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# Paleoproductivity variations related to climatic conditions in the Alboran Sea (western Mediterranean) during the last glacial-interglacial transition: the diatom record

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#### Abstract

The present study addresses the analysis of changes in marine primary productivity derived from climatic variability over the last 30,000 yr. The study area is the Alboran Sea; this is the westernmost basin of the Mediterranean Sea, and was a very sensitive area to changes in sea levels and atmospheric configuration during the last glacial-interglacial transition. We report the results from four sediment cores recovered in both the western and eastern Alboran basins. This allowed us to monitor the W-E evolution of the hydrological structures. The study is mainly based on the diatom record, although total organic carbon (TOC), isotopes and UK'37-SST were also used to gain further information. Paleoceanographic conditions favoured high paleoproductivity rates during the Last Glacial Maximum, although this occurred at Termination 1a and during the Younger Dryas event (YD), increasing towards the easternmost sites. During these periods, meteorological conditions, with increased westerlies, intensified the flow to the east, driving an eastward migration of the Western Anticyclonic Gyre. A possible displacement of the North Alboran upwelling system is also considered. The major component of the diatom assemblage during the YD are the resting spores of Leptocylindrus danicus, which suggest the presence of cooler and less saline surface waters. The opposite behaviour between planktonic and benthic proxies in addition to the high TOC values allows us to infer a phase of reduced deep water renewal between 16.5 and 9 cal ky BP. Fresh-water diatoms and opal phytoliths were used as indicators of aridity and humid land-conditions, respectively. During the glacial period and until 14 cal ky BP we observed a gradual climatic deterioration, accompanied by a loss of grass belts. Climatic conditions were warmer and moister during the Bølling-Allerød while the YD was dry. The re-establishment of vegetation belts and replenishment of lakes was observed at 8 cal ky BP. © 2001 Elsevier Science B.V. All rights reserved.

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# 1. Introduction

The present-day climate of the Mediterranean Sea

is characterised by a very stable high-pressure system that is related to the Intertropical Convergence Zone (ITCZ) (Cramp and O'Sullivan, 1999). Its position is critical in determining the climatic and oceanographic patterns in the present Alboran Sea (Parrilla and Kinder, 1987). During glacial periods, the ITCZ

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migrated to the south (Sarnthein et al., 1982), favouring the location of a low-pressure system over the Mediterranean and prevailing westerlies, and therefore favouring "red snow" over the Mediterranean (Parrilla and Kinder, 1987).

Changes in paleoproductivity related to variations in climatic and paleoceanographic conditions have been widely studied in the area (Abrantes, 1988a; Caralp, 1988; Weaver and Pujol, 1988; Turon and Londeix, 1988; Vergnaud-Grazzini and Pierre, 1991; Targarona et al., 1997a; Sierro et al., 1998). Previous studies by Abrantes (1988a) and Bárcena and Abrantes (1998) established the connection between diatom occurrence in Alboran Sea sediments with upwelling conditions in the overlying waters, and downcore variations in diatom abundance as a result of paleoproductivity changes (Abrantes, 1988a). In the above studies, strong changes in primary productivity related to the last glacial-interglacial transition were described, although controversial results arise when different proxies are compared. An organicrich layer (ORL) has been described in Alboran sediments from the deglaciation, indicating diachronous deposition with the last sapropel (S1) in the eastern Mediterranean (Sierro et al., 1998; Murat, 1999). The origin and inducing factors of this ORL have been addressed previously by Caralp (1988) and Sierro et al. (1998) but several points, especially concerning surface productivity, still remain unresolved.

The present work aims at gaining further insight into the climatic and paleoceanographic changes that have occurred over the past 30 ky in the Alboran Sea. For this approach, four sediment cores are compared using a common chronostratigraphical framework. One of the cores studied, MD 95-2043, has a very high resolution and its value for monitoring rapid climatic variability, such as the Greenland Dansgaard-Oeschger (D-O) cycles (Messe et al., 1997), has been demonstrated previously by Cacho et al. (1999). Regional atmospheric patterns would have induced paleoceanographic and paleoproductivity changes in Alboran high-fertility systems. Moreover, Northern Hemisphere climatic and oceanographic changes would also have affected the Alboran Sea by introducing fresh waters through the Strait of Gibraltar. Climatic changes in North Africa have been studied on the basis of continental microfossils (Bárcena et al., 1997; Flores et al., 2000).

#### 2. Oceanographic aspects of the Alboran Sea

The Mediterranean behaves as a system of antiestuarine circulation, the surface water entering from the Atlantic and leaving the Mediterranean at depth (Bormans et al., 1986). The general circulation has a thermohaline origin due to an excess of evaporation over precipitation and is controlled by the exchange of water through the Strait of Gibraltar.

