

# Variability of the western Mediterranean Sea surface temperature during the last 25,000 years and its connection with the Northern Hemisphere climatic changes

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**Abstract.** Sea surface temperature (SST) profiles over the last 25 kyr derived from alkenone measurements are studied in four cores from a W-E latitudinal transect encompassing the Gulf of Cadiz (Atlantic Ocean), the Alboran Sea, and the southern Tyrrhenian Sea (western Mediterranean). The results document the sensitivity of the Mediterranean region to the short climatic changes of the North Atlantic Ocean, particularly those involving the latitudinal position of the polar front. The amplitude of the SST oscillations increases toward the Tyrrhenian Sea, indicating an amplification effect of the Atlantic signal by the climatic regime of the Mediterranean region. All studied cores show a shorter cooling phase (700 years) for the Younger Dryas (YD) than that observed in the North Atlantic region (1200 years). This time diachroneity is related to an intra-YD climatic change documented in the European continent. Minor oscillations in the southward displacement of the North Atlantic polar front may also have driven this early warming in the studied area. During the Holocene a regional diachroneity propagating west to east is observed for the SST maxima, 11.5-10.2 kyr B.P. in the Gulf of Cadiz, 10-9 kyr B.P. in the Alboran Sea, and 8.9-8.4 kyr B.P. in the Tyrrhenian Sea. A general cooling trend from these SST maxima to present day is observed during this stage, which is marked by short cooling oscillations with a periodicity of 730±40 years and its harmonics.

## 1. Introduction

The Mediterranean Sea is a semienclosed water body surrounded by large continental masses with only one narrow connection to the Atlantic Ocean. The hydrodynamics of the basin is controlled by the North Atlantic water inflow, wind regime, and climate of the surrounding lands. Recent research on the last glacial period has demonstrated a dependence of the western basin from the millennial-centennial climatic and oceanographic variability in the North Atlantic [Rohling *et al.*, 1998; Cacho *et al.*, 1999a; Paterne *et al.*, 1999]. However, the last deglaciation and Holocene periods were also characterized by short-term climatic variability, which was recorded either as intense, e.g., the Younger Dryas (YD), or lower-amplitude oscillations [Bond *et al.*, 1997].

The present study is devoted to elucidate the significance of this short term variability for the western Mediterranean. For this purpose, alkenone-derived sea surface temperatures (SST) and  $\delta^{18}\text{O}$  profiles of *Globigerina bulloides* have been analyzed at

high resolution in a transect of four marine cores for the last 25 kyr. These cores are situated at similar latitude but represent different hydrographic conditions, from open ocean waters (Gulf of Cadiz) to enclosed basin conditions (Tyrrhenian Sea) including the Atlantic-Mediterranean exchange area (Alboran Sea). The sampling resolution and age models afford the identification of centennial-millennial-scale oscillations. The results are discussed in the context of the previous knowledge on marine and terrestrial climatic variability of the Mediterranean and North Atlantic regions.

## 2. Core Location and Oceanographic Setting

The cores studied encompass a series of four sites from similar latitude, located on a W-E latitudinal transect from the Gulf of Cadiz in the Atlantic Ocean to the southern Tyrrhenian Sea (Figure 1). Core M39-008 (39°22.8'N, 7°4.6'W, water depth 576 m, core length of 570 cm) was recovered from the northeastern Gulf of Cadiz by RV *Meteor* [Schott *et al.*, 1999]. Piston core MD 95-2043 was obtained in the central Alboran Sea (36° 8.6' N; 2° 37.3' W; 1,841 m water depth; core length 3600 cm) by RV *Marion Dufresne* during the 1995 International Marine Global Change Studies (IMAGES) cruise. Cores BS79-38 (38°24.7'N, 13°34.6'E, water depth of 1489 m, 515 cm long) and BS79-33 (38°15.7'N, 14°1.8'E, water depth of 1282 m, 485 cm long) were recovered from neighboring locations in the southern Tyrrhenian Sea.

The Mediterranean Sea is a concentration basin exchanging water with the open ocean only through the Strait of Gibraltar. The water transport at this site is conformed by a low-salinity surface layer of Atlantic Water entering into the Mediterranean Sea and a deep salt-rich layer of Mediterranean water flowing out to the Atlantic Ocean. The Atlantic Water inflow comes directly from the Gulf of Cadiz, enters into the Alboran Sea, where it describes two anticyclonic gyres, and continues eastward along the Algerian coast forming eddies (Figure 1). About one third of

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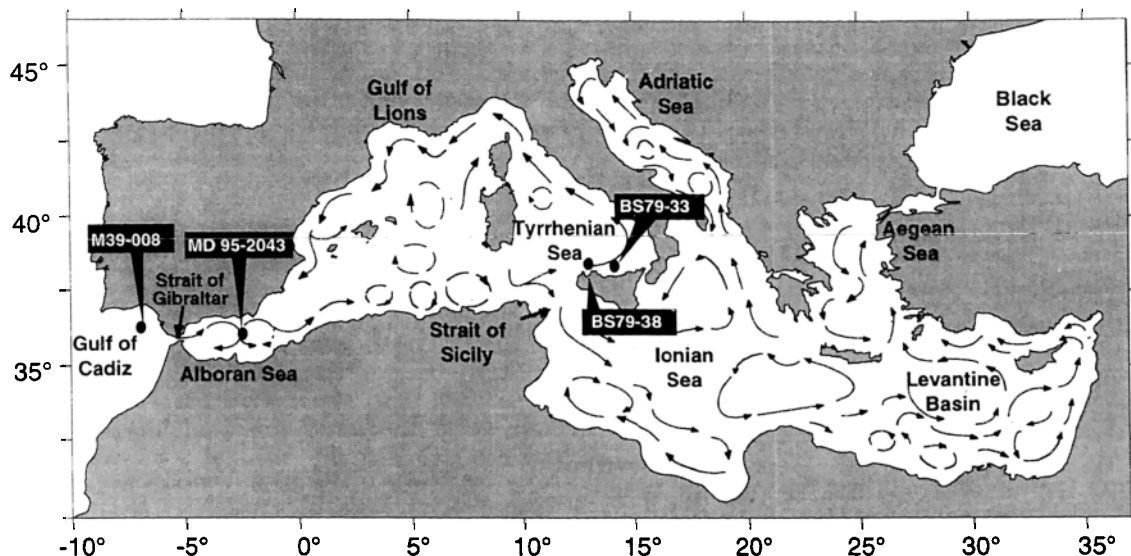
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**Figure 1.** Location of the four cores considered for study. The presently dominant superficial circulation pattern is indicated.

the flow of the now modified Atlantic Water enters the Tyrrhenian Sea following the northern Sicilian coast. The remaining two thirds of the Atlantic Water cross the Strait of Sicily to the eastern Mediterranean [Millot, 1987]. All cores selected for this study are located directly under the present-day path of this Atlantic Water mass.

Vertical mixing in the Alboran Sea occurs because of the strong shear at the interface between inflowing and outflowing water masses through the relatively narrow and shallow Gibraltar sill [Bray *et al.*, 1995]. The anticyclonic gyre described by the inflowing Atlantic jet into the western Alboran Sea forms an upwelling cell which brings cold water to the surface along the Spanish coast between Gibraltar and Malaga [Perkins *et al.*, 1990]. In addition, noticeably vertical and lateral fluxes are related to frontal regions developed at the boundaries between the Atlantic jet and surrounding waters [Peinert and Miquel, 1994]. Therefore, although the cores selected for study are located at almost the same latitude (36°–38°N), they monitor different stages of alteration of the Atlantic Water entering into the Mediterranean Sea.

