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Link between the North and South Atlantic during the Heinrich events of the last glacial period

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Abstract High resolution benthic oxygen isotope records combined with radiocarbon datings, from cores retrieved in the North, Equatorial, and South Atlantic are used to establish a reliable chronostratigraphy for the last 60 ky. This common temporal framework enables us to study the timing of the sub-Milankovitch climate variability in the entire surface Atlantic during this period, as reflected in planktonic oxygen isotope records. Variations in sea surface temperatures in the Equatorial and South Atlantic reveal two warm periods during the mid-stage 3 which are correlated to the warming observed in the North Atlantic after Heinrich events (HL) 5 and 4. However, the records show that the warming started about 1500 y earlier in the South Atlantic. A zonally averaged ocean circulation model simulates a similar north-south thermal antiphasing between the latitudes of our coring sites, when perturbated by a freshwater flux anomaly. We infer that the observed phase relationship between the northern and the southern Atlantic is related to periods of reduced NADW production in the North Atlantic, such as during HL5 and HL4.

1 Introduction

Millennial-scale climate fluctuations have been documented from different paleo-climate archives: marine sediments (Bond et al. 1993), ice cores (Dansgaard et al. 1993), and continental deposits (Lowell et al. 1995). The fact that these "short term" climate variations are recorded globally raises questions regarding their relative timing, and the mechanisms which are responsible for their interhemispheric connections. Changes in the North Atlantic Deep Water (NADW) production played an important role for climate variations on orbital time scales (Broecker and Denton 1989; Imbrie et al. 1992, 1993). For the millennial time scale, however, it is yet not clear whether the world wide transfer of the climate signal is mainly through the ocean (via changes in the thermohaline circulation) or mainly through the atmosphere (via alterations in the transport of heat and moisture) (Mikolajewicz et al. 1997; Fanning and Weaver 1997). To investigate the potential mechanisms for propagating millennial-scale climate events throughout the globe, we need to compare rapid climate changes recorded in different areas. For that purpose, an accurate common chronology is required.

The construction of a common time scale is the major problem in correlating paleoclimatic records from different environments. Successful attempts were obtained for interpolar ice core correlations based on $\delta^{18}O(O_2)$ and CH_4 contents of the entrapped air (Sowers and Bender 1995; Blunier et al. 1997, 1998). To study the oceanic climate on short time scales, one possibility is to correlate the marine records to very high resolution ice core records which clearly register a series of abrupt climate changes. However, this procedure entails several uncertainties. In ice cores, annual layer counting provides an absolute dating (calendar scale) for the climate records (Alley et al. 1997) whereas for marine records radiocarbon dating is used for the last 45 ka BP (Arnold et al. 1987). The calibration of radiocarbon ages on the calendar scale is well constrained for the last 20 ka BP, but inaccuracies remain for the 45 ka – 20 ka period (Bard et al. 1993; Kitagawa and van der Plicht 1998). In addition, the temporal resolution between ice core and marine sediment records often differs by an order of magnitude and may affect the correlation between the climate proxies.

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Here we make an attempt to overcome these problems by correlating high-resolution benthic oxygen records from both hemispheres, in order to establish the required common chronology for an interhemispheric comparison of climate changes in the surface Atlantic Ocean.

This study is based on deep-sea sediment cores from the North, the Equatorial, and the South Atlantic. All the cores presented have been dated by AMS ^{14}C. We first establish a detailed chronostratigraphy for the marine oxygen isotope stage (MIS) 3 constrained with numerous ^{14}C datings for each individual site. Then, a detailed comparison of the high-resolution benthic $\delta^{18}O$ records enables us to correlate the cores. Based on this chronology, the question was whether millennial-scale fluctuations in the surface water hydrology over the last 60 ky, inferred from planktonic $\delta^{18}O$ records, were in phase between the hemispheres. Special emphasis is placed on the sea surface temperature (SST) reconstructions in the North Atlantic associated with Heinrich events 5 and 4 and $\delta^{18}O$-derived SST estimates in the Equatorial and South Atlantic for the same period. The phase relationships between the northern and southern surface ocean are discussed in relation to the paleoceanographic conditions that prevailed during the Heinrich events of the last glacial period.

2 General setting and method

Four high-resolution cores were chosen from the Atlantic Ocean (Fig. 1). The sites NA87-22 (55 °N, 14 °W, 216 m) and SU90-08 (43 °N, 30 °W, 308 m) are located in the North Atlantic within water masses corresponding to the Upper North Atlantic Deep Water (UNADW) and the Lower North Atlantic Deep Water (LNADW) respectively. In both cores, the IRD layers corresponding to the Heinrich events have been identified and ^{14}C dated by AMS. For these sites the stable isotopes records and the detailed chronostratigraphy have been presented in a previous study (Vidal et al. 1997). Core GeoB1515 (4 °14.3N, 43 °4W) was retrieved on the Ceara Rise flank at 3129 m water depth within the lower component of NADW. The sampling spacing was 1.5 cm for the first 3 m of the core and 5 cm for deeper levels. Oxygen and carbon isotopes have been measured on planktonic *Globigerinoïdes sacculifer* (*G. sacculifer*) and benthic *Cibicidoides wuellerstorfi* (*C. wuellerstorfi*) foraminifera shells. All the isotope analyses were made at the Geosciences Department at Bremen University, using a Finnigan MAT 252 mass spectrometer. The measurements are reported in permil versus PBD after calibration with NBS 19. The mean external reproducibility is $\pm 0.05‰$ for both $\delta^{18}O$ and $\delta^{13}C$. AMS ^{14}C datings were made on monospecific *G. sacculifer* samples (Table 1). Core GeoB1711 (23 °18.9S, 12 °22.6E, 196 m) was taken from the Cape Basin, just to the south of the Walvis Ridge. This site is under the influence of the southward flow of NADW (Warren and Speer 1991). Planktonic and benthic isotope records have been obtained from samples taken at 5 cm intervals throughout the core (Little et al. 1997a, b). AMS ^{14}C datings for core GeoB1711 span the interval from 35 ka BP to the late Holocene (Table 2).

