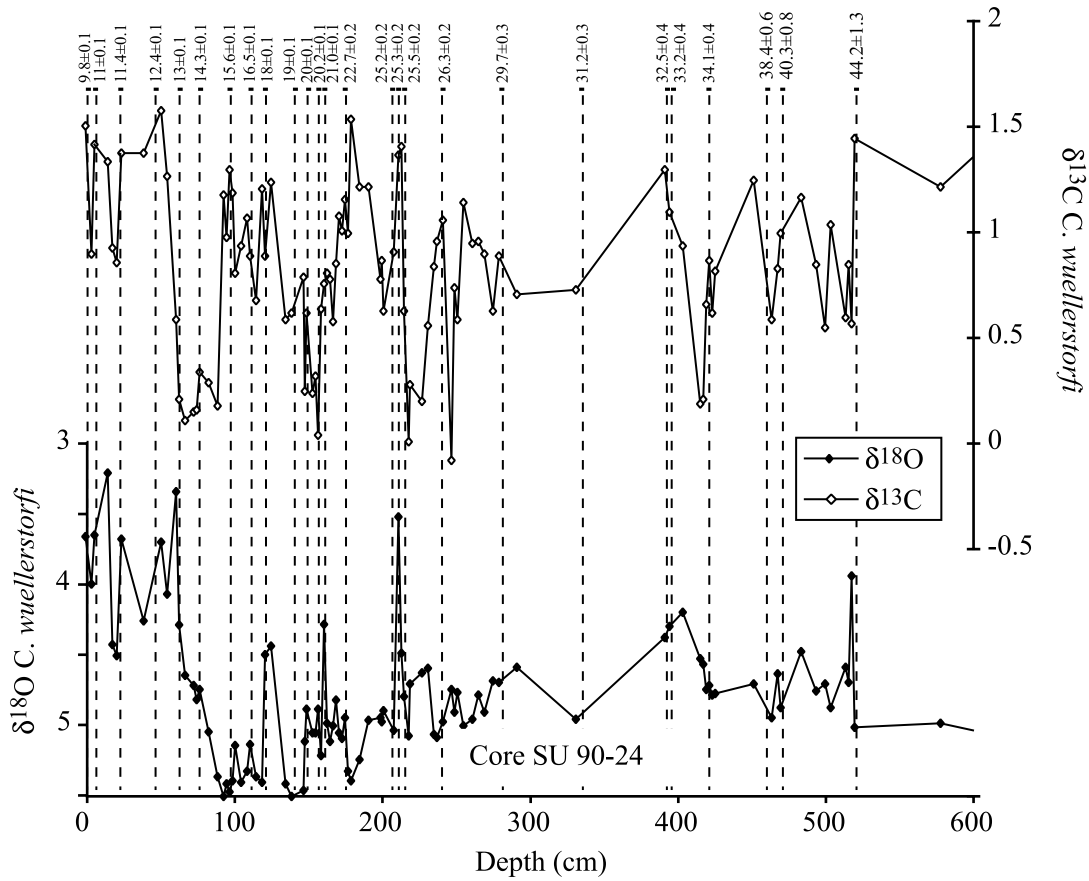

Fig. 2. Records of $\delta^{13}C$ (top) and $\delta^{18}O$ (bottom) of *C. wuellerstorfi* as a function of core depth. The position of the ^{14}C dates are indicated and are reported in Elliot et al. (1998). The $\delta^{18}O$ was corrected by +0.64‰ for specific fractionation of *C. wuellerstorfi* (Duplessy et al., 1984).

corresponding to an average temporal resolution of 400 yr. Isotope measurements and SST estimates have been obtained every 5–2.5 cm corresponding to an average interpolated temporal resolution of 400–200 yr.

The isotope measurements were performed at the Laboratoire des Sciences de l'Environnement et du Climat in Gif-sur-Yvette using an automated preparation line coupled to a Finnigan MAT 251 mass spectrometer. The mean external reproducibility of powdered carbonate standard is ± 0.05‰ for oxygen and carbon. Measurements were obtained from samples weighing a minimum of 40 μg corresponding on average to 2–3 foraminifera shells. The data are reported as per mil difference versus Pee Dee Belemnite standard (PDB) after calibration with National Bureau of Standards (NBS) 19 (Hut, 1987; Coplen, 1988).

2.2. Sea surface temperatures

SSTs in core Na 87-22 are based on planktonic foraminiferal counts of at least 300 individuals using the modern analog technique. Paleotemperatures are estimated by identifying the five most similar core top samples in the North Atlantic data-base (615 core top samples between 0 and 80°N, modified from Pflaumann et al. (1996). Summer and winter SSTs are then estimated by averaging the summer and winter SSTs associated with these most similar core tops (Prell, 1985). Dissimilarity between sample and core-top assemblages, using 32 planktonic taxa, is calculated using the chord distance. Uncertainties in the SST reconstructions correspond to the root mean square error of the top 5 analog temperatures. In each case, the 5 most similar core-top samples are accepted as valid modern analogs for the studied fossil sample, with dissimilarity coefficient lower than 0.2. The average error bars on temperature reconstructions are 1.5°C.

2.3. Age models and correlation with the DO oscillations

Preliminary age models based on ^{14}C dates have previously been presented in (Vidal et al., 1997; Elliot et al., 1998; Elliot et al., in press). However, the errors associated with ^{14}C dating of ocean sediment cores are

Table 1
Average age and duration of the HEs 1–5 calculated from 7 ocean sediment cores across the North Atlantic (SU 90-24, SU 9016 V2381 & ODP 609 e.g. (Bond and Lotti, 1995); Na 87-22 & SU 90-08 (Cortijo et al., 1997; Vidal et al., 1997) and CH69-K09 e.g. (Labeyrie et al., 1999)[a]

		Age (ka)	Std deviation (ka)	Duration (ka)	Std deviation (ka)
H1	Base	15.1	0.7	1.6	0.9
	Top	13.4	0.3		
H2	Base	22.1	0.8	1.7	0.9
	Top	20.4	0.1		
H3	Base	1.6	1.6	1.3	1.1
	Top	0.6	0.6		
H4	Base	34.9	1.1	0.8	0.4
	Top	33.9	0.7		

[a] These ages are presented in ^{14}C ka and have been determined using the same method as that presented in (Elliot et al., 1998), for more detail see also (Elliot, 1999). These ages have been obtained by assuming a constant reservoir age of 400 yr. There is evidence that the reservoir age may have varied temporally during these events (Waelbroeck et al., 2001) however this would affect the ^{14}C age of the IRD layer at all locations. Regional differences of ^{14}C reservoir ages across the North Atlantic during the HEs e.g. (Sarnthein et al., 2000) due to the distribution of the meltwater lens could account for some of the ^{14}C age differences observed at each location.

at least 1000 yr at 25–30 ka, due to the inherent error of the method, a lack of constrains on reservoir ages e.g. (Sarnthein et al., 2000; Waelbroeck et al., 2001). In order to compare the records, particularly during the HEs, we have adjusted our initial age models, within the error bars of ^{14}C dates. The age of the IRD layers corresponding to the HEs were constrained to fit their average age as defined from seven sediment cores in the North Atlantic Ocean (Table 1).

Our comparison of the benthic records of δ^{13}C with the DO oscillations are based on the correlation of MS and SST records with the Greenland ice core record. A wealth of data has given evidence that the changes of MS observed throughout the Nordic regions, in the Irminger Basin and south of Iceland are related to changes in deep-water circulation (Rasmussen et al., 1996a, b; Moros et al., 1997; Dokken and Jansen, 1999). Changes in the velocity of boundary currents marked by different size and concentration of magnetic minerals have been associated with the MS records and strongly favor this hypothesis (Kissel et al., 1999). Low MS values, which correspond to decreased velocity of the contour currents, have been correlated with cold stadials. Conversely, high MS have been correlated with warm interstadials e.g. (Rasmussen et al., 1996a, b; Kissel et al., 1999; Elliot et al., in press). We will use the MS record of core SU 90-24 (Fig. 4) as a means to compare the temperature oscillations in Greenland and the changes in benthic δ^{13}C record. Around the Rockall Plateau, the SST record show frequent millennial-scale oscillations of 3–5°C, corresponding to oscillations of the polar front which can also be correlated to the Greenland temperature record, as done by (Bond et al., 1993) (Fig. 3).

