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Dansgaard-Oeschger and Heinrich event imprints in Alboran Sea paleotemperatures

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Abstract. Past sea surface temperature (SST) evolution in the Alboran Sea (western Mediterranean) during the last 50,000 years has been inferred from the study of C_{37} alkenones in International Marine Global Change Studies MD952043 core. This record has a time resolution of ~200 years allowing the study of millennial-scale and even shorter climatic changes. The observed SST curve displays characteristic sequences of extremely rapid warming and cooling events along the glacial period. Comparison of this Alboran record with $\delta^{18}O$ from Greenland ice (Greenland Ice Sheet Project 2 core) shows a strong parallelism between these SST oscillations and the Dansgaard-Oeschger events. Five prominent cooling episodes standing out in the SST profile are accompanied by an anomalous high abundance of *Neogloboquadrina pachyderma* sinistral which is confined to the duration of these cold intervals. These features and the isotopic record reflect drastic changes in the surface hydrography of the Alboran Sea in association with Heinrich events H1-5.

1. Introduction

Rapid climatic oscillations of 1000 to 3000 years have been described in the glacial record of Greenland ice cores [*Dansgaard et al.*, 1993; *Grootes et al.*, 1993; *Meese et al.*, 1997]. These oscillations, named Dansgaard-Oeschger (D-O) Stadials and Interstadials (cold and warm phases, respectively), may consist of air thermal shifts of about 15°C [*Johnsen et al.*, 1995; *Schwander et al.*, 1997].

Detailed studies of North Atlantic cores at high time resolution demonstrated that these drastic atmospheric oscillations were also recorded in the ocean sediments. Abrupt changes in foraminiferal assemblages and isotopic composition [*Bond et al.*, 1993; *Voelker et al.*, 1998] or carbonate and magnetic susceptibility records [*Keigwin and Jones*, 1994; *Moros et al.*, 1997] have been related to intensive high-latitude North Atlantic Ocean sea surface temperature (SST) shifts or deep water circulation oscillations, respectively.

Correlations of sedimentalogical changes in lake sediments with Greenland's records have also shown that these atmospheric changes extended southward to the European [*Thouveny et al.*, 1994] and North American [*Phillips et al.*, 1994] continents. The at least hemispheric extension of these events has been documented by their correspondence with deep oxygenation phases in Santa Barbara basin [*Behl and Kennett*, 1996], ice-rafted detritus records in the subarctic Pacific Ocean [*Kotilainen and Shackleton*, 1995], productivity changes in the monsoon-induced upwelling from the Arabian Sea [*Schultz et al.*, 1998], loess sequence from western China [*Chen et al.*, 1997], and sea surface salinity changes in the Japan and China Seas [*Wang and Oba*, 1998]. However, all these evidences are somehow fragmentary, and a global picture describing the D-O phenomena and how they influenced ocean hydrodynamics is still missing.

This lack of knowledge contrasts with the present degree of understanding of other abrupt climatic changes such as the Heinrich events (HE) (rapid periods of massive iceberg melting in the North Atlantic Ocean [*Bond et al.*, 1992; *Broecker*, 1994]) which left a strong imprint in marine sediments. Thus these events can be well correlated along most of the North Atlantic Ocean [*Grousset et al.*, 1993]. Rapid reorganization of the thermohaline circulation has been inferred from reconstruction of surface and deep North Atlantic water conditions in concurrence with some of these massive iceberg discharges [*Cortijo et al.*, 1997; *Vidal et al.*, 1997; *Zahn et al.*, 1997].

Rapid changes in thermohaline circulation have also been invoked in relation to D-O events [*Broecker et al.*, 1990; *Keigwin and Jones*, 1994; *Adkins et al.*, 1997; *Rasmussen et al.*, 1997]. However, accurate SST measurements are required for the modelization of these oscillations [*Paillard and Labeyrie*, 1994; *Rahmstorf*, 1994; *Manabe and Stouffer*, 1995].

We report here the first SST record using C_{37} unsaturated alkenones with sufficient temporal resolution to permit comparison with the millennial-scale events recorded in ice cores from Greenland [*Dansgaard et al.*, 1993]. This SST profile allows the quantification of the intensity of these rapid changes and offers the oportunity to study detailed closeups of both oceanic and atmospheric responses to D-O and HE events. The results evidence clear differences in the marine hydrodynamic changes between both episodes.

2. Area of Study and Oceanographic Setting

The present study concerns the top 16 m of a piston core recovered in the eastern Alboran Sea (MD 952043; 36°8'598''N, 2°37'269''W; 1841 m water depth) by the R/V *Marion Dufresne* during the 1995 International Marine Global Change Studies cruise (Figure 1).