2.1 Construction of a common time scale

For core GeoB1515, the age model was constructed from a polynomial fit through the ^{14}C age control points and the isotopic event

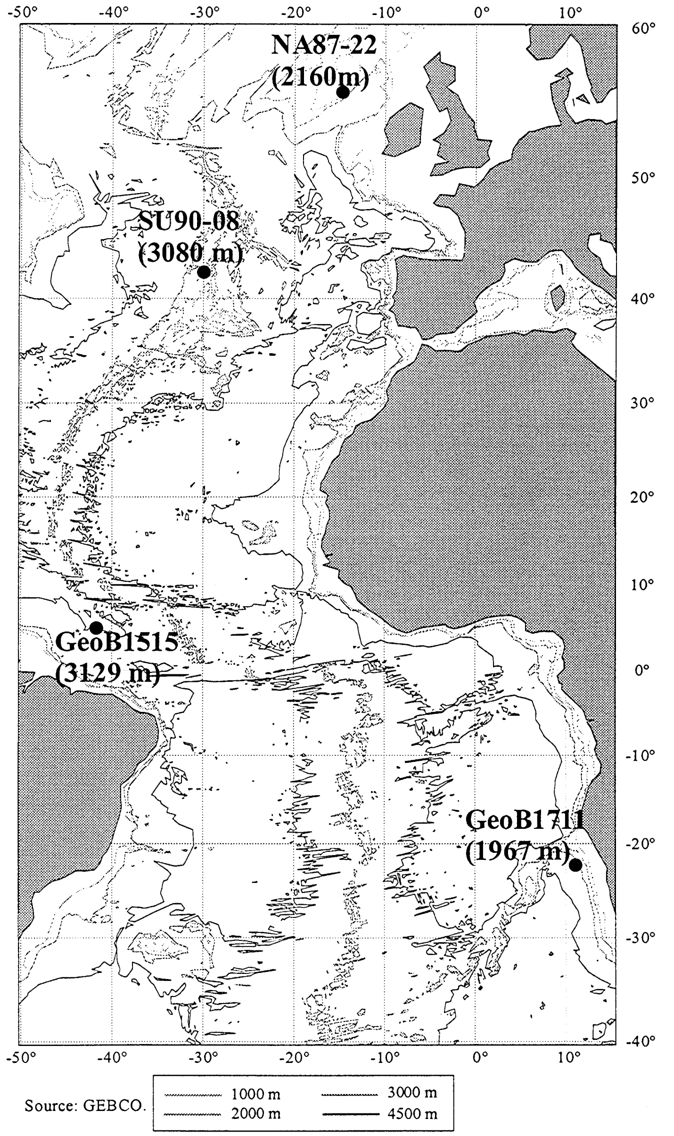

Fig. 1 Location of the cores studied

corresponding to the transition between MIS 4 and 3 according to Pisias et al. (1984) (Fig. 2a). For core GeoB1711, the ^{14}C ages were linearly interpolated (Fig. 2a). The time scales are presented in radiocarbon ages. A value of 400 y was subtracted from the ^{14}C ages in order to account for the reservoir effect (Bard et al. 1988). For deeper levels (older than 60 ka), the age model was obtained by correlating the benthic $\delta^{18}O$ records to the SPECMAP stack (Imbrie et al. 1984). Sedimentation rates at the four sites are relatively high and continuous: 4 cm/ka for GeoB1515, 8 cm/ka for GeoB1711 (Figs. 2a and 3), 5.2 cm/ka for SU90-08, and 10 cm/ka for NA87-22 (Vidal et al. 1997; Fig. 3).

The benthic $\delta^{18}O$ record for each core exhibits typical glacial-interglacial changes mainly documenting the changes in continental ice volume (Labeyrie et al. 1987) (Fig. 3). In addition, low $\delta^{18}O$ peaks lasting 2–3 ky are apparent between 50 and 25 ky BP in the MIS 3 (Fig. 3). The benthic record from the shallow North Atlantic core NA87-22 (2160 m) reveals a higher variability than the other cores. Variations in the benthic $\delta^{18}O$ records and in $\delta^{18}O$ gradients between different cores in the North Atlantic have shown that isotopic

Table 1 Radiocarbon ages[a] (AMS) for samples of the planktonic foraminifera *G. sacculifer* from core GeoB1515

	Laboratory code	Depth (cm)	Composite depth (cm)[b]	^{14}C age (ka)	Age-0.4 (ka)	Error age (+)	Error age (−)
GKG	KIA3266	1	1	2.360	1.96	0.07	− 0.07
GKG	KIA3271	1	1	2.460	2.06	0.05	− 0.05
GKG	KIA3272	1	1	2.250	1.85	0.05	− 0.05
GKG	KIA1877	13	13	6.22	5.82	0.04	− 0.04
GKG	KIA1842	13	13	6.14	5.74	0.03	− 0.03
GKG	KIA1841	25	25	9.65	9.25	0.05	− 0.05
GKG	KIA1876	34	34	11.74	11.34	0.07	− 0.06
Core	KIA1840	16.5	25.5	9.55	9.15	0.05	− 0.05
Core	KIA1839	24.5	33.5	11.8	11.4	0.07	− 0.07
Core	KIA3264	30	39	13.28	12.88	0.07	− 0.07
Core	KIA1839	38	47	15.43	15.03	0.08	− 0.07
Core	KIA1837	48	57	16.92	16.52	0.11	− 0.11
Core	KIA1313	70	79	20.81	20.41	0.29	− 0.28
Core	KIA1314	94.5	83.5	26.59	26.19	0.59	− 0.55
Core	KIA1315	111	120	29.63	29.23	0.89	− 0.8
Core	KIA1316	129.5	138.5	32.77	32.37	1.33	− 1.14
Core	KIA1317	129.5	138.5	32.67	32.27	1.32	− 1.13
Core	KIA3263	144.5	153.5	34.94	34.54	0.47	− 0.44
Core	KIA4114	164.5	173.5	45.57	45.17	1.93	− 1.55
Core	KIA1319	166.5	175.5	42.39	41.99	5.6	− 3.27
Core	KIA3262	166.5	175.5	38.05	37.65	0.64	− 0.59

[a]Graphite samples were prepared at the University of Bremen and analyzed at the Leibniz-Labor AMS facility, Christian Albrechts University, Kiel, Germany, following the procedures described by Nadeau et al. (1997)

[b]The composite depth corresponds to corrected depth (+9 cm) of the core samples relative to the boxcore samples (GKG)

Table 2 Radiocarbon ages[a] (AMS) for samples of planktonic foraminifera from core GeoB1711