3. Results

δ^{13}C *C. wuellerstorfi* is commonly used as a tracer of changes in deep-water circulation as it reproduces the δ^{13}C of ΣCO_2 of ambient waters (Duplessy et al., 1988; Sarnthein et al., 1994). Several studies have shown that the δ^{13}C variations in the deep North Atlantic are mainly modulated by the relative flux of well-ventilated (high δ^{13}C) NADW and deep nutrient rich (low δ^{13}C) southern source waters (Boyle and Keigwin, 1982; Oppo and Lehman, 1995; Vidal et al., 1997).

δ^{13}C *C. wuellerstorfi* values of core Na 87-22 vary from +1.1‰ during late Holocene (data not shown see Vidal et al. (1997) to around +0.7 and +0.8‰ during the glacial maximum, between 15 and 19 ^{14}C ka (Fig. 3). In the Irminger Basin, Holocene deposits are absent in core SU 90-24 (Fig. 4). However, modern values of benthic δ^{13}C should be high ($\sim$ +1.3 to +1.4‰) as they derive directly from the overflow of deep waters formed in the Nordic Seas (Fig. 1) where benthic Holocene δ^{13}C values are around +1.3 to +1.5‰ (Veum et al., 1992; Bauch et al., 2001). The down-core record of δ^{13}C *C. wuellerstorfi* of core SU 90-24 shows values around +1 to +0.9‰ during the glacial maximum (Fig. 4). These results are in agreement with a shift of NADW formation to intermediate depth (above $\sim$2000 m) during the glacial maximum e.g. (Duplessy et al., 1988; Oppo and Lehman, 1993).

High-frequency changes can be observed. δ^{13}C *C. wuellerstorfi* of core Na 87-22 drop to values around +0.2‰ during H1 and around +0.4 to +0.5‰ for H3, H4 and H5 (Fig. 3). The abundance of *C. wuellerstorfi* is too low during the H2 interval to resolve a signal. In core SU 90-24, there are distinct and well-marked decreases of δ^{13}C *C. wuellerstorfi* during each the HE to +0.2‰, well below the values observed during the glacial maximum (Fig. 4).

On the Rockall Plateau, in the interval between H4 and H3, 3 periods of increased IRD deposit associated with cold SST can be identified. These detrital events are not systematically marked by decreases in δ^{13}C *C. wuellerstorfi* (Fig. 3). In some cases, for example the events centered on 31, 41 and 43 ^{14}C ka, the δ^{13}C *C. wuellerstorfi* increases to around +1.1‰ reflecting rather well ventilated waters at the sediment–water interface. Even when the δ^{13}C *C. wuellerstorfi* decreases the values remain around +0.7‰, well above the values observed during the HEs. The same observation can be made for the interval between H4 and H5, when the increased IRD deposits are associated with high values of around +1.1‰. In the Irminger Basin, similarly to

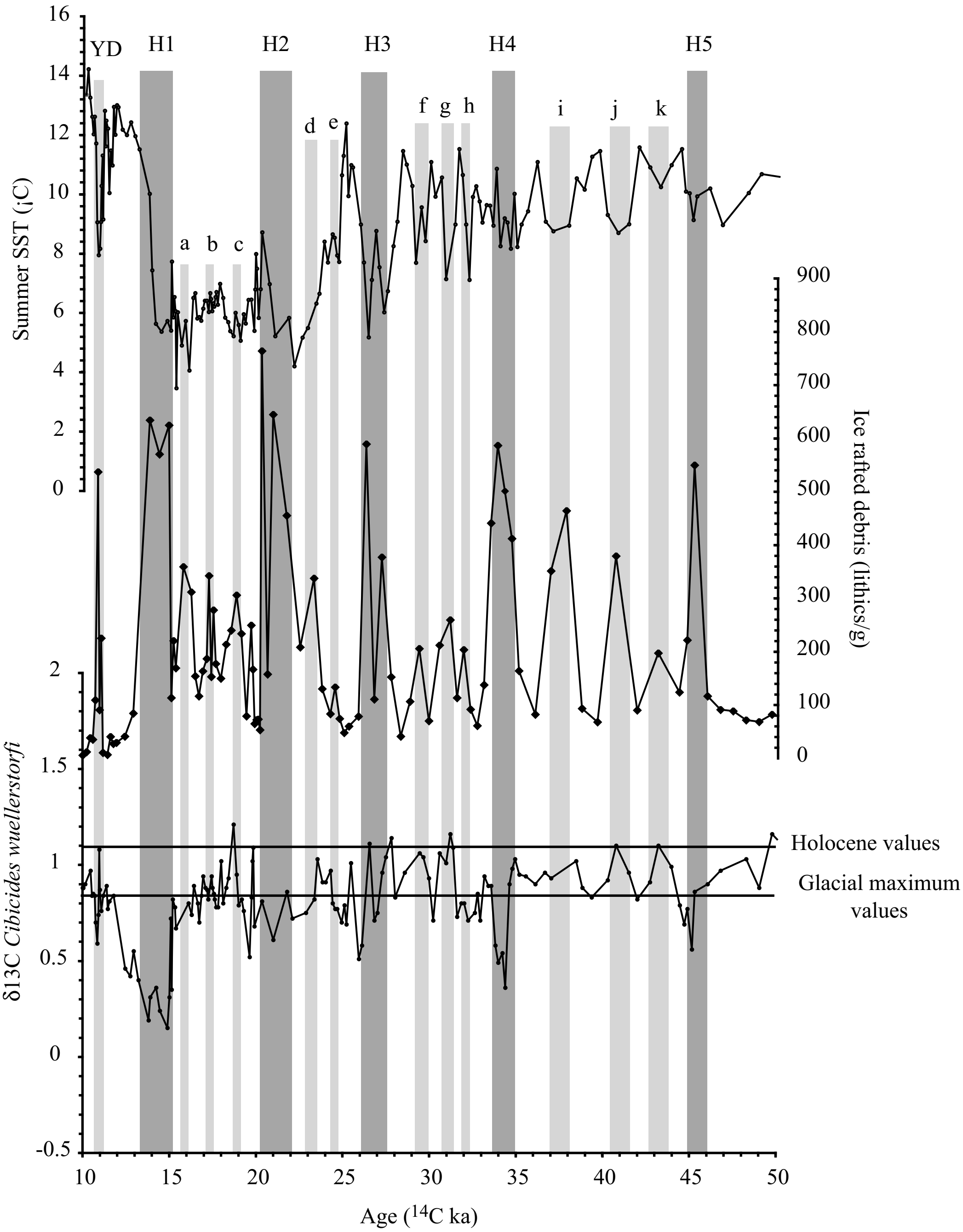

Fig. 3. Results of core Na 87-22 from the Rockall Plateau. Summer SST (top), concentration of IRD in number of lithic grains (>150 μm) per gram of dry sediment (middle), δ^{13}C *C. wuellerstorfi* (bottom). A lower resolution record of IRD abundance were previously published in (Cortijo et al., 1997; Vidal et al., 1997) but were expressed in relative percentage of IRD versus total entities. We have recounted the IRD abundance with higher temporal resolution using the method described in Elliot et al. (1998). The variations in IRD expressed in number of lithic fragments (>150 μm) per gram of dry sediment clearly reveal high-frequency variations of ice rafting episodes not seen in the previous record. An initial age model was based on 25 ^{14}C dates between 0 and 44 ^{14}C ka (see Cortijo et al., 1997; Vidal et al., 1997). We present adjusted ages for the Heinrich layers which were constrained to fit the average age as determined from seven sediment cores from the North Atlantic Ocean (see Table 1). These adjustments are done within the errors of the ^{14}C dates. The dark bands highlight the periods corresponding to the average age and duration of the Heinrich layers as defined in Table 1. The light gray bands highlight periods of increased IRD content associated with the sea surface coolings. We have numbered these detrital events *a–k* in analogy with the numbering of Bond and Lotti (1995).

the results from Na 87-22, low values of MS, correlated with cold stadial events, are not systematically associated with reduced values of δ^{13}C *C. wuellerstorfi*, for example the event prior to H2 at 23 ka. There is a lack of resolution in the interval between H4 and H5, when the MS record can be well correlated with the DO events.