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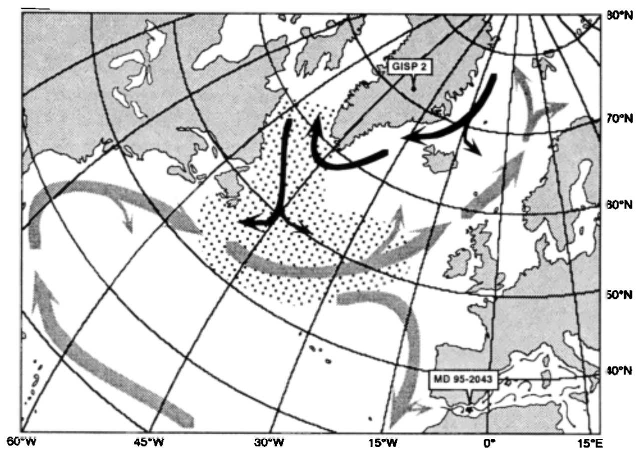

Figure 1. Location of the Alboran core (MD952043) and the Greenland Ice Sheet Project 2 (GISP2) ice core. Arrows indicate the present-day dominant circulation pattern of surface waters in the North Atlantic Ocean and in the western Mediterranean Sea. The dotted area indicates the belt of maximum discharge of ice-rafted debris [*Bond et al.*, 1992; *Groussetet al.*, 1993].

The Alboran Sea is the westernmost basin of the Mediterranean Sea. Its surface water originates from North Atlantic Surface Waters (NASW) which are partially modified on the way through the Gibraltar Strait [*La Violette*, 1984]. This inflow crosses the Alboran Sea, describing two anticyclonic gyres [*Heburn and La Violette*, 1990; *Perkins et al.*, 1990], and then extends to the whole Mediterranean Sea. High-productivity systems are generated in the Alboran Sea in association with the hydrological structures of the inflowing NASW [*Tintoré et al.*, 1988; *Minas et al.*, 1991; *Arnone*, 1994], which contrast with the general oligotrophic character of the Mediterranean Waters. Although this basin is not very distant from the zone of ice-rafted detritus in the North Atlantic [*Grousset et al.*, 1993], so far there is no evidence that it has received direct discharges of ice debris.

3. Methodology

3.1. SST Reconstruction

SST was obtained from the relative composition of C_{37} unsaturated alkenones ($U^{K'}_{37}$ index [*Brassell et al.*, 1986]) by using the calibration equation of *Müller et al.* [1998], $U^{K'}_{37} = 0.033$ SST + 0.044. Recently, some studies of $U^{K'}_{37}$ in suspended particulate matter from the northwestern Mediterranean Sea [*Ternois et al.*, 1997; *Cacho et al.*, 1999] have shown that large biases are obtained from SST estimations when using the general equations for the conversion of $U^{K'}_{37}$ into SST [*Prahl et al.*, 1988; *Müller et al.*, 1998]. In contrast, a good agreement with present SST data is obtained after development of a specific calibration for this area [*Ternois ng et al.*, 1997]. Nevertheless, sedimentary $U^{K'}_{37}$ measurements in this particular zone [*Cacho et al.*, 1999] show a good agreement with average annual SST when calculated with the general equations [*Prahl et al.*, 1988; *Müller et al.*, 1998]. On the other hand, the Alboran Sea is under the influence of the inflowing NASW. Thus the most appropriate conversion equation is that derived from the global $U^{K'}_{37}$ calibration which has also included sediment samples from the North Atlantic Ocean [*Müller et al.*, 1998].

Experimental conditions for $U^{K'}_{37}$ measurement are described by *Villanueva et al.* [1997]. Briefly, sediment samples (~2 g) were freeze-dried and manually ground for homogeneity. After addition of an internal standard mixture containing *n*-nonadecan-1-ol, *n*-hexatriacontane, and *n*-tetracontane, dry sediments were extracted in an ultrasonic bath with dichloromethane. The extracts were hydrolyzed with 6% potassium hydroxide in methanol to eliminate interferences from wax esters. The extracts were derivatized with bis(trimethylsilyl)trifluoroacetamide before instrumental analysis.

Alkenones were analyzed with a Varian gas chromatograph Model 3400 equipped with a septum-programmable injector and a flame ionization detector. The instrument was equipped with a CPSIL-5 CB column coated with 100% dimethylsiloxane (film thickness: 0.12 µm). Hydrogen was the carrier gas (50 cm/s). The oven temperature was programmed from 90° to 140°C at 20°C/min, then to 280°C at 6°C/min (holding time: 25 min), and, finally, to 320°C at 10°C/min (holding time of 6 min). The injector was programmed from 90°C (holding time: 0.3 min) to 320°C at 200°C/min (final holding time: 55 min). Five times replication of a sediment sample with similar lipid content and $U^{K'}_{37}$ index showed a standard deviation of ±0.15°C in temperature estimation.