Samples	Laboratory code	Depth (cm)	^{14}C age (ka)	Age-0.4 (ka)	Error age (+)	Error age (−)	
Core	KIA2590	28	4.52	4.12	0.05	− 0.05	*G. inflata*
Core	KIA2589	43	7.07	6.67	0.04	− 0.04	*G. inflata*
Core	KIA2588	103	12.15	11.75	0.05	− 0.05	*G. inflata*
Core	KIA557[b]	108	12.68	12.28	0.07	− 0.07	*G. inflata*
Core	KIA4109	138	14.06	13.66	0.06	− 0.06	*G. bulloides*
Core	KIA556[b]	183	17.14	16.74	0.13	− 0.12	*G. inflata*
Core	KIA555[b]	243	18.74	18.34	0.13	− 0.13	*G. inflata*
Core	KIA4110	288	29.13	28.73	0.24	− 0.24	*G. bulloides*
Core	KIA554[b]	333	35.37	34.97	1.09	− 0.96	*G. inflata*

[a]Graphite samples were prepared at the University of Bremen and analyzed at the Leibniz-Labor AMS facility, Christian Albrechts University, Kiel, Germany, following the procedures described by Nadeau et al. (1997)

[b]^{14}C ages published in Little et al. (1997b)

The date at 243 cm was not considered to calculate the age model (Fig. 2a)

events observed in shallow cores result from local processes (Vidal et al. 1998). In this study we only consider the isotopic events recorded in both the shallow and the deep North Atlantic cores (NA87-22 and SU90-08) showing a lowering in the δ^{18}O values by $\sim 0.3‰$ (light δ^{18}O peaks 1 and 2 are marked LP1 and LP2 in Fig. 3). These δ^{18}O lowerings were attributed mainly to variations of ocean isotopic composition due to changing ice volume (Vidal et al. 1998). The benthic oxygen records from the Equatorial and South Atlantic cores GeoB-1515 and -1711 also show marked instances of low δ^{18}O values peaks over the time interval 50–25 ky BP (Fig. 3).

The primary aim of this study is the construction of a reliable common chronology for the last glacial period (60–10 ka BP) which is essential for the study of climate response on millennial time scales. We try to improve the time correlation of the high-resolution benthic δ^{18}O records between the North Atlantic cores (NA87-22 and SU90-08), and the Equatorial (GeoB1515) and South Atlantic cores (GeoB1711). Our time correlation is based on benthic δ^{18}O stratigraphy as it provides a sequence of events that can be identified on a global scale. Indeed δ^{18}O changes resulting from ocean mean isotopic shifts due to changing ice volume on the continent are uniform on a global basis, within the mixing time of the ocean, at most 1 or 2 ky depending on the core location (Shackleton and Opdyke 1973). Isotopic events that can be consistently recognized basin wide allow us to correlate the different records in order to construct a common depth/time scale (Prell et al. 1986). The depths of isotopic events known to be identified on a global scale (E1, E2,

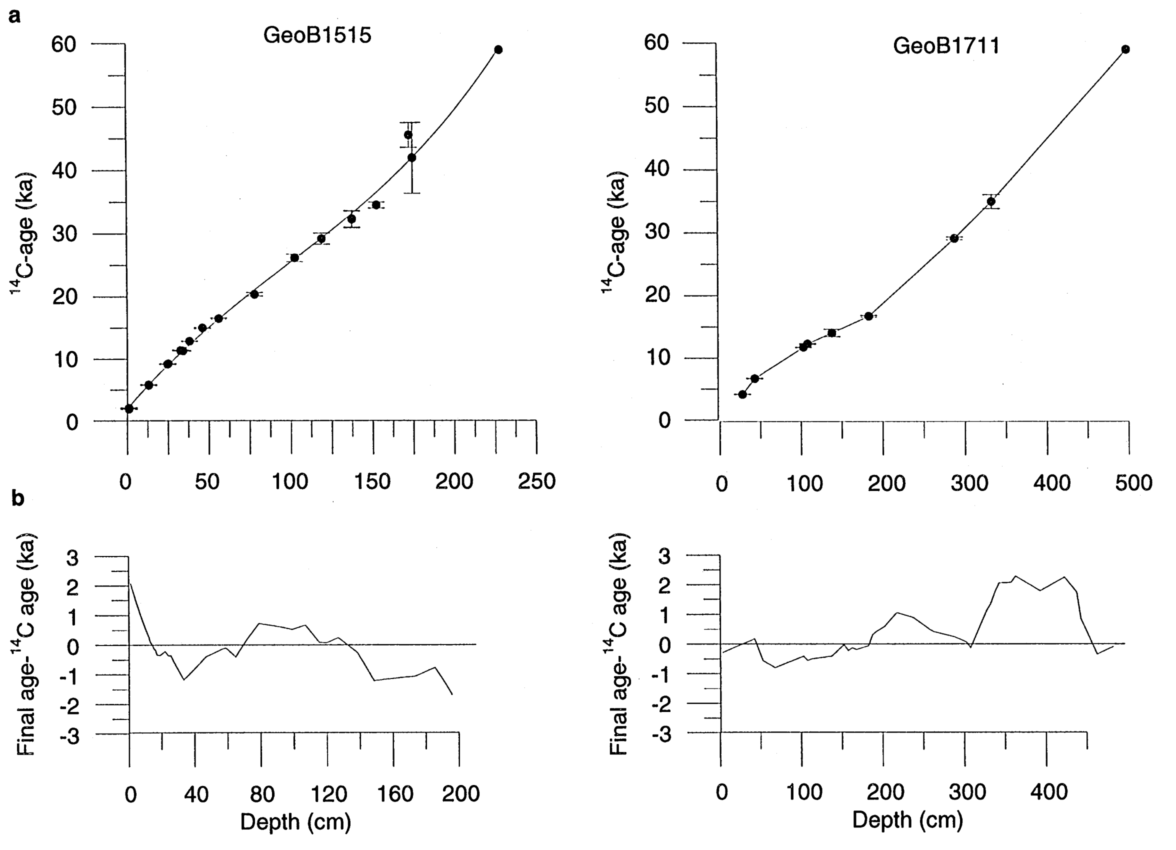

Fig. 2 a Age model for cores GeoB1515 and GeoB1711 based on reservoir-corrected radiocarbon dates and the boundary between MIS 4 and 3 at 59 ka BP. For core GeoB1515, a polynomial fit of 3rd order was used. **b** Differences between ages in the final age model and ages based on the model in **a** for cores GeoB-1515 and -1711. A two-sided Mann test (Kendall and Gibbons 1990) indicates no significant trend at the 1% probability level of the age difference with depth in cores GeoB-1515 and -1711