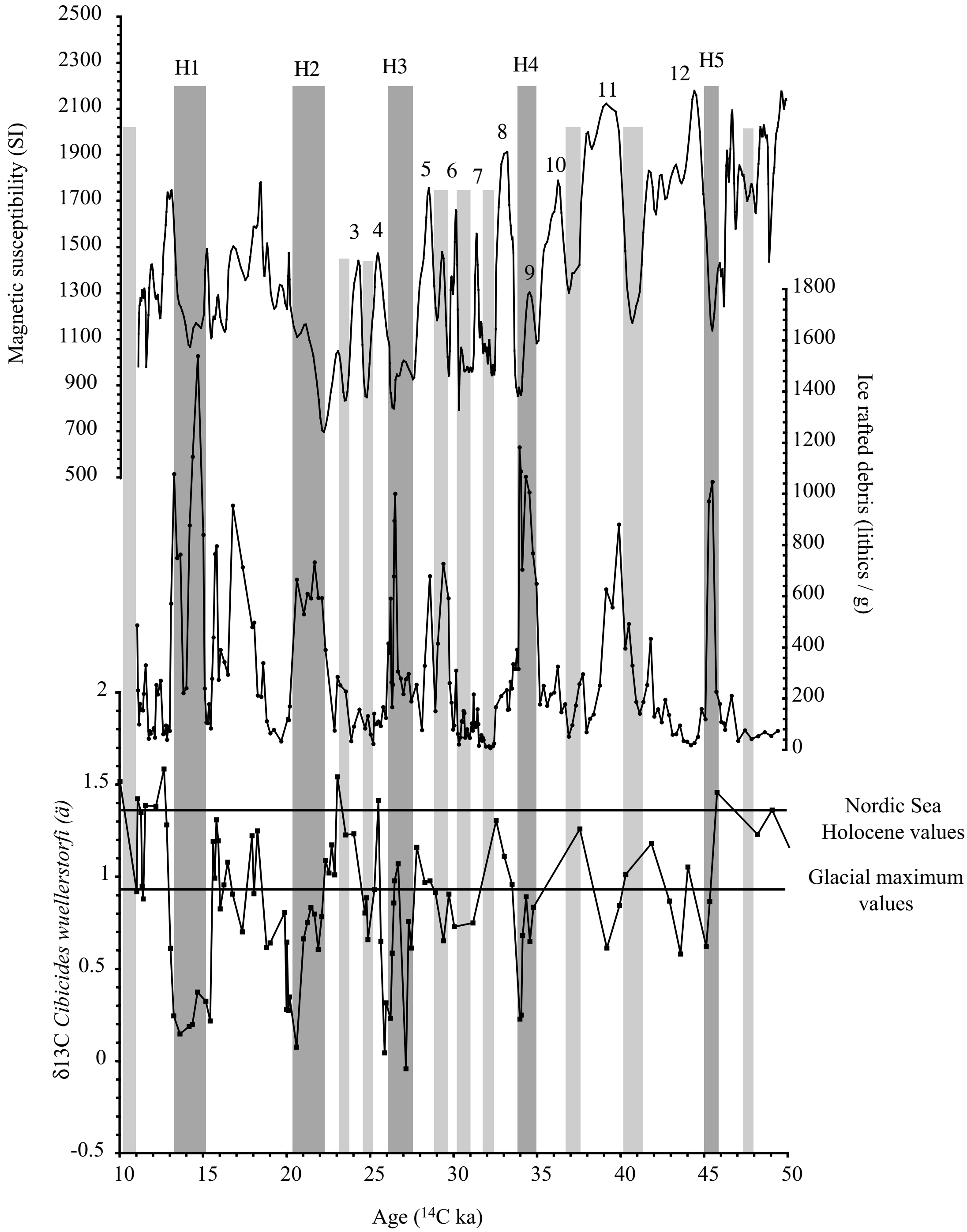

Fig. 4. Results of core SU 90-24 from the Irminger Basin. MS (top), concentration of ice-rafted debris expressed in number of lithic grains per gram of dry sediments (middle), $\delta^{13}C$ *C. wuellerstorfi* (bottom). The initial age model based on a total of 32 ^{14}C dates between 10 and 45 ^{14}C ka is presented in Elliot et al. (1998). As in caption 3, the age of the Heinrich layers (except for H3, see Fig. 6) were constrained to fit the average age as determined from seven sediment cores from the North Atlantic Ocean (see Table 1). The dark bands highlight the periods corresponding to the average age and duration of the Heinrich layers as defined in Table 1. The light gray bands highlight the periods of minimum MS correlated to the cold stadial events recorded in the Greenland ice core. We have numbered the high MS events 3–12, which can be correlated to interstadial events 3–12 of the Greenland ice core record.

However, MS event 11, correlated to a warm interstadial event 11, is marked by a decrease of $\delta^{13}C$ *C. wuellerstorfi* down to +0.6‰. Finally, the Younger–Dryas event is characterized by a strong cooling of around 4°C and an increase in IRD deposits around the Rockall Plateau (Fig. 3). We can observe a coeval decrease of $\delta^{13}C$ *C. wuellerstorfi* in both cores. Values of $\delta^{13}C$ drop to +0.6‰ in core Na 87-22 and to +0.9‰ in core SU 90-24. However, similarly to the other detrital events, the perturbation of deep-water formation associated with this event remains moderate in comparison with those associated with the HE.

4. Discussion

4.1. Deep-water circulation changes associated with the HEs

The marked depletion of $\delta^{13}C$ *C. wuellerstorfi* suggest that HE appear to be associated with a strong northward migration of southern source deep waters to the Rockall Plateau and up to 62°N in the Irminger Basin. This result is in agreement with previous studies stipulating that HEs are associated with drastic reduction or even a total shut down of deep-water convection in the North Atlantic (Oppo and Lehman, 1995; Vidal et al., 1997). However, the $\delta^{13}C$ *C. wuellerstorfi* values observed in the Irminger Basin (+0.1 to +0.2‰) are lower than those observed in the Rockall Plateau, except during H1 (Figs. 3 and 4). The sediment cores compared were both obtained at ~2000 m water depth and thus suggest that there could be a supplementary source of nutrient rich (low $\delta^{13}C$) deep waters in the Irminger Basin during the HE.

A more intense northward migration of southern source waters in the western basin of the North Atlantic Ocean, associated with a stronger glacial lower deep water (see Fig. 1) could explain the lower values observed in the Irminger Basin, but would contradict previous observations. The changes of $\delta^{13}C$ *C. wuellerstorfi* associated with Heinrich event 4 (Vidal et al., 1998) showed that the northward migration of deep southern source waters was more intense in the eastern than in the western North Atlantic Basin, as it is today. Howeever, the density of sediment cores studied was higher in the eastern basin. Also, Vidal et al. (1998) showed that the $\delta^{13}C$ *C. wuellerstorfi* values observed during the HE in sediment cores beneath 4000 m are around 0‰ to −0.2‰ in the western basin at the base of the Newfoundland margin (see also (Labeyrie et al., 1999), whereas they are around 0 to +0.35‰ in the east. These two observations suggest that a more intense northward migration of southern source waters in the Western Basin could have occurred during these events.