3.2. *Neogloboquadrina pachyderma* (sinistral) Percentages

Samples for planktonic foraminifera analyses were washed over a 63 µm mesh sieve and dry sieved through a 150 µm mesh sieve. The residues were split, when necessary, into suitable aliquots of about 500 specimens, which were then counted. The percentage of *Neogloboquadrina pachyderma* (sinistral (s)) represents the proportion of these specimens in relation to the total number of planktonic foraminifera counted in the sample.

3.3. Isotope Measurement

Stable isotope measurements of *Globigerina bulloides* were performed in a SIRA mass spectrometer using samples of 25-30 specimens picked from the 300-355 µm size range. The mass spectrometer was fitted with the VG isocard common acid bath system. The analytical precision of laboratory standards was better than ±0.08‰ for $\delta^{18}O$. Calibration to Vienna Peedee belemnite (VPDB) was via the NBS19 standard.

4. Chronostratigraphy

Accelerator mass spectrometry (AMS) radiocarbon ages were measured at the University of Utrecht on 21 picked monospecific foraminiferal samples (Table 1). For the last 21,000 years, ^{14}C ages have been converted into calendar ages using the Calib4.1 program [*Stuiver and Reimer*, 1993], which uses an updated data set [*Stuiver et al.*, 1998] and already includes the 400 year correction for the ocean surface reservoir effect [*Bard et al.*, 1994]. The conversion of ^{14}C ages after 21,000 years is difficult because of the poor knowledge on the quantitative production of cosmogenic ^{14}C [*Laj et al.*, 1996; *Voelker et al.*, 1998] along isotopic stage 3, showing various rises and plateaus [*Bard*, 1998; *Bard et al.*, 1990]. Calibration of the four ^{14}C ages older than 21,000 years to calendar years using different calibration models [*Bard et al.*, 1990; *Laj et al.*, 1996; *Kitagawa and van der Plicht*, 1998] gives rise to deviations greater than 1000 years. In view of these uncertainties, we have constructed two different chronostratigraphic models for the core section older than 21,000 years which do not consider these previous age calibrations.

The first model is based on oxygen isotopic stratigraphy and the AMS ^{14}C ages. Three isotopic events (3.1, 3.13, and 3.3 [*Martinson et al.*, 1987]) have been identified in the $\delta^{18}O$ curve from *G. bulloides* (Table 1). Constant sedimentation rates between the last four carbon ages calibrated with Calib4.1 and the isotopic event

Table 1. The ^{14}C AMS Radiocarbon Ages Measured in Core MD952043 for Chronology and the Isotopic Events Identified in the $\delta^{18}O$ Curve from *G. bulloides* of Core MD952043 That Have Been Used for Age Model 1 (Section 4)

Depth, cm	Sample type	^{14}C age	Error	Calendar years
14	*G. bulloides*	1,980	±60	1,527*
54	*G. bulloides*	3,216	±37	3,011*
96	*G. bulloides*	4,275	±41	4,396*
178	*G. bulloides*	5,652	±42	6,018*
238	*G. bulloides*	6,870	±50	7,401*
298	*N. paquiderma*	8,530	±47	8,976*
348	*G. bulloides*	9,200	±60	9,827*
418	*N. paquiderma*	9,970	±50	10,815*
487	*N. paquiderma*	10,560	±60	11,659*
512	*N. paquiderma*	10,750	±60	12,021*
588	*N. paquiderma*	11,590	±60	13,081*
595	*G. bulloides*	11,880	±80	13,399*
682	*G. bulloides*	12,790	±90	14,289*
708	*N. paquiderma*	13,100	±90	14,762*
758	*N. paquiderma*	14,350	±110	16,617*
802	*N. paquiderma*	15,440	±90	17,871*
858	*N. paquiderma*	18,260	±120	21,116*
958	*N. paquiderma*	22,450	±160	25,981†
1044	*G. bulloides*	24,840	±280	30,040†
1077	*N. paquiderma*	26,220	±240	31,597†
1324	*N. paquiderma*	32,600	±400	38,812†
Depth, cm	**Isotopic Event**	**Age**	**Reference**	
1026	3.1	29,500	*Sarnthein and Tiedemann*, 1990	
1490	3.13	43,880	*Martinson et al.*, 1987	
1600	3.3	50,210	*Martinson et al.*, 1987	

*Converted to calendar ages with Calib4.1 program [*Stuiver and Reimer*, 1993].

†Converted to calendar ages as described in section 4 (age model 1).