E3, E4 in Fig. 3) in cores GeoB-1515 and -1711 were plotted versus the depth of the same events identified in the reference section core SU90-08, thereby defining a correlation line between the cores (Fig. 4a and Table 3). Other selected isotopic events should fall on this correlation line if they are contemporaneous within the mixing time of the ocean (Prell et al. 1986). The high variability in the benthic records from the North, Equatorial and South Atlantic, characterized by light δ^{18}O peaks during MIS 3, raises the question whether some of these peaks reflect the same isotopic events. If the δ^{18}O changes are temperature-induced or influenced by local processes, they are unlikely to be uniform on a global scale. If they result from changes of ocean isotopic composition, due to changing ice volume, one expects basin-wide uniformity. This is tested by comparing the stratigraphic position of the short-lasting isotopic events in the different cores. The most pronounced light δ^{18}O anomalies (defined as LP1 and LP2) in cores GeoB-1515 and -1711 fit to the correlation line defined from the global scale isotope events (Fig. 4a). This strongly suggests that these events were simultaneous on the Atlantic basin scale. Although the magnitude of the light δ^{18}O peaks in the benthic records is in the range of the within-sample variability (0.2–0.5‰) when individual (1 or 2) foraminiferal tests are measured (Curry 1996), these isotopic events are believed to reflect a common climatic signal. Double and triple δ^{18}O measurements from samples with low δ^{18}O values usually agree to within 0.15‰. Moreover, the radiocarbon ages obtained for these levels are in good agreement from one core to another (Table 3) which all have continuous sedimentation rates for the studied interval (Figs. 2a and 3).

The benthic δ^{18}O records from cores GeoB-1515 and -1711 were then correlated to the reference benthic δ^{18}O record from core SU90-08, using the sequence of isotopic events recognized in all the records (Fig. 4b). Uncertainties in this chronological correlation depend on the mixing time of the Atlantic ocean and on the temporal resolution of the records. In the Atlantic, the present age of deep waters is 200 y (Broecker et al. 1991) while ~700 y (±200 y) has been estimated for the Last Glacial Maximum (Broecker et al. 1990). Sample resolution is close to 500 y in all the cores studied. We estimate therefore an average uncertainty of ±1000 y for our chronology. The differences between ages derived from benthic isotopes correlation and the measured ^{14}C ages are within the 2 sigma error of the ^{14}C datings for cores GeoB-1515 and -1711 (Fig. 2b). No trend is observed in the age differences which would give indications for older ages for the events of the more southern cores. Although the proposed chronology does not provide a better accuracy than a chronology based only on radiocarbon constraints, it is supported by the detailed comparison of the high-resolution benthic records for the different cores.

The occurrence of light benthic δ^{18}O peaks (LP1 and LP2) in cores SU90-08, GeoB-1515 and -1711, at ~43 ka BP and ~33 ka BP, strongly suggests relatively rapid changes in the δ^{18}O of seawater during MIS 3 (Fig. 4b). Relative to the Heinrich events, these peaks of low δ^{18}O occur during and shortly after the iceberg discharges in the North Atlantic associated with HL5 and HL4 as previously observed in δ^{18}O records from North Atlantic cores (Vidal et al. 1998; Fig 4b). Sea level rises of 2 to 5 m have been estimated for iceberg discharges during the Heinrich events

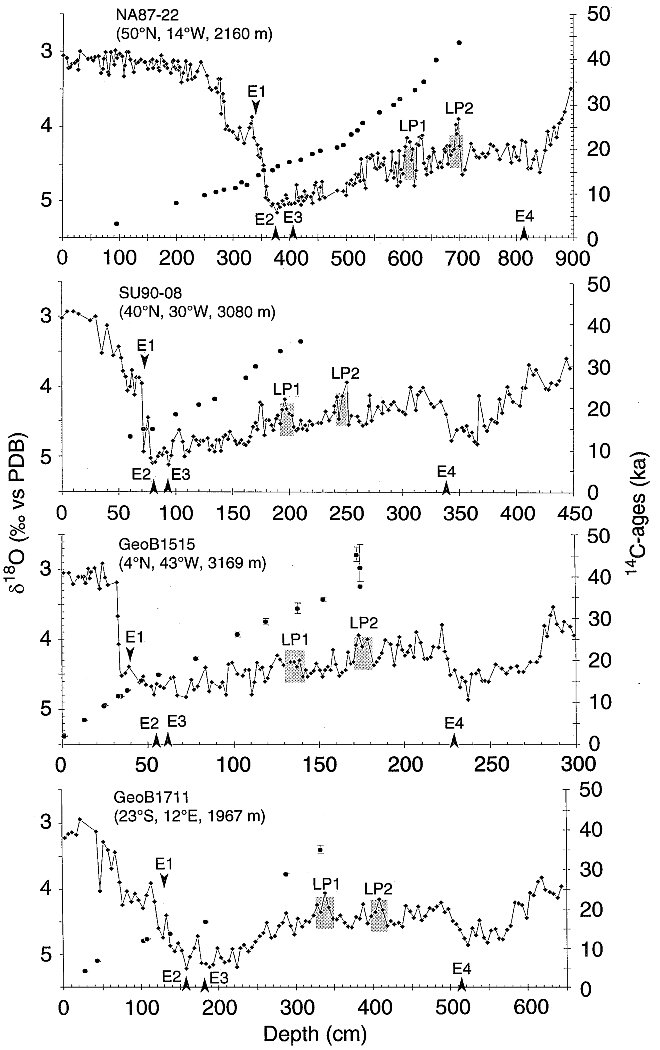

Fig. 3 Benthic $\delta^{18}O$ records versus depth measured on *C. wuellerstorfi* in cores NA87-22, SU90-08, GeoB1515 and GeoB1711. The ^{14}C dated levels are indicated by *black dots*. *Black arrows* in each panel indicate the position of isotopic events known to have a global signature: Deglaciation (first low $\delta^{18}O$ values peak in Termination I) (*E1*), Maximum in $\delta^{18}O$ values (*E2*), Last Glacial Maximum (*E3*), Transition between MIS 4 and 3 ($\delta^{18}O$ values at the midpoint between glacial $\delta^{18}O$ values at the end of stage 4 and the first light $\delta^{18}O$ peak in MIS 3) (*E4*). *LP1* and *LP2* define the light $\delta^{18}O$ peak 1 and light $\delta^{18}O$ peak 2. The height of the *shaded area* corresponds to 0.4‰ on the $\delta^{18}O$ scale