Another explanation could be an influx of depleted deep waters from the Nordic Seas via the Denmark Strait overflow into the Irminger Basin during the HE. Low benthic $\delta^{18}O$ anomalies have been observed in the Nordic regions and in the North Atlantic Ocean during the HE (Vidal et al., 1998; van Kreveld et al., 2000) and off the coast of Norway (Dokken and Jansen, 1999). Both studies point to the existence of deep water formed by brine rejection in the Nordic Seas during these episodes. These water masses appear to have migrated into the sub-polar North Atlantic where benthic foraminifera have also recorded small $\delta^{18}O$ incursions (Vidal et al., 1998). Similar benthic $\delta^{18}O$ incursions have been observed in a sediment core off the Portuguese margin but were interpreted as reflecting changes in sea level during the HE (Shackleton et al., 2000), making it unclear whether these low benthic $\delta^{18}O$ values, observed around the North Atlantic Ocean during the HE intervals, are related to deep waters formed by brine rejection in the Norwegian Sea, ice volume, deep-water temperature variations or a combination of these factors. The benthic $\delta^{18}O$ record of core SU 90-24 (Fig. 2) shows some low $\delta^{18}O$ events which could correspond to the inflow of deep waters produced by brine rejection in the Nordic seas and be a source of low $\delta^{13}C$ in the Irminger Basin. However, the low $\delta^{18}O$ events lag the lowest values of benthic $\delta^{13}C$ around H3 and H4 making it unlikely that these deep-water masses were a supplementary source of low $\delta^{13}C$ to the Irminger Basin during the HEs. This lag between the perturbation of NADW and the initiation of deep waters formed by brine rejection has been previously observed (Sarnthein et al., 2000; van Kreveld et al., 2000), and has been interpreted as being responsible for the initiation of the sudden warming post-HE.

We should however recognize that the observed difference in the amplitude of the decreased $\delta^{13}C$ events could be due to different temporal resolution of the records. The flux of planktonic foraminifera and the abundance of benthic foraminifera are higher during the ice rafting episodes in the Irminger Basin (Elliot et al., 1998). This is opposite to observations made within the main IRD belt, between 40 and 55°N where sedimentation rates are reduced. It is also important to note that the changes in benthic $\delta^{13}C$ during the HE do not appear to be the result of enhanced decay of organic matter which could potentially lower the $\delta^{13}C$ waters at the sediment interface. Each layer rich in IRD is associated with similar increases in faunal abundance (Elliot et al., 1998), but not similar decreases in benthic $\delta^{13}C$. Core SU 90-24 could thus be recording the full amplitude of the $\delta^{13}C$ incursions due to enhanced deposition rates during these events. This could explain why previous reconstructions from mid-latitude sediment cores did not encounter such low values of $\delta^{13}C$ *C. wuellerstorfi* at intermediate depth. We postulate that southern source depleted waters did migrate into the Irminger Basin during the HE. The observed differences in the amplitude of the reduced $\delta^{13}C$ could be related to an enhanced temporal resolution during the HE in core SU 90-24 or to a more intense northward migration of southern source depleted waters into the western Basin of the North Atlantic Ocean.

4.1.1. Heinrich event 4: the precursor events?

The $\delta^{13}C$ *C. wuellerstorfi* evolution around the H4 interval shows a marked decrease to lower values at both locations (Fig. 5). The decrease in $\delta^{13}C$ *C. wuellerstorfi* in core SU 90-24 occurs in a two step manner during H4. The first deposits of IRD during H4 are associated with low values of $\delta^{13}C$, around +0.7‰ whereas the lowest values, around +0.2‰, are observed towards the end of the IRD layer (Fig. 5). Similarly, the lowest values of $\delta^{13}C$ during H2, H3 and to a lesser degree during H5 occur during the later phase of the IRD deposits (Fig. 5). The initial deposits of IRD occur during MS event 9 correlated to interstadial 9, whereas the lowest values of $\delta^{13}C$ occur during the minimum MS, prior to interstadial 8 (Fig. 5). It is also at this time that small quantities of detrital carbonate originating from the Laurentide ice sheet have been reported in the Irminger Basin (Bond et al., 1999), see Fig. 5. Thus the lowest values $\delta^{13}C$ and the northernmost migration of southern source waters appear to occur during the later phase of the IRD layer in the Irminger Basin. Conversely, in core Na 87-22, $\delta^{13}C$ values during the H4 interval, drop rapidly in phase with the increased IRD deposits to +0.4‰, and remain low during the entire event.

We are thus able to define a sequence of events surrounding this HE, which recalls the "precursor event" hypothesis proposed by Bond and Lotti (1995) and recently documented by Grousset et al. (2001). The discovery of increases in volcanic glass and hematite stained grains at the base of the detrital carbonate layers led to the concept that the timing of the major collapses from the Laurentide ice sheet may have been climatically controlled. The initial deposits of IRD of core SU 90-24 could thus slightly precede the timing of the Laurentide ice sheet collapses and correspond to the precursor event observed by these authors. In such a case small adjustments would have to be made to the age model surrounding the HE. Owing to inaccurate age constrains within an IRD layer, particularly at around 35 ^{14}C ka, it is not possible to speculate on the relative timing of the different sources of IRD deposits. Two scenarios can however be proposed.

If the detrital carbonate fragments in both cores mark the timing of the Laurentide ice sheet collapses into the Labrador Sea. The icebergs bearing these rock fragments could have been transported into the Irminger Basin via the glacial Irminger Current, small adjustments would have to be made to the age models see Fig. 5 (Model 1). In this scenario the northward migration of deep southern source waters would affect the entire North Atlantic Basin in the east and west. However, the icebergs bearing the detrital carbonate fragments could also originate from the Arctic (Bischof and Darby, 1997). These rock fragments would have been transported to the location of core SU 90-24 via the

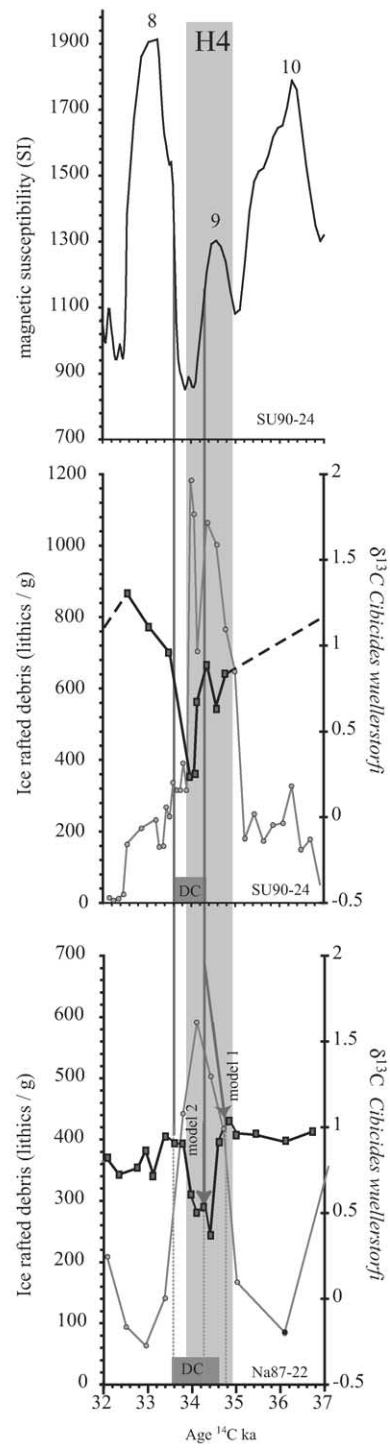

Fig. 5. Zoom on the H4 interval from 32 to 37 ^{14}C ka. MS of core SU 90-24 (top). IRD (circles—left axis), $\delta^{13}C$ *C. wuellerstorfi* (squares—right axis) of core SU 90-24 (middle). Concentration of ice-rafted debris (circles—left axis) $\delta^{13}C$ *C. wuellerstorfi* (squares—right axis) of core NA 87-22 (bottom). The two solid lines highlight the position of the minimum MS correlated to the cold stadial event prior to interstadial 8. The small dark gray boxes (DC) highlight the position of the detrital carbonate layers in core SU 90-24 see (Bond et al., 1999) and core Na 87-22. The two solid lines in the bottom panel (core Na 87-22) represent the two potential age models discussed in the manuscript. The light gray boxes highlight the average age (in ^{14}C ka) and duration of H4 as in Figs. 3 and 4.