### 3. Experimental Section

#### 3.1. Sea Surface Temperature (SST) Reconstruction

SST was obtained from the relative composition of  $C_{37}$  unsaturated alkenones through the  $U^{K'_{37}}$  index [Brassell *et al.*, 1986]. Sediment samples (~2 g) were freeze-dried and manually grounded. 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, and the alkenones were recovered with hexane. The solvent was evaporated to dryness with a  $N_2$  stream, and the extracts were finally redissolved with toluene and 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 of 0.12  $\mu$ m). 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 of 0.3 min) to 320°C at 200°C/min (final holding time was 55 min). For further details on analytical conditions used in this study, see Villanueva *et al.* [1997]. Replication of a sediment sample with similar lipid content and  $U^{K'_{37}}$  index ( $n = 5$ ) showed a standard deviation of  $\pm 0.15^\circ\text{C}$  in temperature estimation.

The cores from this study are located within the influence of the inflowing North Atlantic Surface Water. The most appropriate conversion equation is therefore the general  $U^{K'_{37}}$  calibration ( $U^{K'_{37}} = 0.033 \text{ SST} + 0.044$ ), which includes sediment samples from the North Atlantic Ocean [Müller *et al.*, 1998]. Recent  $U^{K'_{37}}$  studies carried out on suspended particulate matter from the Gulf of Lions and surrounding area have shown that the general  $U^{K'_{37}}$  calibration [Prahel *et al.*, 1988; Müller *et al.*, 1998] provides anomalous SST in this zone [Ternois *et al.*, 1997; Bentaleb *et al.*, 1999; Cacho *et al.*, 1999b]. In this area the water column measurements reflect a good correlation with present SST data when a specific calibration is used [Ternois *et al.*, 1997]. Nevertheless, even in this particular zone, the sedimentary  $U^{K'_{37}}$  indices only provide reliable average annual SST through the general equation [Cacho *et al.*, 1999b]. This general equation [Müller *et al.*, 1998] is therefore the one chosen for transformation of the  $U^{K'_{37}}$  data into SST in the present study.

#### 3.2. Isotopic Measurements

Stable isotope measurements of *G. bulloides* for cores MD 95-2043, BS79-33, and BS79-38 were made in a SIRA mass spectrometer at the Godwin Laboratory (University of Cambridge) using samples of 25–30 specimens picked from the 300–355  $\mu$ m size range. The mass spectrometer is fitted with the VG isocarb common acid bath system. Analytical reproducibility of laboratory standards is better than  $\pm 0.08\text{‰}$  for  $\delta^{18}\text{O}$ . Calibration to Vienna Peedee belemnite (VPDB) is via the NBS19 standard.

In core M39-008, stable isotope measurements were carried out with 9–11 specimens of *G. bulloides* from the size fraction

>250  $\mu\text{m}$ , each 5 cm. The tests were cracked and ultrasonically rinsed in methanol prior to analysis. Measurements were done at GEOMAR isotope laboratory with a CARBO KIEL automated carbonate preparation device coupled to a FINNIGAN MAT 251 mass spectrometer. Reproducibility for  $\delta^{18}\text{O}$  was  $\pm 0.056\text{‰}$  ( $1\sigma$  value,  $n=42$  internal carbonate standard measurements).

## 4. Stratigraphy

### 4.1. Age Models for Sediment Cores

Radiocarbon ages for all cores (Table 1) have been converted into calendar ages with the Calib. 4.1 program which uses an updated calibration data set [Stuiver et al., 1998] and includes the correction for ocean surface reservoir effects [Bard et al., 1994]. The age model for core M39-008 has been constructed by linear interpolation between 10 calibrated  $^{14}\text{C}$  accelerator mass spectrometry (AMS) carbon ages and two more age control points determined from the  $\delta^{18}\text{O}$  profile of *G. bulloides*. The  $^{14}\text{C}$  ages were determined in monospecific samples of *Globigerinoides ruber* (white) containing 555-1037 tests from the >150  $\mu\text{m}$  size fraction. The samples were cleaned with 15-30 %  $\text{H}_2\text{O}_2$  in an ultrasonic bath before analysis. Radiocarbon ages were determined at the Leibniz-Labor AMS facility of Kiel University [Nadeau et al., 1997]. The precision of the ages ranges from  $\pm 25$  to  $\pm 150$  years (standard deviation). The two additional  $\delta^{18}\text{O}$  points were selected by correlation with the *G. bulloides*  $\delta^{18}\text{O}$  profile of core SU81-18, the neighbor core previously dated by several  $^{14}\text{C}$  AMS ages [Bard et al., 1987]. These two points confirm the timing of the ice-rafted detritus (IRD) layer corresponding to Heinrich event H1 (Figure 2).

Core MD 95-2043 has an accurate chronostratigraphy based on 18  $^{14}\text{C}$  AMS ages for the last 20 kyr (Table 1). AMS measurements were determined in the University of Utrecht with a precision ranging from  $\pm 37$  to  $\pm 120$  years. The older section has been dated by correlation of the alkenone SST profile with the  $\delta^{18}\text{O}$  record of the Greenland ice core GISP2 (see more details in the work by Cacho et al. [1999a]). The correlation coefficient between MD 95-2043 SST and GISP2  $\delta^{18}\text{O}$  over the  $^{14}\text{C}$  dated interval is extremely high ( $R=0.92$ ).

Seven AMS carbon ages have been used for the age model of core BS79-33 (Tyrrhenian Sea). These were measured in monospecific samples of *G. ruber* or *G. bulloides* (Table 1) in the laboratory of the Centre des Faibles Radioactivités (Gif-sur-Yvette), providing a precision between  $\pm 70$  and  $\pm 220$  years. Subsequent improvement of the age model has been obtained by correlation of the  $\delta^{18}\text{O}$  records from *G. bulloides* and *G. ruber* with those of cores AC 85-4, GT 85-5 and ET 91-18 (Tyrrhenian Sea), which were dated with a large number of  $^{14}\text{C}$  AMS measurements [Capotondi et al., 1999]. Combination of the age points resulting from the two dating series has provided a consistent age model. In core BS79-38 the planktonic  $\delta^{18}\text{O}$  profiles have been correlated with those reported from Capotondi et al. [1999].

### 4.2. Chronostratigraphy of the Last Deglaciation

Traditionally, paleoceanographical and paleoclimatological studies covering the last deglaciation, Termination I (TI), use a classical sequence of climatic periods (Oldest Dryas, Bølling, Older Dryas, Allerød, Younger Dryas, and Preboreal), which was first proposed for the Fenoscandian region on the basis of pollen evidence [Mangerud et al., 1974]. The world wide use of this terminology has been widely criticized because it has several

different connotations (chronozones and biozones) and boundary age uncertainties [Ammann and Lotter, 1989; Broecker, 1992; Walker, 1995; Wohlfarth, 1996; Björck et al., 1998]. To avoid confusion derived from the classical nomenclature, a new event stratigraphy has been proposed for the North Atlantic region on the basis of the  $\delta^{18}\text{O}$  from Greenland ice cores [Björck et al., 1998; Walker et al., 1999]. For this purpose, the GRIP  $\delta^{18}\text{O}$  profile is divided in a series of Greenland stadial (GS) and interstadial (GI) periods following the procedure used in marine isotopic stratigraphy. This event stratigraphy provide a continuous and more detailed subdivision for the Last Glacial-Interglacial transition oscillations, fixing clearly the location and timing of the different boundaries without the problems derived from the  $^{14}\text{C}$  dating [Ammann and Lotter, 1989; Björck et al., 1996; Hughen et al., 1998]. This new stratigraphy is used in the present study, but the terminology of some of the traditional events is occasionally used in the text. For clarification of possible confusion, the new stratigraphy is compared with the traditional terminology in Figure 3.