3.1 have been assumed, and the regression line (r = 0.9962) has been calculated. Calendar ages for the ^{14}C dated samples at 958, 1044, and 1077 cm have been assigned using the curve-fitted equation (Table 1). The ratio between calendar and ^{14}C ages in these three samples is nearly constant (mean 1.2, standard deviation 0.03) and has been used for the calendar age calibration of the ^{14}C measurement at 1324 cm (Table 1). The resulting value is fully consistent with the isotopic data (Figure 2) and in agreement with the age calibration suggested by *Stuiver et al.* [1998], based on only two coral meassurements.

The second age model is based on the correlation of SST in MD952043 with the $\delta^{18}O$ ice core record from the Greenland Ice Sheet Project 2 (GISP2) [*Grootes et al.*, 1993; *Meese et al.*, 1997]. The high parallelism of the SST changes with the D-O events recorded in Greenland allows a peak to peak correlation between the two signals. The resulting chronology does not differ significantly from that in the first age model, and the small misalignments are well within dating uncertainties (Figure 2). The two chronologies show that the studied section provides a continuous record over the last 52,000 years with very high sedimentation rates (average of 30 cm kyr^{-1}). The core was sampled at 4-6 cm intervals, corresponding to an average temporal resolution of about 200 years.

5. Results and Discussion

5.1. Sea Surface Temperatures in the Alboran Sea

The obtained SST profile (Figure 3a) shows a continuous record for the last 52 kyr covering most of the last glacial period, Termination I, and Holocene with high time resolution. Relatively stable temperatures are observed during the Holocene which strongly contrast with the abrupt shifts in the glacial period. SST oscillations in the Holocene are not greater than 2°C (20°-18°C), while the rapid SST changes of the glacial period range between 9° and 16°C. Termination I involved an overall warming of nearly 10°C interrupted by few cooling events, among which a prominent event at about 12 ka is clearly correlated with the Younger Dryas. This event is marked by a 4°C drop and a duration of about 800 years.

The abrupt and intense thermal oscillations recorded in the Alboran core in the glacial period are coherent with the Greenland $\delta^{18}O$ ice record in GISP2 [*Grootes et al.*, 1993; *Meese et al.*, 1997] (Figure 3d). Linear correlation of GISP2 $\delta^{18}O$ with Alboran SST for the last 52,000 years using our age models gives correlation coefficients of r = 0.89 and 0.92 for the first and second, respectively (section 4), which are remarkably high when compared with other analogous correlations between sedimentological data and Greenland ice $\delta^{18}O$ profiles [*Thouveny et al.*, 1994]. These strong correlations evidence a close connection between climate over Greenland and oceanographic conditions in the Alboran Sea. The errors in the ^{14}C measurements (Table 1) and the even greater uncertainties of the conversion from ^{14}C to calendar ages for periods older than 21,000 year B.P. prevents assumption of synchroneity [*Crowley*, 1999]. Nevertheless, the strong agreement between the two records suggests that the SST changes in the Alboran Sea are phase locked with respect to the D-O events described in the GISP2 core.

This good parallelism indicates that surface water conditions in the Alboran Sea remained closely linked to water advection from the Atlantic Ocean during the last glacial and interglacial periods. MD952043 SST data give further evidence for the dominance of the modern Mediterranean antiestuarine circulation system throughout the studied period, which includes the most recent eastern Mediterranean sapropel (S1: 9000-6000 years), supporting previous geologic evidences [*Zahn and Sarnthein*, 1987] and re-

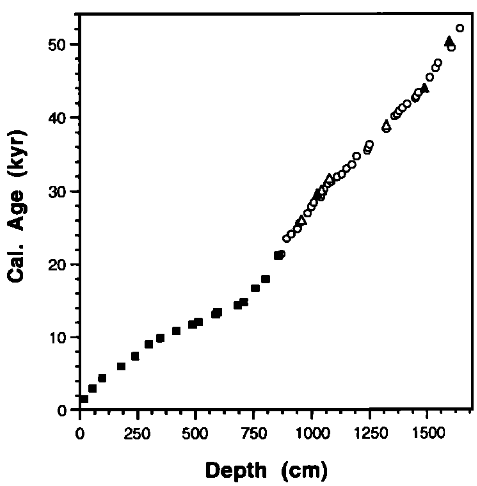

Figure 2. Age control points used in the two different age models of core MD952043. The solid squares refer to the accelerator mass spectrometry (AMS) ^{14}C ages calibrated with the Calib4.1 program that are used in the two age models. The triangles refer to the points used in the first age model, including the isotopical events (solid triangles) and the AMS ^{14}C ages calibrated with our model (open triangles). The open circles refer to the points used in the second age model based on peak to peak correlation of the MD952043 sea surface temperature (SST) record with the GISP2 $\delta^{18}O$ profile [*Grootes et al.*, 1993; *Meese et al.*, 1997].