glacial East Greenland Current. Given the degree of cooling associated to the HE and the evidence given by a sediment core on the Reykjanes Ridge (van Kreveld et al., 2000), it is unlikely that the warmer Irminger Current reached the location of core SU 90-24 favoring this second hypothesis. The detrital carbonate fragments in core SU 90-24 would then mark the surges from the northern part of the Laurentide ice sheet into the Arctic Sea. In such a scenario the deposit of detrital carbonate fragments may not be synchronous and the northward migration of the southern source deep waters could have operated progressively throughout H4, affecting first the deep waters around the Rockall Plateau, then the deep waters in the Irminger Basin (Model 2, Fig. 5).

4.1.2. Heinrich event 3: The noncharacteristic Heinrich event

H3 is commonly referred to as the "noncharacteristic" HE. When compared to the other HEs, H3 is characterized by a European sources of IRD and origin of icebergs (Gwiazda et al., 1996; Hemming et al., 1998), has different MS signature in the sub-polar North Atlantic (Grousset et al., 1993). H3 has no precursor (Bond and Lotti, 1995) and finally does not appear clearly synchronous (Elliot et al., 1998). Over the H3 interval, the δ^{13}C *C. wuellerstorfi* records show a "W" feature at both locations (Fig. 6). During the initial phase of H3, values of δ^{13}C drop to around +0.6‰ in the Rockall Plateau and peak down to much more depleted values to −0.2‰ in the Irminger Basin (this value is however determined by a single analysis). Then, the δ^{13}C shows an increase to +1.1‰ at both locations and decreases again to +0.5‰ in the Rockall Plateau and to +0.2‰ in the Irminger Basin.

The initial drop in δ^{13}C *C. wuellerstorfi* (~27.4 ^{14}C ka) rapidly follows a peak of IRD observed mainly in core Na 87-22. The second phase of reduced δ^{13}C (~26.1 ^{14}C ka) rapidly follows coeval increases of the abundance of IRD at both locations (Fig. 6). MS values also remain low throughout this interval (Fig. 6) and this episode has been correlated to the abnormally long stadial prior to interstadial 4. This specific interval was subdivided into two phases, interstadial 4a and b based on SST estimates (van Kreveld et al., 2000) and a similar pattern can be observed in the SST record of core Na 87-22 (Fig. 3). These results suggest the existence of a more complex perturbation of deep-water circulation during H3. This particular event seems associated with a two-phased incursion of southern source waters into the North Atlantic which could explain the longer duration of this cold stadial event and the origin of the "noncharacteristic" signature of H3.

In conclusion, results from core SU 90-24, which has enhanced temporal resolution during the HEs shows, that changes in benthic δ^{13}C and modification in deep-water formation lag the initial deposit of IRD during the

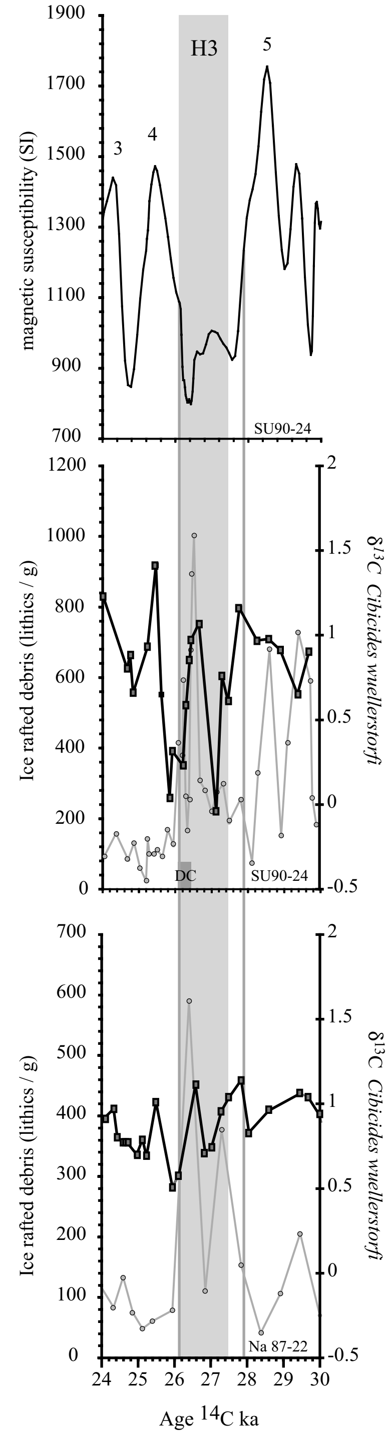

Fig. 6. Zoom of the H3 interval from 24 to 30 ^{14}C ka (see caption as in Fig. 5). The two solid lines highlight the position of the minimum MS correlated to the cold stadial event prior to interstadial 4. The small dark gray box (DC) highlights the position of the detrital carbonate layers in core SU 90-24, see (Bond et al., 1999). The light gray boxes highlight the average age and duration of H3 as in Figs. 3 and 4. The age of H3 in core Na 87-22 was constrained to fit the average age of this HE as determined in Table 1. Only the age of the top of H3 has been adjusted in core SU 90-24 as the age of the base could not be adjusted within the errors of ^{14}C dates to fit the average age as determined in Table 1.

HEs. Although there is still uncertainty around the precise timing of the Laurentide versus the Nordic and Arctic ice sheet instabilities, the modifications of the deep-water formation appear to be a result rather than a trigger of these periods of iceberg discharges which contradicts previous observations (Zahn et al., 1997; Curry et al., 1999). Furthermore at 62°N, the evolution of the $\delta^{13}C$ post-Heinrich event shows a rapid return to well-ventilated deep waters (Figs. 4–6). This observation suggests that deep-water formation resumed rapidly in the Nordic region at the end of the HE.

4.2. Deep-water circulation changes during the inter-Heinrich intervals

The records presented here offer the possibility to compare the changes of deep-water circulation during the inter-Heinrich intervals at intermediate depth. Firstly, our results suggest that changes in the velocity of deep boundary currents are not linearly related to changes in the ventilation rates of NADW formation (Fig. 4). Indeed, whereas H3, H4 and H5 are associated with both reduced velocity of boundary currents (low MS values) and reduced ventilation of deep North Atlantic waters (low $\delta^{13}C$ values), this observation cannot be translated to the inter-Heinrich intervals. Secondly, our records show that the decreased SST (Na 87-22) and the reduced MS (SU 90-24) correlated to the cold stadial events intercalated between the HE are not associated with similar reductions of $\delta^{13}C$ *C. wuellerstorfi*. $\delta^{13}C$ values remain high, closer to that observed during the LGM or even Holocene (Figs. 3 and 4). This observation suggests that the changes in ventilation rates and the perturbations of the THC were not as drastic as those associated with the HEs. If changes in ventilation rates do occur during these cold stadials, as has been shown by Keigwin and Boyle (1999), they affect only the deeper water masses below 2000 m in the North Atlantic Ocean. These changes in the rates of NADW formation thus appear subtle when compared to the THC perturbations associated with the HEs.