## 5. Results

### 5.1. Isotope Records

All studied  $\delta^{18}\text{O}$  profiles from *G. bulloides* (Figure 2) show a good agreement in both the general trends and the timing of the major oscillations. This feature verifies the consistency of the different core chronologies. The onset of the isotopic depletion associated to the last deglaciation appears at  $\sim 16.5$  kyr B.P., and a second depletion phase started at 15 kyr B.P.. The YD is represented in all the cores by a short enrichment (0.25‰ in the Gulf of Cadiz and 0.75‰ in the Mediterranean cores), which lasted  $\sim 1100$  years. All profiles reached Holocene values at 10 kyr B.P..

Absolute values are always lower in the Gulf of Cadiz than in the Mediterranean cores, a difference that is larger during the glacial period (1-1.5‰) than during the Holocene (0.5‰). All Mediterranean cores show similar values during the Holocene, while glacial and deglaciation values are lower by  $\sim 0.5\text{‰}$  in the Alboran Sea than in the Tyrrhenian Sea.

### 5.2. Sea Surface Temperature Records

SST curves display similar trends than those from the isotopes (Figure 4) but in general, they show more abrupt oscillations. Minimum values are recorded during the glacial period. Warming of the last deglaciation lasted  $\sim 6000$  years, with total SST increases of  $\sim 7^\circ$  and  $\sim 9^\circ\text{C}$  in the Gulf of Cadiz and the Mediterranean Sea, respectively. Holocene values were always reached before 11 kyr B.P., arriving to maximum temperatures in the early Holocene, although not synchronously between the different areas. Thereafter, Holocene SST follow a weak cooling trend towards present day. The major difference between Holocene SST in the different areas concerns the first phase of the deglaciation warming. This occurred as an extremely abrupt pulse in the Gulf of Cadiz while it was slower in the Mediterranean sites. The YD is well marked in all cores by a  $3^\circ$ - $4^\circ\text{C}$  cooling which lasted  $\sim 700$  years.

Absolute values are always higher in the Gulf of Cadiz than in the Mediterranean which is in agreement with the present-day oceanographical pattern (see section 2). This Atlantic-Mediterranean gradient was larger during the glacial period ( $\sim 4^\circ\text{C}$ ) than during the Holocene ( $\sim 1$ - $2^\circ\text{C}$ ). Similar values are recorded between the Mediterranean cores, although the Tyrrhenian cores always show larger amplitude in the short oscillations, especially during the Holocene.

**Table 1.** Dated Points Used in the Age Models Indicating the Measured  $^{14}\text{C}$  Years and the Calibrated Ages Obtained With the Calib 4.1 Program<sup>a</sup>

Depth, cm	Laboratory Number	Age Source	Dated Species	$^{14}\text{C}$ age	Calendar years	Sedimentation Rates, cm/kyr	Sampling Resolution, years
<i>CORE M39008 (Gulf of Cadiz)</i>							
8	KIA6969	$^{14}\text{C}$ AMS	<i>G. ruber</i>	2,660	2,332	5.1	975
18	KIA7639	$^{14}\text{C}$ AMS	<i>G. ruber</i>	4,195	4,282	10.4	479
28	KIA7640	$^{14}\text{C}$ AMS	<i>G. ruber</i>	4,875	5,240	9.4	530
38	KIA7641	$^{14}\text{C}$ AMS	<i>G. ruber</i>	5,910	6,301	28.0	178
91	KIA7642	$^{14}\text{C}$ AMS	<i>G. ruber</i>	7,755	8,193	133.4	37
190	KIA7643	$^{14}\text{C}$ AMS	<i>G. ruber</i>	8,445	8,935	323.4	15
288	KIA7644	$^{14}\text{C}$ AMS	<i>G. ruber</i>	8,755	9,238	51.0	98
358	KIA8131	$^{14}\text{C}$ AMS	<i>G. ruber</i>	9,860	10,610	22.3	224
403		<i>Bard et al.</i> [1987]	11,080 <sup>b</sup>	12,626	15.6	320	
458		<i>Bard et al.</i> [1987]	13,980 <sup>b</sup>	16,143	33.1	150	
529	KIA8581	$^{14}\text{C}$ AMS	<i>G. ruber</i>	15,800	18,285	11.8	422
564	KIA8132	$^{14}\text{C}$ AMS	<i>G. ruber</i>	18,370	21,243		
<i>CORE MD 95-2043 (Alboran Sea)</i>							
14		$^{14}\text{C}$ AMS	<i>G. bulloides</i>	1,980	1,527	27.0	186
54		$^{14}\text{C}$ AMS	<i>G. bulloides</i>	3,216	3,011	30.3	165
96		$^{14}\text{C}$ AMS	<i>G. bulloides</i>	4,275	4,396	50.6	99
178		$^{14}\text{C}$ AMS	<i>G. bulloides</i>	5,652	6,018	43.4	115
238		$^{14}\text{C}$ AMS	<i>G. bulloides</i>	6,870	7,401	38.1	131
298		$^{14}\text{C}$ AMS	<i>N. pachyderma</i>	8,530	8,976	58.8	85
348		$^{14}\text{C}$ AMS	<i>G. bulloides</i>	9,200	9,827	70.9	71
418		$^{14}\text{C}$ AMS	<i>N. pachyderma</i>	9,970	10,815	81.8	61
487		$^{14}\text{C}$ AMS	<i>N. pachyderma</i>	10,560	11,659	69.1	72
512		$^{14}\text{C}$ AMS	<i>N. pachyderma</i>	10,750	12,021	71.7	70
588		$^{14}\text{C}$ AMS	<i>N. pachyderma</i>	11,590	13,081	22.0	227
595		$^{14}\text{C}$ AMS	<i>G. bulloides</i>	11,880	13,399	97.8	51
682		$^{14}\text{C}$ AMS	<i>G. bulloides</i>	12,790	14,289	54.9	91
708		$^{14}\text{C}$ AMS	<i>N. pachyderma</i>	13,100	14,762	27.0	185
758		$^{14}\text{C}$ AMS	<i>N. pachyderma</i>	14,350	16,617	35.1	143
802		$^{14}\text{C}$ AMS	<i>N. pachyderma</i>	15,440	17,871	17.3	290
858		$^{14}\text{C}$ AMS	<i>N. pachyderma</i>	18,260	21,116	61.9	81
870		GISP2	-	-	21,310	11.4	438
894		GISP2	-	-	23,410	33.9	148
914		GISP2	-	-	24,000	35.1	142
940		GISP2	-	-	24,740	12.7	395
950		GISP2	-	-	25,530	25.8	194
984		GISP2	-	-	26,850	18.0	278
1000		GISP2	-	-	27,740	18.2	275
1010		GISP2	-	-	28,290	40.5	123
1040		GISP2	-	-	29,030	13.1	382
1054		GISP2	-	-	30,100	20.5	244
<i>CORE BS79-38 (Tyrrhenian Sea)</i>							
37		boundary 1-2		2,800 <sup>c</sup>	2,530	19.2	260
65		boundary 2-3		4,000 <sup>c</sup>	3,985	23.2	216
150		boundary 3-4		7,200 <sup>c</sup>	7,656	30.6	98
210		boundary 4-5		9,000 <sup>c</sup>	9,615	19.6	256
237		boundary 5-6		10,000 <sup>c</sup>	10,995	23.1	449
282		boundary 6-7		11,400 <sup>c</sup>	12,945	21.3	235
333		boundary 7-8		13,200 <sup>c</sup>	15,343	35.6	281
405		boundary 8-9		15,000 <sup>c</sup>	17,365	14.1	354
470		boundary 9-10		19,000 <sup>c</sup>	21,968		

Table 1. (continued)