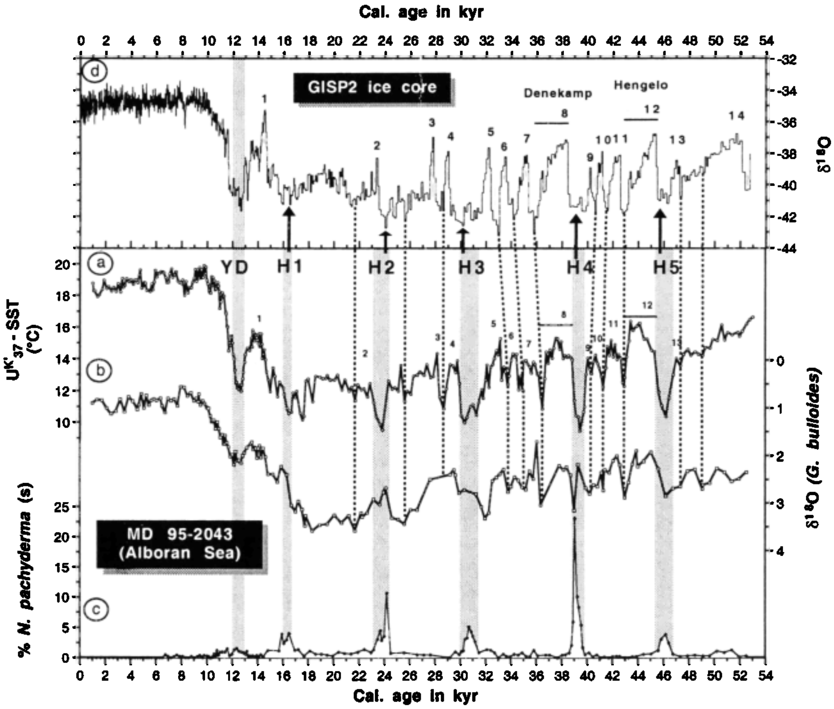

Figure 3. (a) Sea surface temperature in the Alboran Sea (core MD952043) based on measurements of the relative composition of unsaturated C_{37} alkenones, (b) $\delta^{18}O$ in *G. bulloides* in core MD952043 (plotted with reversed *Y* axis), (c) Relative abundance (%) of *N. pachyderma* (s) in core MD952043, and (d) $\delta^{18}O$ record from the GISP2 ice core [*Grootes et al.*, 1993; *Meese et al.*, 1997]. Figures 3a-3c are plotted using the first age model described in section 4 based on AMS^{14}C dates and $\delta^{18}O$ isotopic stratigraphy. Numbers over the graph mark the position of the Dangaard-Oeschger interstadials [*Dansgaard et al.*, 1993]. Arrows indicate the correspondence of $\delta^{18}O$ in GISP2 and Heinrich events as defined by *Bond et al.* [1993, 1997].

sults on paleocirculation models [*Myers et al.*, 1998]. Thus the results reported in the present study disagree with the possibility of a reversal to estuarine conditions in the Mediterranean Sea during the time of sapropel formation [*Thunell and Williams*, 1989].

5.2. Heinrich Events

The five prominent minima standing out in the SST profile of the glacial period occur always at the end of a series of abrupt progressively shorter warming and cooling events which compare with the Bond cycles [*Bond et al.*, 1993; *Broecker*, 1994]) (Figure 3a) and are time coincident with HE 1-5 described in the North Atlantic Ocean [*Bond et al.*, 1992; *Grousset et al.*, 1993; *Broecker*, 1994; *Bond and Lotti*, 1995; *Cortijo et al.*, 1995, 1997; *Rasmussen et al.*, 1997; *Chapman and Shackleton*, 1998; *Elliot et al.*, 1998; *Vidal et al.*, 1997]. These SST minima are observed concurrently with raises in *N. pachyderma* (s), 5-20% of total foraminifera (Figure 3c). The high percentages of these species in core MD952043 are anomalous for the Alboran Sea, being confined to these events. *N. pachyderma* (s) is usually found in the North Atlantic Ocean during HE [*Bond et al.*, 1993; *Broecker*, 1994; *Fronval et al.*, 1995], reaching 100% of total foraminifera at high latitudes [*Bond et al.*, 1993] and 25-50% in the south Portuguese margin [*Lebreiro et al.*, 1996; *Cayre et al.*, 1999]. The concurrence of SST drops and *N. Pachyderma* (s) raises at the time of H1-5 in the Atlantic gives further ground to a link of the SST changes in the Alboran Sea and the development of HE. Furthermore, the occurrence of foraminiferal species down core having no analogue in the present Mediterranean underlines the value of C_{37} unsaturated alkenones for temperature estimation rather than relying only on the foraminiferal faunas as temperature proxy.