Rind et al. (in press) have recently shown that there is a linear response of the THC to increasing freshwater perturbations. Although no precise estimates exist, the HEs are most probably associated with much larger input of low saline waters as it is during these episodes that the major collapses of the Laurentite Sheet occur. The difference in volume of freshwater input to the North Atlantic and also the location of this input, the inter-Heinrich events seem to be related mainly to instabilities of the Fennoscandian ice sheet (Elliot et al., in press), could thus explain the different amplitude of THC perturbation. What is most intriguing is that, given these differences, the degree of cooling observed in the Greenland ice core is of similar amplitude during each stadial event.

4.3. Implications

There is evidence which shows that, although the timing of the HEs may have been climatically controlled (Bond and Lotti, 1995), the amplitude of the continental ice sheets collapses was catastrophic in character. The inter-Heinrich events appear to be the expression of the millennial-scale climate oscillator which operates during cold glacial and warm interglacial periods (Bond et al., 1997; de Menocal et al., 2000). Other paleo-climatic data have illustrated the global effect of these climatic rebounds. These can be observed at high and low latitudes e.g. (Sachs and Lehman, 1999; Peterson et al., 2000) and in the South Atlantic, Indian and Pacific Ocean (Behl and Kenneth, 1996; Schulz et al., 1998; Ninnemann et al., 1999; Kiefer et al., 2001). It is intriguing to note that most of these records show that, periods corresponding the HE and those corresponding to the other cold stadials are often characterized by similar signatures. Records from Antarctica do however show that only the periods surrounding the HEs are associated with significant warming of atmospheric temperatures (Blunier et al., 1998) which could indicate that the THC "sea saw" is only fully operational during these episodes and would agree with our results.

The global imprint of these climatic events, characterized by coeval perturbations of sea surface hydrology and modifications of atmospheric circulation, is not explained by current model simulations of THC perturbations. Model results, which simulate a complete shut down of the THC ($\sim$the Heinrich event situation), show that the impacts of this change in ocean circulation would primarily affect the North Atlantic sector and to a lesser degree the Antarctic Ocean (Manabe and Stouffer, 1995; Manabe and Stouffer, 1997; Schiller et al., 1997). Conversely, a reduced THC ($\sim$the inter-Heinrich situation) does not explain the degree of cooling above Greenland, as illustrated by recent model simulations (Rind et al., in press). Our results thus question the role of changes in THC as the unique explanation for the amplitude and the widespread signature of the millennial scale climatic oscillations during the last glacial. Or, perhaps there is an important amplifying and feedback mechanism not taken into account in current model simulations to extend the effects of a reduced deep-water formation in the North Atlantic to the other regions of the globe.

5. Conclusion

Two high-resolution records of benthic foraminifera $\delta^{13}C$ records from mid-water depth ($\sim$2000 m) in the North Atlantic give new insight on the role of changes in ventilation rates of North Atlantic Deep Waters and perturbations of the thermohaline circulation associated with the Dansgaard–Oeschger (DO) temperature oscillations. Our results are based on the correlation of SST changes in the sub-polar North Atlantic and MS records from the Irminger Basin with the DO temperature oscillations. Drastic reductions of benthic $\delta^{13}C$ *C. wuellerstorfi* are associated with the HEs interpreted as reflecting a northward migration of southern source depleted waters to 62°N in the Irminger Basin. The large depletions of $\delta^{13}C$ observed in the Irminger Basin could be explained either by a more intense northward migration of Southern source deep waters into the western basin of the North Atlantic Ocean or to higher depositional rates at this location during the HE at this location. The variations of benthic $\delta^{13}C$ during the inter-Heinrich events, associated to the cold stadials observed in Greenland ice core, are of lesser amplitude. The changes in the THC during the inter-Heinrich stadial events were thus of different magnitude although the degree of cooling above Greenland were of similar amplitude. Given our present understanding of the effect of a shutdown of the thermohaline circulation, as simulated by model experiments, there is need for an important amplifying mechanism to generate both the widespread imprint and amplitude of the climatic changes associated with these events.

Acknowledgements

This research was supported by the NOAA Post-doctoral Program in Climate and Global Change, administered by the University Corporation for Atmospheric Research. Basic support from the CEA-CNRS to the LSCE, programs Geosciences marines, PNEDC (INSU), ATP, EU environmental program, IFRTP and the IMAGES program. The isotopic data were obtained in Gif sur Yvette; B. Lecoat, D. Dole and J. Tessier were in charge of the measurements. H. Leclaire's support and advice for micropaleonthology were of great help. We also acknowledge M. Arnold, in charge of the accelerator mass spectrometry, where the ^{14}C dates have been obtained. The role of IFREMER and Genavir in operating the coring cruises Paleocinat 1 and French R/V Le suroit is acknowledged. I wish to thank Tom Marchitto for helpful comments on the manuscript. Two anonymous reviewers and the Editor also greatly helped improve the quality of the manuscript. This is an LDEO contribution number 6266.

References

Aagaard, K., Carmack, E.C., 1989. The role of sea ice and other fresh water in the Arctic circulation. Journal of Geophysical Research 94 (C10), 14485–14498.

Bauch, H.A., Erlenkeuser, H., Spielhagen, R.F., Struck, U., Matthiessen, J., Thiede, J., Heinemeir, J., 2001. A multiproxy reconstruction of the evolution of deep and surface waters in the subarctic seas over the last 30,000 yr. Quaternary Science Reviews 2001, 659–678.

Behl, R.J., Kenneth, J.P., 1996. Brief interstadial events in the Santa Barbara basin, NE Pacific, during the past 60 kyr. Nature 379, 243–246.

Bianchi, G.G., McCave, I.N., 1999. Holocene periodicity in North Atlantic and deep-ocean flow south of Iceland. Nature 397, 515–517.

Bischof, J.F., Darby, D.A., 1997. Mid- to late Pleistocene ice drift in the Western Arctic Ocean: evidence for a different circulation in the past. Science 277, 74–78.

Blunier, T., Chappellaz, J., Schawnder, J., Dällenbach, A., Stauffer, B., Stocker, T.F., Raynaud, D., Jouzel, J., Clausen, H.B., Hammer, C.U., Jonhsen, S.J., 1998. Asynchrony of Antarctic and Greenland climate change during the last glacial period. Nature 394, 739–743.

Bond, G.C., Lotti, R., 1995. Iceberg discharges into the North Atlantic on millenial time scales during the last glaciation. Science 267, 1005–1010.

Bond, G., Broecker, W., Johnsen, S., McManus, J., Labeyrie, L., Jouzel, J., Bonani, G., 1993. Correlations between climate records from North Atlantic sediments and Greenland ice. Nature 365, 143–147.

Bond, G., Showers, W., Cheseby, M., Lotti, L., Almasi, P., deMonecal, P., Priore, P., Cullen, H., Hajdas, I., Bonani, G., 1997. A pervasive millennial-scale cycle in North Atlantic Holocene and glacial climates. Science 278, 1257–1266.

Bond, G.C., Showers, W., Elliot, M., Evans, M., Lotti, R., Johnson, S., 1999. Tracing the North Atlantic's ~1.5 kyr climate cycle to 80,000 Years BP-relationship to Heinrich events, Dansgaard/Oeschger cycles, and the little ice age. In: Clark, P.U., Webb, R., Keigwin, L. (Eds.), Mechanisms of Millennial-Scale Global Climate Change, AGU Monographs. American Geophysical Union, Washington, DC, Snowbird, Utah, pp. 35–58.