(McAyeal 1993; Labeyrie et al. 1995). This would cause a shift in the mean ocean $\delta^{18}O$ of 0.02‰ to 0.05‰ which is less that 25% of the changes recorded in the benthic $\delta^{18}O$ light peaks (LP1 and LP2). Assuming a negligible effect of deep water temperature changes, we suggest additional melting of continental ice sheets associated with or slightly following the Heinrich events. Still the mechanisms responsible for ice volume changes on short time scales are not fully understood. Interestingly, $\delta^{18}O$ anomalies with similar amplitudes at about 43 ka and 33 ka BP, were documented in a benthic oxygen record from the deep Pacific (Lund and Mix 1998). This supports the hypothesis that the light isotopic events observed in the Atlantic cores represent a global signal.

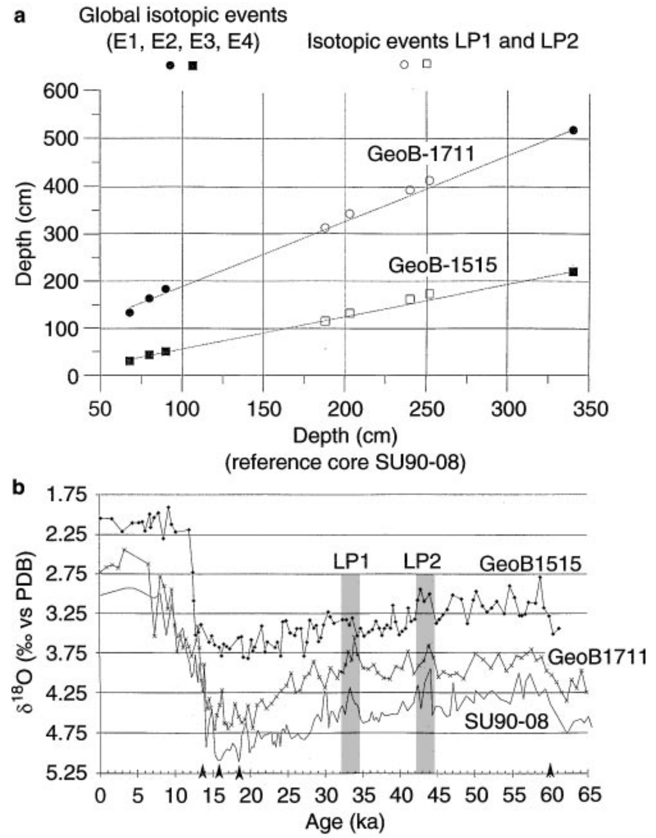

Fig. 4 a Correlation between depths in cores GeoB1515 and GeoB1711 and depths in the reference section core SU90-08. *Black symbols* indicate the depth of isotopic events corresponding to changes in ocean mean $\delta^{18}O$ in each core (see Table 3). *Open symbols* correspond to the depth of the isotopic events LP1 and LP2 in each core (see Table 3 and Fig. 3). **b** Benthic $\delta^{18}O$ records versus age for cores SU90-08, GeoB1515 and GeoB1711. The age model is based on the correlation to the reference core SU90-08. The $\delta^{18}O$ records of GeoB1711 and GeoB1515 are shifted by -0.5‰ and -1‰ respectively for better clarity

3 Surface water changes in the South Atlantic versus North Atlantic

The planktonic $\delta^{18}O$ record for the surface dwelling species *G. sacculifer* in core GeoB1515 (4 °N) shows several short term excursions over the last glacial period, particularly at ~43 ka and ~33 ka BP (Fig. 5). For these specific events, the $\delta^{18}O$ variations in the planktonic record are about 0.5–0.6‰, while reaching 0.2–0.3‰ in the benthic record (Fig. 5). At the Ceara Rise, changes in surface water salinity would result only in minor $\delta^{18}O$ changes due to low fractionation of ^{18}O from ^{16}O (Craig and Gordon 1965). It seems then unlikely that observed planktonic $\delta^{18}O$ changes are solely a result of salinity changes in the western Atlantic. Sea surface temperature reconstruction in tropical areas is problematic as transfer functions are

Table 3 Ages of isotopic events corresponding to global signal in the benthic $\delta^{18}O$ records from cores GeoB1515, GeoB1711, SU90-08, and NA78-22

	GeoB1515			GeoB1711			SU90-08		NA87-22	
	Depth (cm)	^{14}C age (ka)	*Age (ka) after correlation*	Depth (cm)	^{14}C age (ka)	*Age (ka) after correlation*	Depth (cm)	^{14}C age (ka)	Depth (cm)	^{14}C age (ka)
E1	39	12.6	*13.45*	133	13.43	*13.85*	68	13.48	338	13.64
E2	52	15.59	*15.86*	163	15.02	*15.5*	82	16.04	390	15.96
E3	59	17.13	*17.23*	183	16.74	*16.8*	90	17.62	410	16.97
LP1	124	30.36	*30*	313	32.19	*32*	188	31.97	600	31.74
	141	33.97	*34.6*	343	36.42	*34.4*	203	34.5	620	34.01
LP2	171	41.11	*42.15*	393	43.7	*41.93*	240	42.02	688	42.66
	182	44.07	*44.92*	408	46.62	*44.5*	252	44.34	700	44.29
E4	229	59.6	*61*	518	61.27	*61.56*	340	61	815	61

E1, E2, E3, E4 define the isotopic events corresponding respectively to the deglaciation, maximum $\delta^{18}O$ values, LGM, transition between MIS 4 and 3 (see Fig. 3). For GeoB1515 and GeoB1711, ages are obtained both from ^{14}C datings (second column) and correlation to the North Atlantic core SU90-08 (third column). The ^{14}C dates for GeoB-1515 and GeoB1711 are given in Tables 1 and 2