Depth, cm	LaboratoryNumber	Age source	Dated Species	<sup>14</sup> C age	Calendar years	Sedimentation Rates, cm/kyr	Sampling Resolution, years
<i>CORE BS79-33 (Tyrrhenian Sea)</i>							
57		boundary 1-2	-	2,800 <sup>c</sup>	2,530	26.0	532
98		boundary 2-3	-	4,000 <sup>c</sup>	3,985	10.0	396
115		<sup>14</sup> C AMS	<i>G. rubber</i>	6,310	6,747	4.4	1,721
120		boundary 3-4	-	7,200 <sup>c</sup>	7,656	14.2	300
136		<sup>14</sup> C AMS	<i>G. rubber</i>	8,160	8,615	12.5	1,002
150		boundary 4-5	-	9,000 <sup>c</sup>	9,615	21.9	633
175		boundary 5-6	-	10,000 <sup>c</sup>	10,995	9.0	1,230
185		<sup>14</sup> C AMS	<i>G. bulloides</i>	10,830	12,225	17.9	189
204		boundary 6-7	-	11,400 <sup>c</sup>	12,945	19.0	551
225		<sup>14</sup> C AMS	<i>G. bulloides</i>	12,910	14,343	22.4	625
249		boundary 7-8	-	13,200 <sup>c</sup>	15,343	39.4	1,062
295		<sup>14</sup> C AMS	<i>G. bulloides</i>	15,480	17,917	41.0	859
340		<sup>14</sup> C AMS	<i>G. bulloides</i>	16,990	19,655	17.9	947
360		boundary 9-10	-	19,000 <sup>c</sup>	21,968	71.6	503
450		<sup>14</sup> C AMS	<i>G. bulloides</i>	24,120	27,608		

<sup>a</sup> See *Stuiver et al.* [1998] for Calib 4.1 program. Sedimentation rates and the average time resolution for the SST measurements are also indicated.

<sup>b</sup> Data are taken from *Bard et al.* [1987] without correction for reservoir effect.

<sup>c</sup> Data are taken from *Capotondi et al.* [1999] without correction for reservoir effect.

## 6. Discussion

The isotopic and thermal gradients between the Gulf of Cadiz and the western Mediterranean Sea remained throughout the last 20 kyr B.P.. These gradients changed in intensity but never in direction, supporting the hypothesis that the western Mediterranean Sea worked as a concentration basin along this full period [*Rohling and De Rijk*, 1999]. The largest differences are observed in the glacial periods, reflecting an increasing isolation of the Mediterranean Sea which operated as an amplifier of the prevailing cold and arid glacial conditions. The highest SST oscillations recorded in the Tyrrhenian cores may reflect the stronger continental influence of this area in contrast to the more oceanic regime of the Atlantic site.

### 6.1. Glacial Period

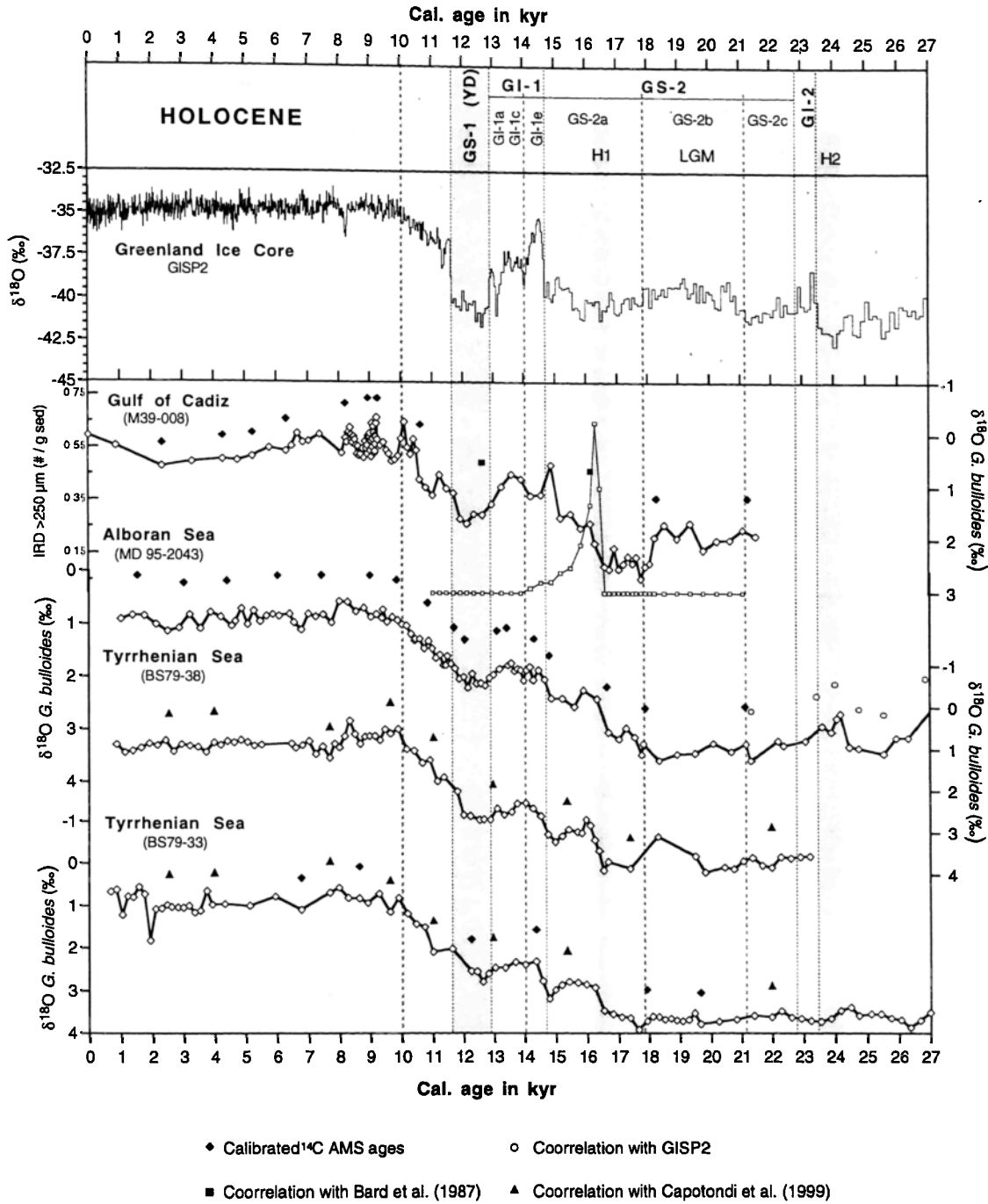
The SST profiles show relatively cold values for the glacial period, but the lowest values are not coincident with the Last Glacial Maximum (LGM) (18.5-20 kyr B.P.). These SST minima are confined to rapid events related with the Heinrich events (HE) as observed in core MD 95-2043 [*Cacho et al.*, 1999a]. H1 and H2 are well correlated in the Alboran and Tyrrhenian SST profiles, which highlights H2 as the coldest period for the last 30 kyr (Figure 4). Polar water entrance through the Strait of Gibraltar has been proposed as the main mechanism responsible for this water cooling in the Alboran [*Cacho et al.*, 1999a] and in the Tyrrhenian [*Paterne et al.*, 1999] Seas. This hypothesis is strongly supported now by the documentation of IRD occurrence during H1 in the Gulf of Cadiz (Figure 2). However, intensification of the Northern Hemisphere wind systems could also have played an important role in these Mediterranean water coolings [*Rohling et al.*, 1998; *Cacho et al.*, 1999a]. An atmospheric transmission of rapid climate change is supported by the oscillations in the relative abundance of tree pollen at Lago Grande di Monticchio and Lago di Mezzano (southern Italy) following the HE and Dansgaard-Oeschger (D-O) oscillations [*Watts et al.*, 1996; *Allen et al.*, 1999].