Some previous studies have documented the arrival of icebergs farther south to the North Atlantic Ice-Rafted Detritus Belt (IRDB), mainly along the Iberian margin [*Lebreiro et al.*, 1996; *Zahn et al.*, 1997; *Cayre et al.*, 1999]. Recently, rapid drops in sea surface temperature and salinity related to the occurrence of IRDB have been described in a core as south as 37°N in the south Iberian margin [*Cayre et al.*, 1999]. Our results reinforce these evidences, suggesting that the extension of the polar waters arrived farther south than was previously believed [*Ruddiman and McIntyre*, 1981], allowing its entrance into the Mediterranean Sea. Therefore, in contrast with previous assumptions [*Rohling et al.*, 1998], our results support the occurrence of important polar water invasions via the Strait of Gibraltar during H1-5. The unreported presence of *N. pachyderma* in previous studies of the area [*Pujol and Vergnaud-Grazzini*, 1989; *Rohling et al.*, 1995] was probably due to problems of core resolution.

The oxygen isotope record from *G. bulloides* (Figure 3c) has a lower resolution than the SST curve, but there is a general agreement between the two records. During the HE, the correlation is made difficult by the absence of heavy $\delta^{18}O$ values, as expected for such cold SSTs. On the contrary, some HE, namely, H2 and H4, apparently contain an isotopic depletion. Although it can not be discarded that this loss of parallelism between the SST and the $\delta^{18}O$ curves during the HE may reflect some poorly understood process in the Mediterranean Sea, the discrepancy is probably related to surface water salinity reduction [*Duplessy et al.*, 1992] in the source region of the North Atlantic during massive iceberg melting [*Zahn et al.*, 1997].

The magnitudes of the SST reductions are very similar for all the HE (*ca* 4°C) and similar to those described in midlatitudes of the North Atlantic Ocean [*Chapman and Shackleton*, 1998]. Even H3, which has not been identified in some North Atlantic Ocean areas [*Cortijo et al.*, 1995; *Chapman and Shackleton*, 1998] nor in the Iberian margin [*Zahn et al.*, 1997], is well represented in the Alboran record. H2 and H4 display slightly colder temperatures than the others, in agreement with maximum *N. pachyderma* percentages pointing to these HE as the coldest phases of the studied record. The cooling rates are different for each event, H4 and H2 being the fastest. H4 appears to exhibit the most abrupt SST cooling in the whole record, involving a drop of 2.5°C in less than a century. In contrast, H3 is characterized by a relatively slow cooling (3.7°C over 1300 years) and a final sharp warming (4°C in 500 years). These rapid warmings are characteristic of the Heinrich episodes [*Bond et al.*, 1992; *Broecker*, 1994] and have been attributed to enhancement of the ocean's thermohaline circulation after ice sheet retreat [*Paillard and Labeyrie*, 1994].

5.3. Differences Between the Dansgaard-Oeschger Stadials and the Heinrich Events

The cold D-O Stadials are also recorded in the Alboran Sea by significant SST drops (1°-3°C), but they are of lesser intensity than those related to the HE. This feature contrasts with the trend observed in Greenland cores, where the temperature drop reach almost the same values during all the D-O stadials, even those associated with the HE [*Grousset et al.*, 1993; *Meese et al.*, 1997]. In addition, the absence of *N. pachyderma* during the stadials (Figure 3b) also points to a less intense cooling related to these events. Another distinctive feature is observed in the isotope curve, which shows small but clear $\delta^{18}O$ enrichment during D-O stadials, in contrast with the HE (Figure 3c), suggesting a temperature change without salinity dilution [*Duplessy et al.*, 1992]. All these evidences indicate that during the D-O Stadials, surface water in the Alboran Sea does not change as drastically as during the HE although this marine region was also very sensitive to these climatic changes.

Causes and consequences of D-O coolings are not well understood, and their relation with the HE is a present subject of debate. Layers of ice-rafted detritus in synchrony with the D-O stadials have been identified in sediments from the North Atlantic Ocean [*Bond et al.*, 1993; *Bond and Lotti*, 1995; *Fronval et al.*, 1995; *Moros et al.*, 1997], although their influence is restricted to northern latitudes [*Bond et al.*, 1992; *Broecker*, 1994]. These evidences, added to the feature that all D-O stadials, including the HE, show a very similar signature in Greenland ice cores [*Grootes et al.*, 1993; *Meese et al.*, 1997], point to a common trigger mechanism for all these cold events [*Stocker et al.*, 1999]. Reduced convection in the Norwegian Sea and slower deep water circulation in the stadials than during the interstadials was proposed by *Rasmussen et al.* [1996], suggesting that deep water convection moved farther south than the present position of the subpolar Atlantic Ocean during the stadial periods. To this end, the changes in SST and surface salinity associated with D-O observed in the Bermuda Rise [*Keigwin and Jones*, 1994; *Adkins et al.*, 1997] have also been associated with shifts in NADW flux.