Boyle, E.A., 2000. Is the ocean thermohaline circulation linked to abrupt stadial/interstadial transitions? Quaternary Science Reviews 19, 255–272.

Boyle, E.A., Keigwin, L.D., 1982. Deep circulation of the North Atlantic over the last 200,000 years: geochemical evidence. Science 218, 784–787.

Broecker, W., 1997. Thermohaline circulation, the Achilles Heel of our climate system: will man-made CO_2 upset the current balance? Science 278, 1582–1588.

Broecker, W.S., Denton, G.H., 1989. The role of ocean-atmosphere reorganizations in glacial cycles. Geochimica et Cosmochimica Acta 53, 2465–2501.

Broecker, W.S., Sutherland, S., Peng, T.H., 1999. A possible 20th-century slowdown of Southern Ocean deep water formation. Science 286, 1132–1135.

Cane, M., Clement, A., 1999. A role of the tropical pacific coupled ocean-atmosphere system on Milankovitch and millennial time-scales. PartII: global impacts, in mechanisms of millennial-scale global climate change. In: Clark, P.U., Webb, R., Keigwin, L. (Eds.), AGU Monographs. American Geophysical Union, Washington, DC, Snowbird, Utah.

Chapman, M.R., Shackleton, N.J., 2000. Evidence of 550-year and 1000-year cyclicities in North Atlantic circulation patterns during the Holocene. The Holocene 10 (3), 287–291.

Charles, C.D., Lynch-Stieglitz, J., Ninnemann, U.S., Fairbanks, R.G., 1996. Climate connections between the hemisphere revealed by deep sea sediment core/ice core correlations. Earth and Planetary Science Letters 142 (1-2), 19–27.

Clement, A., Cane, M., 1999. A role of the tropical pacific coupled ocean-atmosphere system on Milankovitch and Millennial time-scales. Part I: a modeling study of tropical pacific variability. In:

Clark, P.U., Webb, R., Keigwin, L. (Eds.), Mechanisms of Millennial-Scale Global Climate Change, AGU Monographs. American Geophysical Union, Washington, DC, Snowbird, Utah.

Coplen, T.B., 1988. Normalization of oxygen and hydrogen isotope data. Chemical Geology (Isotope Geoscience Section) 72, 293–297.

Cortijo, E., Labeyrie, L., Vidal, L., Vautravers, M., Chapman, M., Duplessy, J.-C., Elliot, M., Arnold, M., Turon, J.-L., Auffret, G., 1997. Changes in sea surface hydrology associated with Heinrich event 4 in the North Atlantic Ocean between 40° and 60°N. Earth and Planetary Science Letters 146, 29–45.

Curry, W.B., Marchitto, T.M., McManus, J.F., Oppo, D.W., Laakamp, K.L., 1999. Millennial-scale changes in ventilation of the thermocline, intermediate, and deep waters of the glacial North Atlantic. In: Clark, P.U., Webb, R., Keigwin, L. (Eds.), Mechanisms of Millennial-scale Global Climate Change, AGU Monographs. American Geophysical Union, Washington, DC, Snowbird, Utah, pp. 59–76.

Dansgaard, W., Johnsen, S.J., Clausen, H.B., Dahl-Jensen, D., Gundestrup, N.S., Hammer, C.U., Hvidberg, C.S., Steffensen, J.P., Sveinbjörnsdottir, A.E., Jouzel, J., Bond, G., 1993. Evidence for general instability of past climate from a 250 kyr ice-core record. Nature 364, 218–220.

de Menocal, P.B., Ortiz, J., Guilderson, T., Sarnthein, M., 2000. Coherent high- and low-latitude climate variability during the Holocene warm period. Science 288, 2198–2202.

Dickson, R.R., Brown, J., 1994. The production of North Atlantic deep water: sources, rates and pathways. Journal of Geophysical Research 99, 12319–12341.

Dokken, T.M., Jansen, E., 1999. Rapid changes in the mode of deep and intermediate water formation linked with atmospheric and ice sheet variations during the last glacial. Nature 401, 458–461.

Duplessy, J.-C., Shackleton, N.J., Matthews, R.K., Prell, W., Ruddiman, W.F., Caralp, M., Hendy, C.H., 1984. ^{13}C record benthic foraminifera in the last interglacial ocean: implications for the carbon cycle and the global deep water circulation. Quaternary Research 21, 225–243.

Duplessy, J.C., Shackleton, N.J., Fairbanks, R.G., Labeyrie, L.D., Oppo, D., Kallel, N., 1988. Deepwater source variations during the last climatic cycle and their impact on the global deepwater circulation. Paleoceanography 3 (3), 343–360.

Elliot, M., 1999. Variabilité millénaire du climat et de l'hydrologie de l'océan Atlantique nord lors de la dernière période glaciaire (60,000–10,000 ans). University of Orsay, Orsay.

Elliot, M., Labeyrie, L., Bond, G., Cortijo, E., Turon, J.-L., Tisnerat, N., Duplessy, J.-C., 1998. Millennial-scale iceberg discharges in the Irminger Basin during the last glacial period: relationship with the Heinrich events and environmental settings. Paleoceanography 13 (5), 433–446.

Elliot, M., Labeyrie, L., Dokken, T., Manthé, S., 2001. Coherent patterns of ice rafted debris deposits in the nordic regions during the last glacial (10–60 ka). Earth and Planetary Science Letters 194, 151–163.

Ganopolski, A., Rahmstorf, S., 2001. Rapid changes of glacial climate simulated in a coupled climate model. Nature 409, 153–158.

Grousset, F., Labeyrie, L., Sinko, J., Cremer, M., Bond, G., Duprat, J., Cortijo, E., Huon, S., 1993. Patterns of ice rafted detritus in the glacial North Atlantic (40–55°N). Paleoceanography 8, 175–192.

Grousset, F.E., Cortijo, E., Huon, S., Hervé, L., Richter, T., Burdloff, D., Duprat, J., Weber, O., 2001. Zooming in on Heinrich layers. Paleoceanography 16 (3), 240–259.

Gwiazda, R.H., Hemming, S.R., Broecker, W.S., 1996. Provenance of icebergs during Heinrich event 3 and the contrast to their sources during other Heinrich episodes. Paleoceanography 11 (4), 371–378.

Heinrich, H., 1988. Origin and consequences of cyclic ice-rafting in the northeast Atlantic Ocean during the past 130,000 years. Quaternary Research 29, 142–152.

Hemming, S.R., Broecker, W.S., Sharp, W.D., Bond, G.C., Gwiazda, R.H., McManus, J.F., Klas, M., Hajdas, I., 1998. Provenance of Heinrich layers in core V28-82, northeastern Atlantic: $^{40}Ar/^{39}Ar$ ages of ice rafted hornblende, Pb isotopes in feldspar grains, and Nd–Sr–Pb isotopes in fine sediment fraction. Earth and Planetary Science Letters 164, 317–333.

Hughen, K.A., Southon, J.R., Lehman, S.J., Overpeck, J.T., 2000. Synchronous radiocarbon and climate shifts during the last deglaciation. Science 290 (5498), 1951–1954.

Hut, G., 1987. Stable isotope reference samples for geochemical and hydrological investigations. Consultant Group Meeting IAEA, Vienna, 16–18 September 1985, Report to the Director General, International Atomic Energy Agency, Vienna.

Kanfoush, S.L., Hodell, D.A., Charles, C.D., Guilderson, T.P., Mortyn, P.G., 2000. US Ninneman, millennial-scale instability of the Antarctic ice sheet during the last glacial. Science 288, 1815–1818.

Keigwin, L.D., Boyle, E.A., 1999. Surface and deep ocean variability in the northern Sargasso Sea during marine isotope stage 3. Paleoceanography 14, 164–170.