All the SST profiles depict relatively stable values during most of GS-2 (GS-2b and GS-2c) following an overall slightly warming trend. Such a stability is in strong contrast with the rapid SST variability documented for isotopic stage 3 in the Alboran Sea [*Cacho et al.*, 1999a]. A similar stable pattern is observed during isotopic stage 2 in North Atlantic mid latitude sites, as evidenced from foraminiferal-derived paleotemperatures (core SU90-03 at 40°N [*Chapman and Shackleton*, 1998] and core CH69-09 at 41°N [*Waelbroeck et al.*, 1998]) and alkenone SST (core MD 95-2037 at 37°N [*E. Calvo et al.*, New insights into the glacial latitudinal temperature gradients in the North Atlantic: Results from U<sup>K</sup><sub>37</sub>-sea surface temperatures and terrigenous inputs, submitted to *Earth and Planetary Science Letters*, 2000]). In contrast, the cores from northern latitudes show well-defined SST minima covering the whole GS-2 (core BOFS5K at ~51°N [*Maslin et al.*, 1995]) or exhibiting a clear cooling trend (core SU90-08 at 43°N [*Villanueva et al.*, 1998; *Paterne et al.*, 1999]). This latitudinal contrast has been discussed by *Chapman and Maslin* [1999] and was interpreted to result from seasonal insolation increase over low latitudes, which led to enhanced transport of atmospheric moisture. Higher atmospheric moisture stimulated Northern Hemisphere ice sheet growth during the LGM.

The results of the present study indicate that the equatorial SST warming extended into the Mediterranean region by the eastern boundary current from the Atlantic subtropical gyre. This situation would produce a strong meridional gradient between 40° and 50°N in the North Atlantic Ocean [*Chapman and Maslin*, 1999] in agreement with the LGM position of the polar front reconstructed by *Climate: Long-Range Investigation, Mapping and Precipitation (CLIMAP) Project Members* [1976].

### 6.2. First Phase of the Last Deglaciation

In all locations considered in the present study the last deglaciation warming started immediately after H1, prior to the onset of the late glacial interstadial (GI-1) in the chronostratigraphic control of Greenland ice cores (Figure 3 and



**Figure 2.** Profiles of *G. bulloides*  $\delta^{18}O$  from the studied cores and ice-rafted detritus (IRD) for core M39-008 compared with the  $\delta^{18}O$  record from the GISP2 Greenland ice core [Grootes et al., 1993]. The event chronology proposed by the Integration of Ice-core, Marine and Terrestrial records (IMITATE) group [Björck et al., 1998] (see Figure 3) is indicated. The points drawn over the  $\delta^{18}O$  curve of each core indicate the location of the dated points used in the age models (see Table 1).

4), the so-called Bølling transition. This warming phase is the most distinctive feature amongst the various basins studied. While in the Gulf of Cadiz, this warming occurred as a single and extremely rapid pulse at ~16 kyr B.P.; in the Tyrrhenian and Alboran Seas the warming followed a stepwise pattern. Thus all Mediterranean cores show colder SST at the beginning of the last glacial interstadial (GI-1e or Bølling period) than at the end (GI-1a to GI-1c or Allerød period), just the opposite of what is

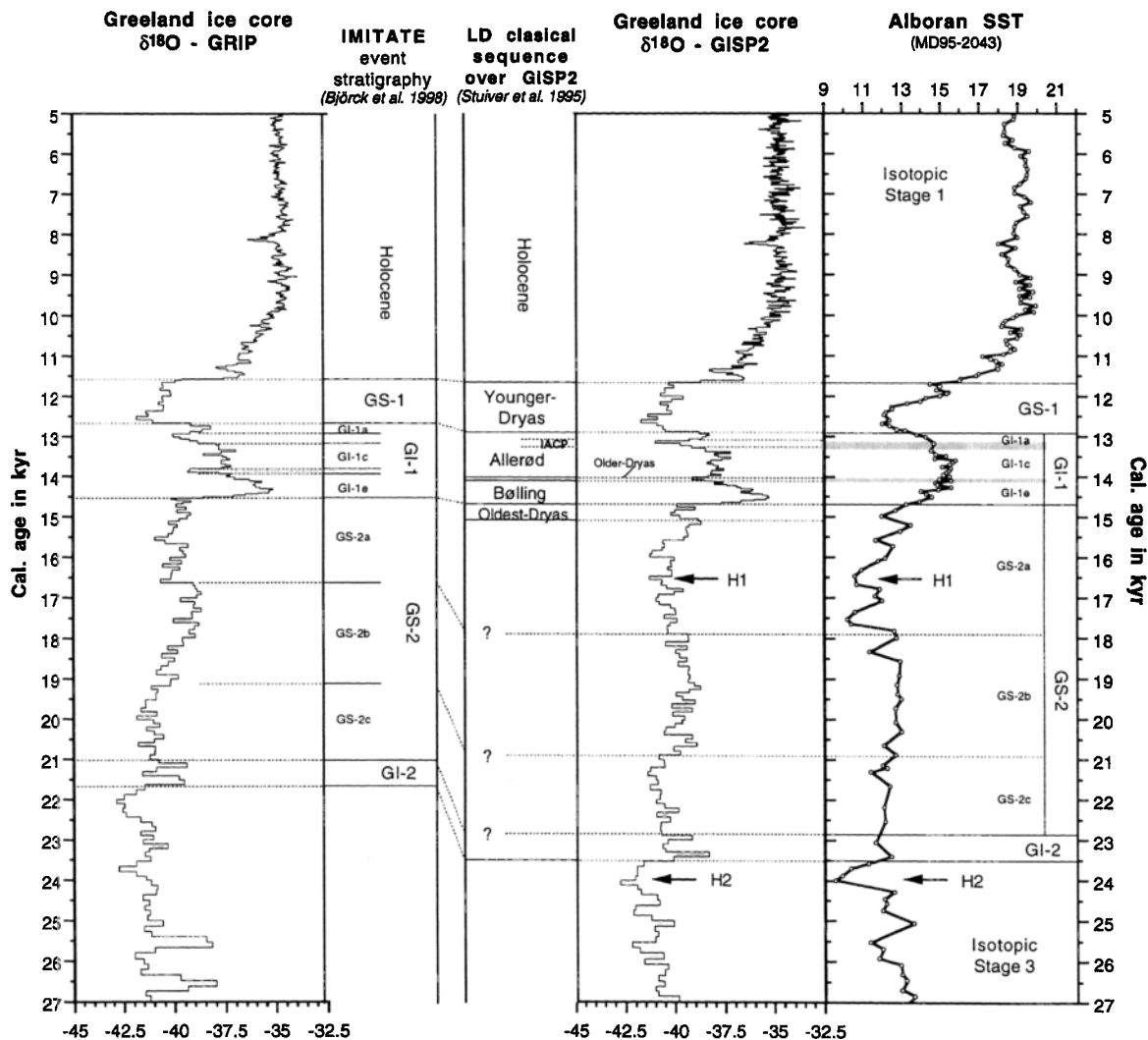
observed in most temperature records previously reported. Pollen, beetles, and also Greenland Ice Core records show that the initial warming in North Atlantic borelands occurred as a single and rapid pulse, reaching maximum values at the beginning of the Bølling episode followed by a cooling trend throughout the full Bølling-Allerød (B-A) period [Johnsen et al., 1992; Lowe et al., 1994; Coope and Lemdahl, 1995; Stuiver et al., 1995; Walker, 1995]. In contrast with the terrestrial records,

the B-A marine pattern varies from place to place. Foraminifera-derived SST from northwest Scotland closely parallel the B-A cooling trend of the terrestrial records [Haflidason *et al.*, 1995; Kroon *et al.*, 1997], while diatom SST reconstructions from the Norwegian Sea show an almost flat B-A period [Koç Karpuz and Jansen, 1992]. This marine heterogeneity points to a rather complex oceanographic response to the global warming related to the first phase of the TI.