Nevertheless, these studies do not propose any distinctive process for the stadials associated with HE. Coupled ocean-atmosphere models have demonstrated that the thermohaline circulation constitutes a major feedback mechanism for the transmission of climatic changes [*Birchfield et al.*, 1994; *Paillard and Labeyrie*, 1994; *Rahmstorf*, 1994; *Manabe and Stouffer*, 1995], showing that climatic oscillations can trigger transitions between different states of the thermohaline circulation. *Stocker et al.* [1999] argued that the complete shutdown of the NADW formation occurred only during HE 1, 4, and 5, leading to a warming in the Southern Hemisphere [*Blumier et al.*, 1998], while during the D-O Stadials, NADW switched between different convection sites in the North Atlantic Ocean.

The different signatures observed in the Alboran record between HE and D-O Stadials (Figure 3) are consistent with different mechanisms of propagation of these climatic changes from high to medium latitudes and/or to their differential modes of operation. The overlaying of the Alboran Sea and Greenland paleotemperature records (Figure 4) evidences that while the oceanographic response to a cooling was amplified during the HE, during the rest of the D-O Stadials it was weakned. This difference suggests the activation of a positive oceanographic feedback during the HE, consistently with the changes in the thermohaline circulation already described for these episodes [*Zahn et al.*, 1997; *Vidal et al.*, 1997]. The complete shutdown of NADW formation proposed by *Stocker et al.* [1999] could explain the southward extension of the polar front allowing the entrance of polar water through the Strait of Gibraltar. Nevertheless, all the H1-5 events are well recorded in the Alboran Sea, reaching similar low SSTs and pointing to mechanisms of the same intensity for all HE events. Thus, if there was not a complete shutdown of the NADW formation during HE 2 and 3 as proposed by *Stocker et al.* [1999], maybe the weakening of the mechanism was effective enough to lead to an impact similar to HE 1, 4, and 5 in the Alboran Sea.

The atmosphere can also involve efficient interlatitudinal transport of rapid climatic changes. Evidences from Greenland ice cores indicate that wind speeds increased during every cold D-O stadial [*Taylor et al.*, 1993]. Moreover, studies of sedimentary sequences from the Chinese Loess Plateau show that strengthened winter Westerlies associated with the HE and D-O Stadials led to cold and dry episodes in the Asian land mass with enhanced loess deposits [*Chen et al.*, 1997; *Wang and Oba*, 1998].

Rapid changes from cold to temperate climatic conditions in consonance with the D-O events have been recorded in lake deposits from the south of France [*Thouveny et al.*, 1994]. Similarly, a connection between the D-O events and oscillations in the evaporation/precipitation ratio of some Italian lakes has been observed [*Ramrath et al.*, 1999]. These evidences of coupling between D-O events and lake hydrology in the Mediterranean region also point to a rapid atmospheric connection between Greenland

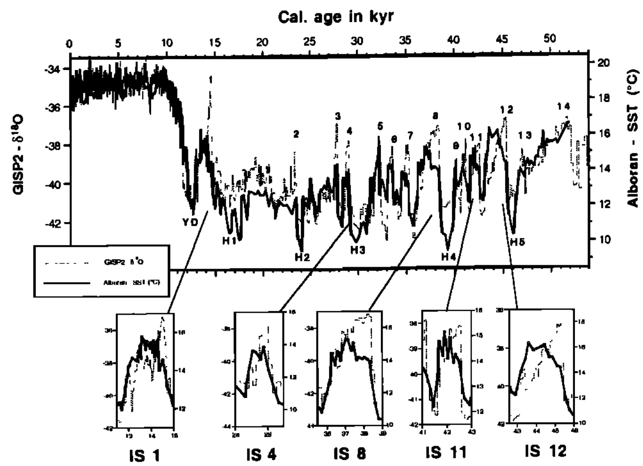

Figure 4. Comparison of Alboran SST in core MD952043 (solid line) and $\delta^{18}O$ in GISP2 [*Grootes et al.*, 1993; *Meese et al.*, 1997] (dotted line) using the second timescale of MD952043 based on peak to peak correlation with GISP2 (section 4). The zoom boxes illustrate the differences in interstadial temperature evolution between GISP2 (air) and Alboran (sea surface).