Kiefer, T., Sarnthein, M., Erlenkeuser, H., Grootes, P.M., Roberts, A.P., 2001. North Pacific response to millennial-scale changes in ocean circulation over the past 60 kyr. Paleoceanography 16, 179–189.

Kissel, C., Laj, C., Labeyrie, L., Dokken, T., Voelker, A., Blamart, D., 1999. Rapid climatic variations during marine isotopic stage 3: magnetic analysis of sediments from Nordic Seas and North Atlantic. Earth and Planetory Science Letters 171 (3), 489–502.

Labeyrie, L., Heclaire, H., Waelbroek, C., Cortijo, E., Duplessy, J.-C., Vidal, L., Elliot, M., Lecoat, B., Auffret, G., 1999. Insolation forcing and millennial scale variability of the north west Atlantic ocean: surface versus deep water changes. In: Clark, P.U., Webb, R., Keigwin, L. (Eds.), Chapman Conference on Millennial Scale Climate Variability, AGU Monographs. American Geophysical Union, Washington, DC, Snowbird, Utah, pp. 77–98.

Lund, D.C., Mix, A.C., 1998. Millennial-scale deep water oscillations: reflections of the North Atlantic in the deep pacific from 10 to 60 ka. Paleoceanography 13, 10–19.

MacAyeal, D.R., 1993. A low-order model of the Heinrich event cycle. Paleoceanography 8, 767–773.

Manabe, S., Stouffer, R., 1995. Simulation of abrupt climate-change induced by fresh-water input to the North Atlantic Ocean. Nature 378 (6553), 165–167.

Manabe, S., Stouffer, R., 1997. Coupled ocean-atmosphere model response to freshwater input: comparison to Younger Dryas event. Paleoceanography 12 (2), 321–336.

Moros, M., Endler, R., Lackschewitz, K.S., Wallrabe-Adams, H.-J., Mienert, J., Lemke, W., 1997. Physical properties of Reykjanes Ridge sediments and their linkage to high-resolution Greenland ice sheet project 2 ice core data. Paleoceanography 12, 687–695.

Ninnemann, U.S, Charles, C.D., Hodel, D.A., 1999. Origin of global millennial scale climate events: constrains from the southern ocean deep sea sedimentary record. In: Clark, P.U., Webb, R., Keigwin, L. (Eds.), Chapman Conference on Millennial Scale Climate Variability, AGU Monographs. American Geophysical Union, Washington, DC, Snowbird, Utah, pp. 99–112.

Oppo, D.W., Lehman, S.J., 1993. Mid-depth circulation of the subpolar North Atlantic during the last glacial maximum. Science 259, 1148–1152.

Oppo, D.W., Lehman, S.J., 1995. Suborbital timescale variability of North Atlantic deep water during the past 200,000 years. Paleaoceanography 10 (5), 900–910.

Peterson, L.C., Haug, G.H., Hughen, K.A., Röhl, U., 2000. Rapid changes in the hydrologic cycle of the Tropical Atlantic during the last glacial. Science 290, 1047–1051.

Pflaumann, U., Duprat, J., Pujol, C., Labeyrie, L., 1996. SIMMAX, a modern analog technique to deduce Atlantic sea surface temperatures from Planktonic foraminifera in deep sea sediments. Paleoceanography 11 (1), 15–35.

Prell, W.L., 1985. The stability of low-latitude sea-surface temperatures: an evaluation of the CLIMAP reconstruction with emphasis on the positive SST anomalies. US Department of Energy, Washington, DC.

Rahmstorf, S., 1994. Rapid climate transitions in a coupled ocean-atmosphere model. Nature 372, 82–85.

Rasmussen, T.L., Thomsen, E., Tjeerd, C.E., Labeyrie, L., van Veering, T.C.E., 1996a. Circulation changes in the Faeroe-Shetland Channel correlating with the cold events during the last glacial period (58–10 ka). Geology 24, 937–940.

Rasmussen, T.L., Thomsen, E., Tjeerd, C.E., van Veering, T.C.E., Labeyrie, L., 1996b. Rapid changes in the surface and deep water conditions at the Faeroe Margin during the last 58,000 years. Paleoceanography 11, 757–771.

Renssen, H., van Geel, B., van der Plicht, J., Magny, M., 2000. Reduced solar activity as a trigger for the start of the Younger Dryas? Quaternary International 68-71, 373–383.

Rind, D., De Menocal, P., Russel, G.L., Seth, S., Collins, D., Schmidt, G.A., Teller, J., 2001. Effects of glacial meltwater in the GISS coupled atmosphere-ocean model: Part I: North Atlantic deep water response. Journal of Geophysical Research 106 (27), 355–366.

Sachs, J.P., Lehman, S.J., 1999. Subtropical North Atlantic temperatures 60,000–30,000 years ago. Science 272, 756–759.

Sarnthein, M., Winn, K., Jung, S.J.A., Duplessy, J.C., Labeyrie, L.D., Erlenkeuser, H., Ganssen, G., 1994. Changes in East Atlantic deep water circulation over the last 30,000 years: eight time slice reconstructions. Paleoceanography 9, 209–267.

Sarnthein, M., et al., 2000. Fundamental modes and abrupt changes in the North Atlantic circulation and climate over the last 60 kyr—concepts, reconstruction and numerical modeling. In: Schafer, P., Ritzau, W., Schluter, M., Thiede, J. (Eds.), The Northern North Atlantic: A Changing Environment. Springer, Berlin, pp. 365–410.

Schiller, A., Mikolajewicz, U., Voss, R., 1997. The stability of the North Atlantic thermohaline circulation in a coupled ocean-atmosphere general circulation model. Climate Dynamics 13, 325–347.

Schulz, H., von Rad, U., Erlenkeuser, H., 1998. Correlation between Arabian Sea and Greenland climate oscillations of the past 110,000 years. Nature 393, 54–57.

Shackleton, N., Hall, M., Vincent, E., 2000. Phase relationship between millennial-scale events 64,000–24,000 years ago. Paleoceanography 15, 565–569.

Stocker, T.F., Wright, D.G., 1991. Rapid transitions of the ocean's deep circulation induced by changes in surface water fluxes. Nature 351, 729–732.

Waelbroeck, C., Duplessy, J.C., Michel, E., Labeyrie, L., Paillard, D., Duprat, J., 2001. The timing of the last deglaciation in the North Atlantic climate records. Nature 412, 724–726.

van Kreveld, S., Sarnthein, M., Erlenkeuser, H., Grootes, P., Jung, S., Nadeau, M.J., Pflaumann, U., Voelker, A., 2000. Potential links between surging ice sheets, circulation changes, and the Dansgaard–Oeschger cycles in the Irminger Sea, 60–18 kyr. Paleoceanography 15 (4), 425–442.

Veum, T., Jansen, E., Arnold, M., Beyer, I., Duplessy, J.C., 1992. Water mass exchange between the North Atlantic and the Norwegian Sea during the past 28,000 years. Nature, 356, 783–785.

Vidal, L., Labeyrie, L., Cortijo, E., Arnold, M., Duplessy, J.C., Michel, E., Becqué, S., van Weering, T.C.E., 1997. Evidence for changes in the North Atlantic deep water linked to meltwater surges during the Heinrich events. Earth and Planetary Science Letters 146, 13–27.

Vidal, L., Labeyrie, L., van Weering, T.C.E., 1998. Benthic $\delta^{18}O$ records in the North Atlantic over the last glacial period (60–10 kyr): evidence for brine formation. Paleoceanography 13, 245–251.

Zahn, R., Schönfeld, J., Kudrass, H.-R., Park, M.-H., Erlenkeuser, H., Grootes, P., 1997. Thermohaline instability in the north Atlantic during meltwater events: stable isotope and ice rafted detritus records from core SO75-26KL, Portuguese margin. Paleoceanography 12 (5), 696–710.