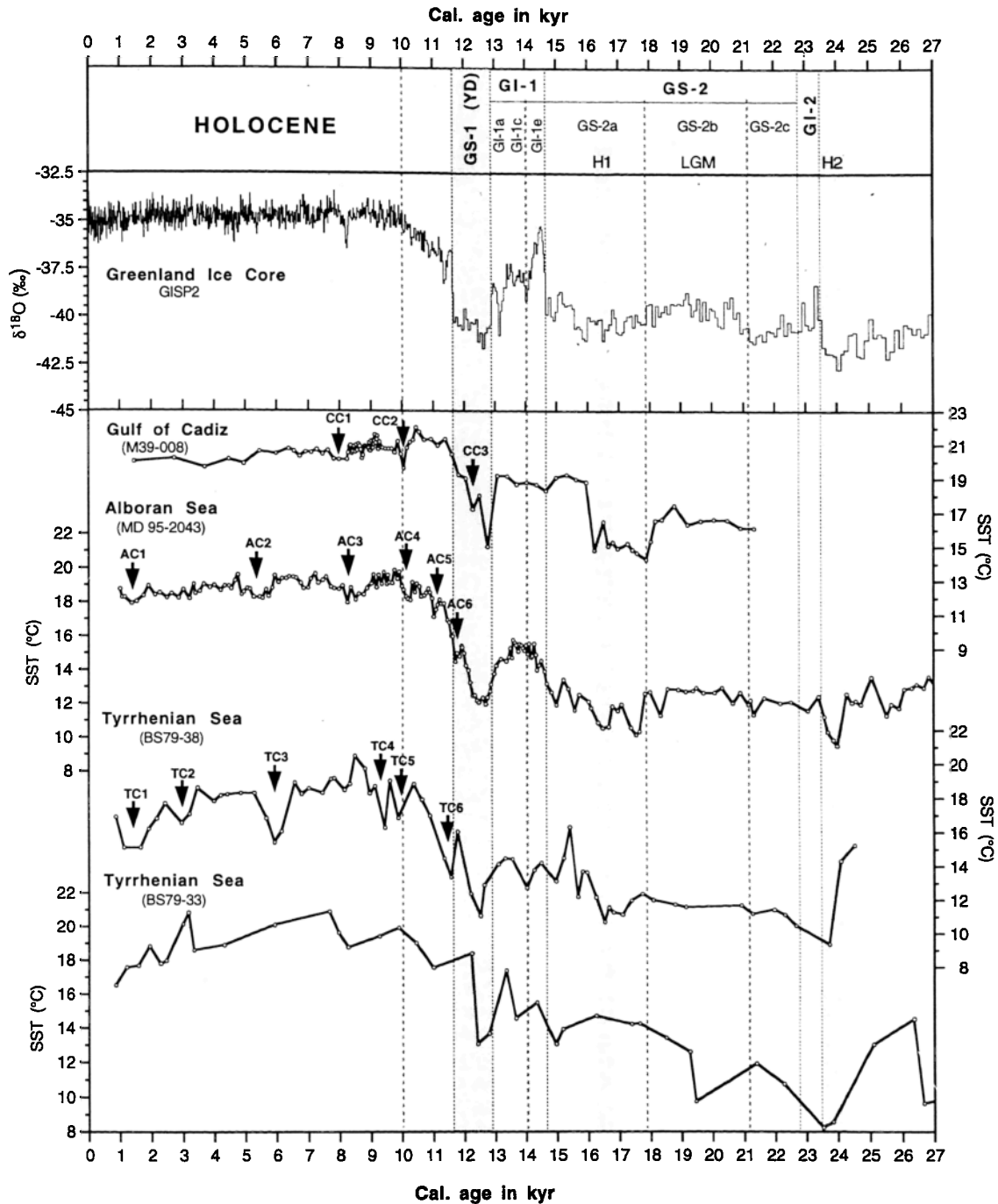
The sharpness of this warming phase (4°C per ~ 225 years) related to GI-1 in the Gulf of Cadiz has also been reported in the Iberian margin [Bard *et al.*, 1987]. It likely corresponds to the rapid northward migration of the polar front after the occurrence of H1, leading to the arrival of surface warm subtropical waters to the Gulf of Cadiz. Surprisingly, this abrupt change in the surface water mass source is not detected in the Alboran Sea (4°C per ~2000 years) nor in the Tyrrhenian Sea (4°C per ~3000 years) which are under the direct influence of the surface inflowing water from the Gulf of Cadiz (see section 2). Local processes, such as an intensification in the vertical water mixing

with colder intermediate waters, may have been the mechanism responsible for an intensive cooling of the warm inflowing waters, thus modifying the Atlantic signal. Enhanced water exchange during this period, probably related to sea level rise, has been previously documented by sedimentological and benthic fauna analyses in the Gulf of Cadiz [Caralp, 1988; Grousset *et al.*, 1988; Sierro *et al.*, 1999]. This vertical mixing should also drive a nutrient fertilization of surface waters, increasing primary productivity as is supported by studies on diatom assemblages [Abrantes, 1988] and planktonic foraminifera [Vergnaud-Grazzini and Pierre, 1991]. The warm B-A SST were only reached after ~14.5 kyr B.P. probably due to a weakening of the water mixing processes as is also suggested by the primary productivity proxies [Abrantes, 1988; Vergnaud-Grazzini and Pierre, 1991].

The SST profiles of the last deglaciation also document some short-term variability by the occurrence of some brief coolings which were more intense in the Tyrrhenian Sea. The  $\delta^{18}\text{O}$  profiles also report some short oscillations as the well



**Figure 3.** Correlation between  $\delta^{18}\text{O}$  from the two Greenland ice cores, GISP2 [Grootes *et al.*, 1993] and GRIP [Johnsen *et al.*, 1992; Dansgaard *et al.*, 1993; Grootes *et al.*, 1993], and the Alboran Sea alkenone sea surface temperature record from core MD 95-2043. Figure 3 shows the classical last deglaciation sequence [Mangerud *et al.*, 1974] indicated over GISP2  $\delta^{18}\text{O}$  profile by Stuiver *et al.* [1995] and its correspondence with the new event chronology proposed by the IMITATE group [Björck *et al.*, 1998].



**Figure 4.** Alkenone SST profiles compared to the  $\delta^{18}\text{O}$  record (top panel) of the Greenland ice core GISP2 [Groote et al., 1993]. The event chronology proposed by the IMITATE group [Björck et al., 1998] (see Figure 3) is shown. Arrows over SST profiles indicate the position of the cold Holocene events described in text and listed in Table 2.

documented enrichment at ~15 kyr B.P., which separates in two phases the isotopic depletion trend of this period (Figure 2). Paleoclimatic records from the North Atlantic Ocean borderlands have documented several minor, short-lived climatic oscillations as the Older Dryas (OD) and the Intra-Allerød Cold Period (IACP) (Figure 3) [Mangerud et al., 1974; Björck et al., 1996]. High-resolution studies from North Atlantic Ocean and Norwegian Sea cores have also reported the occurrence of few

cold spells during this period [Lehman and Keigwin, 1992; Koç Karpuz and Jansen, 1992; Kroon et al., 1997], which have been related to events of decreased flux of warm surface water into the high latitudes. Similar short-term variability has also been recognized in biological productivity records from the Cariaco basin (10°N) being associated to short periods of trade wind strengthening over the tropical North Atlantic, leading to the intensification of the upwelling cell [Hughen et al., 1996].



Mediterranean SST oscillations, especially the cooling recorded at ~15 kyr B.P. (the Oldest Dryas), which has been also reported in pollen records from Italian lakes [Huntley *et al.*, 1999], indicate that Mediterranean climate was also sensitive to such short-term variability. The atmospheric scenario described by Hughen *et al.* [1996] may also involve stronger northwesterlies responsible for short Mediterranean cooling events, similar to the teleconnection mechanisms hypothesized for the cold D-O stadials [Rohling *et al.*, 1998; Cacho *et al.*, 1999a]. Such atmospheric force would explain the higher SST variability of the Tyrrhenian Sea for this period, in contrast to the other studied areas, which were under higher oceanic influence.