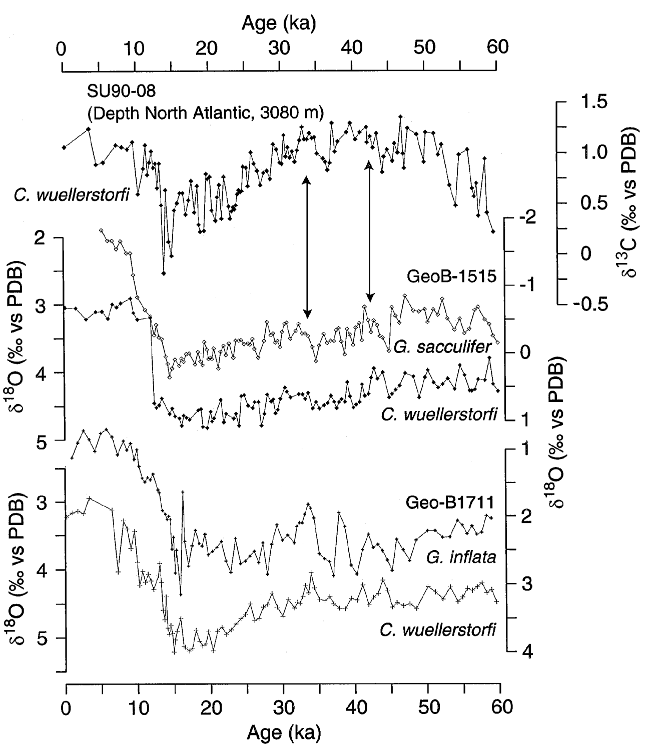

Fig. 5 Planktonic $\delta^{18}O$ records measured on *G. sacculifer* in core GeoB1515 and on *G. inflata* in core GeoB1711. The benthic $\delta^{18}O$ records for these two cores and the $\delta^{13}C$ benthic record for core SU90-08 are also shown

not sensitive enough in the high temperature range (Prell 1985; Curry and Oppo 1997; Wolff et al. 1998). Isotopic-derived temperature estimates lead to 2–3 °C changes when the entire variation of the planktonic $\delta^{18}O$ in core GeoB1515 is attributed to temperature. These values are similar to those estimated from a nearby planktonic record (Curry and Oppo 1997). However, changes in the mean ocean $\delta^{18}O$ could partly contribute to the planktonic $\delta^{18}O$ variability; if changes of ~ 0.2–$0.3‰$ are considered, as inferred from the benthic oxygen isotope data, the estimated SST changes are smaller, reaching 1–2 °C. The planktonic $\delta^{18}O$ variations in core GeoB1515 parallel the benthic $\delta^{13}C$ variations from the North Atlantic core SU90-08 between 55-25 ky BP (Fig. 5). These results are consistent with a previous study showing that warm/cold oscillations in tropical surface waters were synchronous with changes in the production of NADW in the North Atlantic during the last glacial period (Curry and Oppo 1997).

Similar to core GeoB1515, the planktonic $\delta^{18}O$ record for *Globorotalia inflata* (*G. inflata*) from the Cape Basin core GeoB1711 exhibits short-term anomalies characterized by low $\delta^{18}O$ values (Fig. 5). This core, located in the Benguela Current System, is influenced by coastal upwelling which makes the interpretation of this planktonic record more delicate. Under modern condition, *G. inflata* is the most common taxon preserved in the surface sediment along the southwest African margin. This species, however, is poorly represented in coastal samples beneath the upwelling centers but dominates the foraminiferal assemblages in environments with reduced upwelling conditions and relatively low nutrient levels (Giraudeau 1993). Its maximum abundance is reached off the divergence that splits the Benguela current into the coastal branch and the oceanic gyre water (Bé and Tolder 1971). Distribution of *G. inflata* parallels distribution of subtropical/tropical species representative of the oceanic regimes (Giraudeau and Rogers 1994). Similar observations were obtained from downcore distribution of planktonic assemblages in GeoB1711 indicating the *G. inflata* represents hydrological conditions typical for the oceanic branch of the Benguela Current (Little et al. 1997a). In this area, waters which upwell along the coast originate from the South Atlantic Central Waters (SACW) from a depth of 100–300 m (Shannon 1985). Close to the coast, SACW reach the surface while off the shelf, at the location of core GeoB1711, the contribution of these waters to the surface mixed layer is attenuated. As *G. inflata* secretes most of its calcite in the surface mixed layer (Fairbanks et al. 1982) within the depth range 0–75 m (Ravelo and Fairbanks 1992), we believe that its $\delta^{18}O$ is indicative from the near surface waters conditions. This is supported from the downcore distribution of the warm water species assemblage in core GeoB1711 which covaries with the $\delta^{18}O$ record for *G. inflata* (Fig. 6). Variations of the $\delta^{18}O$ record are therefore interpreted as reflecting temperature changes of the gyre surface waters. This, however, does not exclude that part of the observed hydrological changes may have also occurred in sub-surface and intermediate layers, the signal being partly transported to the surface. The equivalent amplitude of the planktonic $\delta^{18}O$ variations in the GeoB1711 core to that observed in the Ceara Rise core suggests similar temperature changes in both areas during MIS 3 (Fig. 5).

Using our time scale, the planktonic $\delta^{18}O$ records from the Equatorial and southeast Atlantic are compares with the SST reconstruction for core SU90-08 from the central North Atlantic for the last 60 ky (Fig. 6). During this period, changes in surface waters conditions in the North Atlantic, indicated by episodic coolings and subsequent warmings, were linked to the Heinrich events (Bond et al. 1992; Labeyrie et al. 1995; Cortijo et al. 1997). Warm transients in the South and Equatorial Atlantic are time correlative with warm periods observed in the North Atlantic surface waters, in particular for those following HL5, HL4, and HL1. However, the warming trends in the South Atlantic (GeoB1711) occurred about 1500 y earlier than in the North Atlantic (Fig. 6).