Table 2. Summary of the Observations in $\delta^{18}O$ GISP2 and SST MD952043 Records During the Dansgaard-Oeschger Stadials and Interstadials and the Heinrich Events and the Inferred Climatic Mechanisms Allowing their Interpretation

	Observations		Mechanisms	
Events	Air Temperature Over Greenland	Sea Surface Temperature at Alboran Sea	Latitude of the Polar Front	Northern Hemisphere Wind Strength
D-O interstadials	early maximum/ cooling trend	late maximum/ warming trend	high latitudes	low
D-O stadials	similar cold values	intermediate cold values	northern 37°N	high
Heinrich events	similar cold values	coldest values	southern 37°N	high

and the Mediterranean Sea. Cold winds and enhanced formation of Western Mediterranean Deep Water in the Gulf of Lions during the HE has been recently suggested by *Rholing et al.* [1998]. Thus colder and more windy conditions could have led to the SST coolings in the Alboran Sea observed during the D-O Stadials and HE, since this is a semienclosed sea with a relatively small water volume that may respond faster than the open ocean to any climatic alteration.

5.4. Dansgaard-Oeschger Interstadials

The present high-resolution study of SSTs allows a detailed analysis of the thermal evolution along every warm D-O interstadials. The close comparison of both Alboran and Greenland records (Figure 4) illustrates a repeated SST pattern with some characteristics differing from those observed in Greenland. Temperature evolution in Greenland consists of rapid warming (within years or a few decades), reaching maximum values at the beginning of the interstadial, and thereafter a continuous stepwise cooling trend. In contrast, warming in the Alboran Sea also is initially rapid, but after this first pulse, SST stabilizes and continues to warm at a lower rate, reaching the highest values toward the end of the interstadial (Figure 4). All interstadial cessations occurred in a single and rapid cooling pulse. Similar abrupt coolings linked to D-O Interstadials have also been observed in the SST record from the Norwegian Sea [*Rasmussen et al.*, 1996]. These rapid coolings point to the occurrence of a rapid interlatitudinal teleconnection between the Alboran Sea and high latitudes. Nevertheless, such a connection mechanism was less effective in cooling Alboran waters than those operating during the HE (Figure 4). We propose that the intensification the Northern Hemisphere wind system, as mentioned above, was the main mechanism responsible of the D-O coolings in the Alboran Sea (Table 2). On the other hand, during HE, changes in the global thermohaline circulation added to the wind intensification would have been responsible for the oceanic amplification of the climatic signal during these periods.

6. Conclusions

The study of a high-resolution core has allowed the establishment of the response of the Alboran Sea to the rapid climatic changes described in the Greenland ice cores. Age models have been established on the basis of several AMS ^{14}C age determinations, identified events in the $\delta^{18}O$ record of *G. bulloides*, and the alignment of some events with the $\delta^{18}O$ GISP2 record, providing correlation coefficients of $r= 0.9$. The good parallelism between the SST in the Alboran Sea and GISP2 $\delta^{18}O$ records, showing the oscillations associated with all HE 1-5 and D-O Stadials and Interstadials, evidences this region as being very sensitive to rapid climatic changes and provides information on the possible connection mechanisms between the Alboran Sea and climate dynamics at high latitudes.

HE 1-5 are well represented in the Alboran Sea by significant SST drops (4°C) and anomalous high percentages of *N. pachyderma*. The close comparison of these results with $\delta^{18}O$ GISP2 points to the existence of an oceanographic mechanism which amplified the climatic signal related to these events. The complete shutdown and/or strong weakening in the global thermohaline circulation would force the southward displacement of the polar front, allowing the entrance of North Atlantic polar water through the Strait of Gibraltar during HE.

Comparison of the $U^{K'}_{37}$ SST from the Alboran Sea and $\delta^{18}O$ from GISP2 core shows cooling events in concurrence with the D-O Stadials in the marine record. However, these events are of lower intensity than the SST drops concurrent with the HE. Foraminiferal and isotopic records also suggest that the oceanographic changes associated with D-O stadials were less severe than during the HE. Intrainterstadial examination of SST evolution in the Alboran Sea evidences a time-dependent temperature pattern different from that observed in the Greenland cores. While $\delta^{18}O$ in GISP2 reaches thermal maximum at the beginning of the Interstadial and then decreases following a stepwise mode, Alboran SST reaches the maximum value at the end of the interstadial and is followed by a sharp drop. These evidences point to the operation of a rapid teleconnection mechanism during the D-O stadials, which led to less drastic modification of surface water properties. Rapid intensification of the Northern Hemiphere winds is quoted as being mainly responsible for Alboran SST coolings during the D-O stadials. During HE, thermohaline circulation operated as the main positive feedback of the climatic signal although atmospheric circulation could have also played some role.

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