### 6.3. Younger Dryas

The YD period or Greenland stadial I (GSI) is very well marked in all the studied profiles (Figs. 4), showing that SST cooled down (3°–4°C) to glacial values very abruptly, in a 50–80 years interval. This rapid cooling occurred in both sides of the Gibraltar Strait. Rapid SST cooling and salinity depletion during YD has been associated, in the Portuguese margin, to fast movements of the polar front driven by changes in the global thermohaline circulation [Bard *et al.*, 1987; Duplessy *et al.*, 1992]. If the polar front moved southward enough to reach the Strait of Gibraltar latitude, this could also be the mechanism responsible for the YD cooling recorded in our cores [Cacho *et al.*, 1999a]. Nevertheless, an atmospheric transmission for the YD SST cooling cannot be excluded. Model experiments suggest that atmospheric circulation played an important role in the hemispheric propagation of the YD cooling by intensification of the wind system in many regions and southward displacement of the winter storm tracks [Renssen *et al.*, 1996; Mikolajewicz *et al.*, 1997; Fawcett *et al.*, 1997]. Such YD atmospheric perturbations are also documented in Greenland ice cores by a decrease in snow accumulation [Alley *et al.*, 1993; Kapsner *et al.*, 1995] and an increase in the concentrations of dust and sea salt [Mayewski *et al.*, 1993; Taylor *et al.*, 1997].

The most surprising feature of SST evolution during the YD in all our profiles concerns to the brief duration of the cold phase (~700 years), which contrasts with extent of the isotopic plateau recorded in the same cores for this period (~1100 years). This means that the onset of the final YD warming started ~600 years earlier in the Mediterranean SST (12,250 years B.P.) than over Greenland (11,650 years B.P.). This warming phase was extremely abrupt in the Alboran Sea (3.3°C per 55 years) and was suddenly interrupted by a short (1°C) reversal which ended in parallel with the GS-1 on GISP2. Cores M39-008 (Gulf of Cadiz) and BS79-38 (Tyrrhenian Sea) also show a short reversal just after the YD warming although the lower resolution and/or age model uncertainties prevent a precise correlation of this extremely short event (260 years). The causes of these timing anomaly in our YD records could be regarded as a chronological problem in the framework of the Greenland Ice Sheet Project 2 (GISP2) ice core and/or the studied profiles.

The YD in Greenland ice records lasted 1200 years with slight differences in the absolute timing, 12,800–11,600 years B.P. in GISP2 and 12,700–11,500 years B.P. in Greenland Ice Core Project (GRIP) [Johnsen *et al.*, 1992; Grootes *et al.*, 1993]. This is not significantly different from <sup>230</sup>Th/<sup>234</sup>U dating of Barbados corals, 13,000–11,700 years B.P. [Fairbanks, 1990], or varved sediment counting estimations from European lakes, 12,700–11,500 years B.P. [Goslar *et al.*, 2000 and references therein]. Core 56/10/36 from northwest Scotland (~56°N) provides the presently available highest-resolution YD record from the North Atlantic Ocean and shows a good matching with GISP2 chronostratigraphy [Kroon *et al.*, 1997]. This correlation

is strongly supported by the dating of the Verde ash layer present in both cores. Annually laminated sediments from Cariaco basin (~10°N) also correlate with the GISP2 chronology for the YD [Hughen *et al.*, 1998]. Thus the timing and duration of the YD in Greenland seem to provide a real chronostratigraphy for climatic and oceanographic oscillations that occurred in the North Atlantic region.

Alboran core MD 95-2043 has a robust age model including two <sup>14</sup>C AMS ages for the YD termination. The YD in the Tyrrhenian core BS79-33 is also <sup>14</sup>C AMS dated. Calibration of <sup>14</sup>C AMS ages during the YD is particularly difficult for the occurrence of two <sup>14</sup>C plateaus, one of them just at the time of the YD termination. This last plateau has a duration of ~250 years and ends 100–200 years after the YD termination [Goslar *et al.*, 1995]. However, the potential calibration error is smaller than the observed differences in our study, and in any case, it would introduce a lag rather than a lead in our calibrated ages. Another potential error source consists of the ventilation age of the water masses from the studied area. Since no major change in the Mediterranean circulation pattern occurred and considering that the Mediterranean water renewal period is 100 years, ventilation ages would not differ strongly from those of the Atlantic Ocean, at least those from the Alboran Sea. These Atlantic Ocean ventilation rates were estimated by Bard *et al.* [1994] who found an atmosphere-sea surface <sup>14</sup>C difference of 700–800 years for the YD (a period of 400 years is already corrected in our age models [Stuiver *et al.*, 1998]).

Bioturbation is another potential factor for disturbance, but taking into account the extremely high sedimentation rates for this period in core MD 95-2043 (70–80 cm/kyr) and the results from previous simulated bioturbation models [Bard *et al.*, 1994], this effect could only be of a few years. On the other hand, the studied  $\delta^{18}\text{O}$  curves provide a clear proof, totally independent from the age model constraints, that there is a lead in the YD end warming of the studied cores. SST started to increase ~400 years before the  $\delta^{18}\text{O}$  depletion associated to the last phase of the deglaciation. Therefore the above reported SST rise must be considered as a real warming lead.

Considering that the YD cooling was driven by the entrance of cold waters through the Strait of Gibraltar, our timing would indicate a rapid northward retirement of the polar front in middle YD. A displacement of a few degrees north would be enough to avoid the entrance of cold waters through the Strait of Gibraltar. The short SST reversal in Alboran core during the YD termination would probably indicate a second brief phase of southward displacement of the polar front but less intense than the previous one. Distinct periods of polar water replacement by warm and salty waters have been identified in the YD profile from the North Atlantic core 56/10/36 [Kroon *et al.*, 1997]. Further work on high-resolution YD records from low North Atlantic latitudes (~35°N) would be required for a better assessment of the polar front movement during this period.

Several records from the northwestern European continent document the occurrence of climatic changes during the YD, pointing to a warmer episode during the latter phase [Walker, 1995]. Sediments from Lake Gosciadz (Poland) show two main phases: the earlier, which was longer, colder, and drier, and the latter, which was shorter and milder [Goslar *et al.*, 1993]. The transition between these two phases is assigned to ~600 years before the YD end. This two fold character of the YD has also been observed recently in a laminated record from Meerfelder Maar (western Germany), where this change toward more humid conditions is dated at 12,250 kyr B.P. [Brauer *et al.*, 2000]. This timing is synchronous with the age of our premature warming in the Alboran Sea. Greenland ice cores also show an intra-YD

weak oscillation around this age in the  $\delta^{18}\text{O}$  record, which is better defined in the sea salt records [Grootes *et al.*, 1993; Mayewski *et al.*, 1993]. Farther south a well  $^{14}\text{C}$  AMS dated pollen record from the Iberian Peninsula also shows that the final warming of the YD advanced, in ~600 years, the SST North Atlantic warming recorded in core SU81-18 [Peñalba *et al.*, 1997]. This warming phase was shortly interrupted by a cooling period similar to that recorded by SST in Alboran Sea. All this evidence strongly supports that this climatic variation over Europe was related to the YD signature in our cores and therefore that wind regime changes were involved in this early warming phase. Comparison of the diverse YD records mentioned above seem to indicate that this climatic disturbance increased its amplitude toward the south.

6.4. Holocene

All studied areas show the warmest SST values at the early Holocene followed by a slow cooling trend. Nevertheless, the

timing of the Holocene maximum SST change between the areas: 11.5-10.2 kyr B.P. in the Gulf of Cadiz, 10-9 kyr B.P. in the Alboran Sea, and 8.9-8.4 kyr B.P. in the Tyrrhenian Sea. This indicates a regional diachroneity in the occurrence of the Holocene optimum, which propagates from west to east.