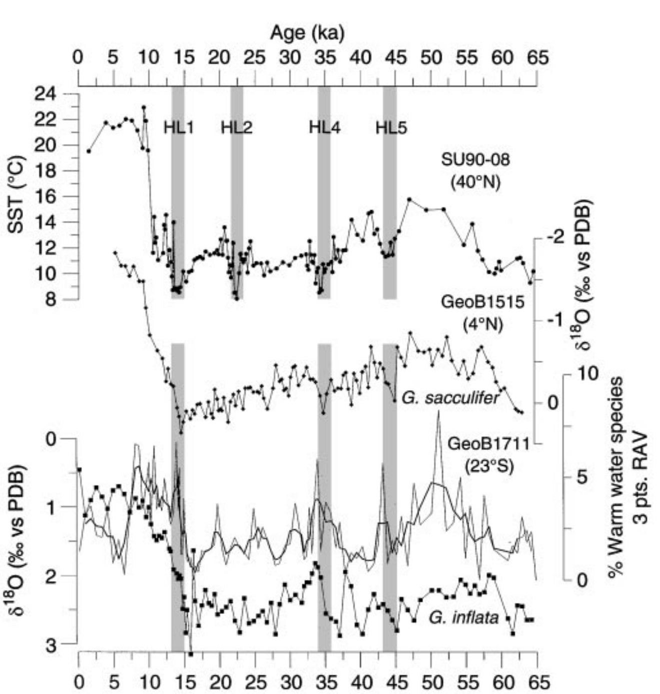

Fig. 6 SST records based on foraminiferal transfer function for core SU90-08 (Cortijo et al. 1997), planktonic $\delta^{18}O$ records for cores GeoB-1515 and -1711, and percents of warm water species in core GeoB1711 (*solid line*) (the *bold line* indicates a 3 points running average). The ages of the Heinrich events are indicated by *shaded areas*

4 Implications

High-resolution planktonic $\delta^{18}O$ records from different cores allow us to recognize sub-Milankovitch surface water temperature variations in the North-, Equatorial and South Atlantic over the last glacial period. Based on a common stratigraphic framework, we found that warm transients in the South Atlantic occurred $\sim$1500 y earlier than in the North Atlantic. Although this value is close to the uncertainty of the proposed chronology, our data set points to an early climate change in the south. Additional adjustments to the age model of core GeoB1711, to obtain a synchronous response between the North and South Atlantic surface ocean, would have led to differences between the measured ^{14}C ages and the ages derived from this correlation, which would have been significantly larger than the error range of the ^{14}C datings. The offset between both hemispheres on sub-Milankovitch time scales is consistent with earlier findings (Charles et al. 1996; Little et al. 1997b; Blunier et al. 1998). However, the mechanisms connecting the northern and the southern oceans on the millennial scale are still a matter of debate (Lowell et al. 1995; Bard et al. 1997; Curry and Oppo 1997; Little et al. 1997b; Broecker 1998; Stocker 1998; Cane 1998). To explain both the time offset and the links between the hemispheres on these time scales, some authors considered the impact of variations in the zonality of low-latitude winds on the northern heat transport (McIntyre and Molfino 1996; Little et al. 1997b). Increased zonality in the trade winds would create optimal conditions (enhanced moisture supply) for the rapid growth of the ice sheets in the northern hemisphere until the ice collapses. This process implies that the subpolar North Atlantic maintained warm SST during the rapid ice-growth phase (Ruddiman and McIntyre 1981) and is inconsistent with the gradual cooling observed in the North Atlantic surface waters during a Bond cycle (Bond et al. 1993). Moreover, the lead of the South Atlantic with respect to the Equatorial Atlantic for HL4 and HL1, does not support the hypothesis of tropical climate variability as a forcing mechanism (Fig. 6).

In the following, possible mechanisms for interhemispheric connections are discussed, with special emphasis on the two warm periods at ~43 ka and ~33 ka BP, which followed HL5 and HL4 respectively (Fig. 6). In the North Atlantic these warm transients are correlated with the prominent interstadials 12 and 8 in the GRIP ice core records (Bond et al. 1993). Massive iceberg discharges in the North Atlantic during the last glacial period led to abrupt changes in the surface water hydrology, reduced NADW flux, and altered deep water circulation patterns (Keigwin and Jones 1994; Oppo and Lehman 1995; Vidal et al. 1997; Zahn et al. 1997). The effects of these discharges were also detected south of the North Atlantic (Charles et al. 1996; Curry and Oppo 1997). After the Heinrich events, the reinitiation of NADW production is accompanied by a warming of North Atlantic surface waters (Wright and Stocker 1993; Paillard and Labeyrie 1994; Oppo and Lehman 1995; Cortijo et al. 1997; Vidal et al. 1997). The early warming observed in the South Atlantic core GeoB1711 (23 °S) is concomitant with the HL5 and HL4 cold periods in the North Atlantic (Fig. 6). At the Equatorial GeoB1515 (4 °N) core location, the SST-proxy does not show a significant offset to the North Atlantic records (Fig. 6).

The thermohaline overturn and NADW production are associated with considerable interhemispheric northward heat transport in the Atlantic. The southward export of NADW is compensated by a return flow of warm near surface water from the South Atlantic (Gordon 1986; Rintoul 1991). Modulation in the thermohaline overturn may therefore have some effects on the cross-equatorial heat transport (Mix et al. 1986), resulting in a net warming in the South when NADW is reduced (Crowley 1992). We argue that reduced NADW production during Heinrich events 5 and 4 could have contributed to the early warming observed in core GeoB1711.

To test this hypothesis we examine the SST changes at the latitude of our cores simulated by a zonally averaged, global ocean circulation model in response to a surface freshwater pulse in the North Atlantic (reference simulation of Marchal et al. 1998). The pulse corresponds to a mean flux of 0.2 Sv which is higher but of the same order of magnitude as the freshwater fluxes estimated for the Heinrich events (Labeyrie et al. 1995). When the pulse is applied, the model simulates a total collapse of the thermohaline circulation in the Atlantic, an extreme response which might not have occurred during these events. However, the associated $\delta^{13}C$ depletion predicted in the deep North Atlantic (Marchal et al. 1998) is qualitatively consistent with the observation of a decrease in benthic foraminiferal $\delta^{13}C$ from sedimentary records of the deep North Atlantic during Heinrich events (Vidal et al. 1997; Zahn et al. 1997). The collapse of the thermohaline circulation in the model