The general Holocene cooling trend is shortly interrupted by the occurrence of some short cooling events (1°-2°C), which are more intense in the Tyrrhenian Sea (2.5°-3°C). This short-term variability can be better analyzed in core MD 95-2043, the one studied at higher resolution and with more abundant  $^{14}\text{C}$  AMS ages. Six short cold events (AC1-AC6; 1°-1.5°C) have been identified (Figure 4 and Table 2). The younger events (AC1-AC3) lasted 900-1500 years, while the older events (AC4-AC6) were considerably shorter (250-400 years) but with the same thermal intensity (1.5°C).

A periodicity of ~730±40 years is observed between the oldest cold spells (AC6-4; see Table 2). Then, these periods lengthen after the Holocene optimum, but they always occur in harmonic tones of this 730±40 years periodicity (AC3: ~730 x 3; AC2:

**Table 2.** List of the Rapid Holocene Cooling Events Recorded in Each of the Studied Areas, Indicating Their Age, Duration, and the Time Period Between Them and List of the Holocene Cold Events Defined in the North Atlantic Ocean Proposed for Correlation With Those Described in the Present Study<sup>a</sup>

Cold Event	Minimum SST Age, kyr	Time Span Since the Previous One, kyr	Interval Age	Duration, kyr	
<i>Alboran Sea</i>					
AC1	1.38	3.98	1.01 - 1.90	0.89	
AC2	5.36	2.88	4.75 - 5.94	1.19	
AC3	8.24	2.04	7.56 - 9.08	1.52	
AC4	10.28	0.73	9.95 - 10.34	0.39	
AC5	11.01	0.69	10.95 - 11.21	0.25	
AC6	11.70	0.77	11.65 - 11.91	0.26	
ACYD	12.47	-	12.00 - 13.10	1.10	
<i>Tyrrhenian Sea</i>					
TC1	1.39	1.56	1.00 - 1.91	0.91	
TC2	2.95	2.98	2.5 - 3.45	0.95	
TC3	5.93	3.52	5.28 - 6.58	1.3	
TC4	9.45	0.42	9.13 - 9.62	0.49	
TC5	9.87	1.69	9.62 - 10.38	0.76	
TC6	11.56	0.95	10.9 - 11.78	0.88	
TYD	12.51	-	12.00 - 13.09	1.09	
<i>Gulf of Cadiz</i>					
CC1	7.98	2.02	7.8 - 8.25	0.45	
CC2	10	2.27	9.9 - 10.2	0.3	
CC3	12.27	0.48	12.2-12.4	0.2	
CYD	12.75	-	12.6-12.9	0.3	
North Atlantic Cold Events	Minimum SST Age, kyr	Time Span Since the Previous One, kyr	Alboran cold events	Tyrrhenian Cold Events	Gulf of Cadiz Cold Events
NAC1	1.4	1.4	AC1	TC1	-
NAC2	2.8	1.5	-	TC2	-
NAC3	4.3	1.6	-	-	-
NAC4	5.9	2.3	AC2	TC3	-
NAC5	8.2	1.3	AC3	-	CC1
NAC6	9.5	0.8	-	TC4	-
NAC7	10.3	0.8	AC4	TC5	CC2
NAC8	11.1	1.4	AC5	-	-
-	-	-	AC6	TC6	CC3
NAYD	12.5	-	AYD	TYD	CYD

<sup>a</sup> See text for list of cooling events in the studied areas, and see Bond *et al.* [1997] for cold events in the North Atlantic Ocean.

~730 x 4; AC1: ~730 x 5). The cooling events observed in the other studied cores are generally coincident with the events in MD 95-2043 (Table 2), although the lower resolution and less accurate age model do not allow a strict correlation between them. Only few of them, AC3 and AC4 in the Alboran Sea and CC1 and CC2 in the Gulf of Cadiz, can be well correlated (Figure 2 and Table 2), and they are coincident in age with the pre-Boreal oscillation and the Greenland 8.2 event, respectively. These cooling events have been abundantly reported in the North Atlantic as being associated with short freshwater influxes [Björck et al., 1996; Alley et al., 1997, and references therein]. However, in the context of the Mediterranean Holocene SST records they do not show remarkably different characteristics from the other coolings, constituting the sequence with the 730 years periodicity pattern.

The Holocene record in Greenland shows four short periods of increased concentrations of sea salt and terrestrial dust [O'Brien et al., 1995]. Such atmospheric perturbations have been associated to a group of SST cooling events (2°C) in the North Atlantic Ocean [Bond et al., 1997]. A cyclicity of 1470±500 years has been related to them and also seems to be related to changes in the deep-ocean flow south of Iceland [Bianchi and McCave, 1999] and climatic variability over the North American continent [Campbell et al., 1998]. The exact timing of the North Atlantic coolings [Bond et al., 1997] is listed in Table 2, where they are tentatively correlated with most of the cold events found in our cores. A periodicity of ~730±40 years in the earlier Holocene can also be observed in the Atlantic coolings. This periodicity involves  $n \times 730$  years harmonics in the late stage. Such cyclicity (~750 years) has also been reported in Holocene records from Asian monsoon-driven sea surface salinity oscillations in the South China Sea [Wang et al., 1999] and Indian monsoon rainfall from the Arabian Sea [Sakar et al., 2000].

All this evidence indicates that this short-term Holocene variability was a wide extended process. Our data suggest that at least part of this cooling was transmitted through the inflowing North Atlantic Water. Nevertheless, the larger Thyrrenian amplitude for this Holocene oscillation indicates a Mediterranean amplification, probably induced by changes in the wind regime. Similar short Holocene cooling events have been previously documented in some remote areas from the eastern Mediterranean Sea, the Aegean and Adriatic Seas [Rohling et al., 1997; De Rijk et al., 1999]. These areas are strongly influenced by the flow of winter continental winds, which could play an important role for these brief coolings. Further work is necessary for absolute correlation between Mediterranean and Atlantic coolings and also for full understanding of the sequence of processes involved in their transmission. Nevertheless, it is interesting to point out that although this variability involved only short thermal changes, it could be related to drastic changes in the deep water convection of the eastern and western

Mediterranean basins [De Rijk et al., 1999; Myers and Rohling, 2000].

## 7. Conclusions

Both long- and short-term variability in the North Atlantic modulated the climatic evolution of the Mediterranean SST over the last 25 kyr. Atmospheric circulation also played a major role during some periods, which currently led to amplifications of the Atlantic signal. The SST and  $\delta^{18}\text{O}$  gradients between the Gulf of Cadiz and the Alboran Sea changed in intensity but never in direction, pointing to the operation of the western Mediterranean in concentration basin conditions all along this period.

SST minima are not recorded during LGM but during H1 and H2. Occurrence of an IRD layer in the Gulf of Cadiz in coincidence with H1 supports the hypothesis that polar water entered in the Mediterranean Sea through the Strait of Gibraltar during the HE. In the first phase of the deglaciation a slow down of the Atlantic Ocean warming signal is observed in the Mediterranean Sea. This attenuation could reflect an intensification of surface mixing processes in this area.

SST warming after the YD occurred ~600 years earlier than the warming described in other marine and continental areas. This advancement is related to the occurrence of two different climatic phases during this period which had a stronger effect in southern Europe. Minor oscillations in the Atlantic thermohaline circulation could be sufficient to avoid the entrance of cold waters through the Strait of Gibraltar and also determine this earlier Mediterranean warming.

Holocene SST show a cooling trend from start to present in all the studied areas and reflect a regional diachroneity, occurring earlier (10-11.5 kyr B.P.) in the areas under higher oceanic influence and later (8.4-8.9 kyr B.P.) in the enclosed basin sites. A sequence of short SST cooling events (1.5°-3°C) observed in the cores studied at high resolution is consistent with a climatic periodicity of ~730±40 years. These coolings are in agreement with the Holocene variability recorded in other areas. They were transmitted to the Mediterranean Sea by the Atlantic inflowing water, but strong winter winds may have amplified these oscillations in the Tyrrhenian Sea.

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