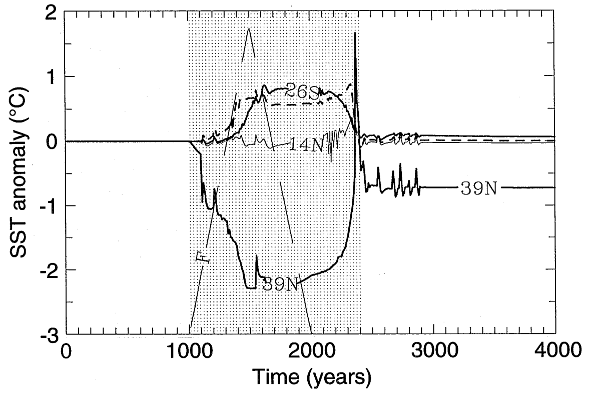

Fig. 7 SST anomaly at 39 °N (*solid line*), at 14 °N (*thin line*), at 0 ° (*short dashed line*), and at 26 °S (*solid line*) in the Atlantic simulated by a zonally averaged circulation model in response to the application of a freshwater input (FI) of $6*10^6$ km^3 over 1000 y (*triangular symmetric discharge*) between 32.5 °N–45 °N in the Atlantic (for details see Marchal et al. 1998). The *FI* (*long dashed line*) is represented by a range of 0 to 0.4 Sv. The *shaded area* denotes the period during which the Atlantic thermohaline circulation (THC) is altered in the simulation (THC drops from 24 Sv at 0 y to a minimum of 3 Sv at 1956 y)

leads to a strong cooling in the North Atlantic (39 °N) and a warming in the South Atlantic (26 °S) (Fig. 7). In the Equatorial Atlantic the modelled-SSTs show a warming at 0 °N and no significant change at 14 °N, suggesting that this area marks the boundary between the warming in the South and the cooling in the North (Fig. 7). One cannot expect that this simplified model simulation reproduces exactly the spatial distribution of our temperature reconstructions. The model reproduces the large-scale features of the north-south thermal antiphasing suggested by the SST reconstructions from the North and South Atlantic cores, but the simulated equatorial warming is inconsistent with the cooling documented in core GeoB1515 (Fig. 6). However, although, the latitude separating regions with cooling and warming in the model may be parameter-dependent, the simulated north-south thermal antiphasing is likely a robust result, consistent with simple physical considerations (Crowley 1992). An asymmetric climate response between Northern and Southern Hemisphere to changes in oceanic overturning is also obtained in three-dimensional, coupled ocean-atmospheric circulation models (Manabe and Stouffer 1988; Schiller et al. 1997; Mikolajewicz et al. 1997). In one of these models, the warming in the South Atlantic surface waters is interpreted as a compensatory effect, reflecting a weakening of the inflow of Southern Ocean subsurface water into the subtropical Atlantic (Schiller et al. 1997). Thus, both our data and model simulations support the hypothesis of a thermal offset between the North and South Atlantic surface waters during Heinrich events 5 and 4.

We suggest that the most likely mechanism for interhemispheric connection on sub-Milankovitch time scales is the oceanic link. Recent synchronisation work of ice isotopic records from Greenland and Antarctica points also to the dominant role of the ocean for coupling the climate of these regions during the last glacial period (Blunier et al. 1998). The observed lag between both hemispheres is attributed to the Southern Hemisphere climate response, in particular in the South Atlantic surface circulation, to abrupt changes in deep water circulation in the North Atlantic during HL5 and HL4. This scenario could also hold for the rapid events during deglaciation and particularly during HL1. The release of icebergs and freshwater during HL1 has been attributed to the internal variability of the Laurentide ice sheet as inferred from terrestrial records from the northern North Atlantic region (McCabe and Clark 1998). For this period our data again indicate an early response of the South Atlantic with respect to the North Atlantic (Fig. 6). Reduced NADW production associated with HL1 would have generated cooling in the Northern Hemisphere and warming in the Southern Hemisphere climate, superimposed on the global deglacial warming trends in both hemispheres. This mechanism would be therefore consistent with the climate antiphasing observed in the ice core records from Greenland and Antarctica over this period (Sowers and Bender 1995; Jouzel et al. 1995; Blunier et al. 1997, 1998).

Other climate records from continental and marine sediments, however, point to interhemispheric synchroneity during abrupt climate changes (Lowell et al. 1995; Bard et al. 1997). The surface ocean response in the Southern Hemisphere may thus not be uniform during such events. In particular, SST reconstructions based on alkenones from a core retrieved in the Indian Ocean (20 °S) show warm transients synchronous with the interstadials observed in a Greenland ice core $\delta^{18}O$ record during the last glacial period (Brad et al. 1997). Decoupled oceanic surface circulation between the South Atlantic and the Indian Ocean during these periods could possibly explain the in-phase evolution of temperature, observed at these two remote areas. On the other hand a recent study based on lake sediments in Madagascar (19 °S) shows an early deglacial warming with respect to the Northern Hemisphere major warming period (Gasse and Van Campo 1998). Clearly it is necessary to increase the number of climate records with reliable time scales in order to better constrain the relative timing of abrupt climate changes in different regions of the globe.

5 Conclusion

A temporal framework, based on high-resolution benthic $\delta^{18}O$ records from North, Equatorial and South Atlantic deep-sea cores has been established for the last 60 ky. Specific isotopic events identified in the benthic $\delta^{18}O$ records have been ^{14}C-dated and used as stratigraphic markers. The light $\delta^{18}O$ events in the benthic records are interpreted as changes in the ocean mean $\delta^{18}O$ occurring at $\sim$43 ka and $\sim$33 ka BP. Based on our chronology, warm periods in the Equatorial and South Atlantic surface waters, are related to the North Atlantic surface water warming following HL5 and HL4. The warming trend in the South Atlantic, however, occurred about 1500 y earlier than in the North Atlantic.

The thermal antiphasing between southern and northern surface Atlantic is attributed to the South Atlantic climate response to abrupt changes in NADW production during the Heinrich events 5 and 4. This suggests that the global ocean circulation connects interhemispheric climate on millennial time scales. This is supported by a zonally averaged ocean model, which simulates a similar antiphasing between the latitudes of our cores in response to a freshwater-induced collapse of the Atlantic thermohaline circulation. NADW is associated with a considerable interhemispheric northward heat transport in the Atlantic. Reducing this transport has therefore two immediate consequences: a cooling in the (high-latitude) Northern Hemisphere but also a warming at least in some parts of the Southern Hemisphere. Further records as well as modelling experiments are required in order to investigate the rapid changes in ocean surface conditions outside the Atlantic region during the last glacial period.

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