The Alboran Sea is the westernmost basin of the Mediterranean. The hydrography in this area can be considered as a two-layer system: the salty and dense mediterranean outflow water (MOW), which is a mixture of Levantine intermediate water (LIW), West Mediterranean deep water (WMDW) and the lighter Atlantic surface water (ASW). The circulation of the Alboran Sea is energetic, and is subject to strong seasonal variations related to fluctuations in the intensity of water exchange through the Strait of Gibraltar (Pistek et al., 1985). Unlike most of the Mediterranean, the Alboran Sea has two systems of high biological productivity which are associated with the Western Anticyclonic Gyre (WAG) and the Almeria-Oran density front. To the northern limb of the WAG, an upwelling of subsurface waters occurs, and hence a frontal system develops along the eastern gyre (Fig. 1). The vertical movement that occurs in the WAG does not seem to be linked to regional atmospheric dynamics (La Violette, 1984). It is probably the result of an instability created by the lateral contact between the ASW and the MOW in addition to the topography of the sea floor (Perkins et al., 1990). The deep, nutrient-enriched waters rise following the continental slope (Ballester and Zavatti, 1983) and then fertilise surface levels. The area also receives an additional contribution of nutrients due to partial enrichment of the ASW as it passes through the Gulf of Cadiz and the Strait of Gibraltar (Minas et al., 1984: Parrilla and Kinder, 1987).

Neither the position nor the behaviour of the gyres is stable; satellite images reveal that the gyres can switch direction and even disappear, and several scenarios have been described (Tintoré et al., 1988; Heburn and La Violette, 1990; Perkins et al., 1990). One of the most significant scenarios shows how the eastern gyre disappears while the WAG remains. Since this scenario occurs at the end of the year when exchange through the Strait of Gibraltar is



Fig. 1. Study area. Location of the four cores studied and of ODP hole 977A. Predominant oceanographic features of the Alboran Sea, the WAG and the Almeria-Oran Front. Shaded areas represent today's high fertility zones. Summer and winter positions of the ITCZ are indicated.

weak due to partial draining of the WMDW, it has been related to thermohaline forcing (Bormans and Garret, 1989). Another example of an extreme scenario is the persistence of the eastern gyre, which is related to meteorological forcing (Parrilla, 1984; Perkins et al. 1990). Moreover, the stability of given circulation patterns in the Alboran Sea over an extended time period must be related to long-term changes in sea-level or the intensity of thermohaline circulation, which is associated with the basin's excess of evaporation. (Targarona et al., 1997a).

#### 3. Materials and methods

The present work includes the study of three gravity cores and one calypso piston core recovered in the Alboran Sea (Fig. 1). Cores KS8230 and KS8231 were taken during the French expedition FAEGAS IV at  $36^{\circ}27'N/3^{\circ}53'W$  and  $36^{\circ}09'N/3^{\circ}16'W$ , at a water depth of 795 and 855 m, respectively. Core TG-5 was recovered during the GC-90-1 Spanish leg at  $36^{\circ}23'N/4^{\circ}15'W$  at a water depth of 626 m. Finally, calypso piston core MD 95-2043 was taken during the 1995 IMAGES cruise at  $36^{\circ}09'N/2^{\circ}37.269'W$  and 1841 m water depth.

## 3.1. Micropaleontological studies

Micropaleontological studies were carried out on samples spaced at 10 cm intervals in core MD 95-2043. The sampling intervals for TG-5 were variable due to the small amount of sediment available. For cores KS8230 and KS8231 we used prior countings published by Abrantes (1988a) together with new countings made at intermediate samples, which yielded 10 cm sampling intervals.

For diatom analyses, samples were prepared according to the randomly distributed microfossils method outlined in Bárcena and Abrantes (1998). For each of the samples used in the micropaleontological study, the initial dry weight, the suspension volume and the volume used to mount the slides were known. Three slides per sample were used in each counting, 5–6 transects per slide were studied, and in general more than 400 fields of view were observed. Magnification was × 1000, and the recommendations of Schrader and Gersonde (1978) were used as a basis for the counting of diatom valves. Number of valves counted for species distribution was higher than 100.

## 3.2. Sea surface temperature assessment

Sea surface temperature (SST) was obtained from the relative composition of  $C_{37}$  unsaturated alkenones (UK'37 index; Brasell et al., 1986 using the calibration equation of Müller et al., 1998 (UK'37 = 0.033 SST + 0.044).

In core TG-5, samples were taken at different core depths; a complete description is offered in Cacho (1995). Alkenones were analysed using a Carlo-Erba model Mega Series 5160 Gas chromatograph equipped with a split-splitless injector, with a 50 m × 0.32 mm i.d. CP-Sil5-Ultra1 glass capillary column coated with a film of silica gel (100% dimethylpolysiloxane, thickness 0.17 µm) and flame ionisation detection. The oven temperature programme was from 150 to 300°C at 10°C/min, and from 300 to 330°C at 2°C/min. The final temperature was maintained for 18 min.

In core MD 95-2043, samples were taken every 4 cm. The alkenones were analysed with a Varian

gas chromatograph Model 3400 equipped with a programmable septum injector and a flame ionisation detector. The instrument was equipped with a CPSIL-5 CB column coated with 100% dimethylsiloxane (film thickness, 0.12 m). Hydrogen was used as carrier gas (50 cm/s). 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).

The gas chromatographs used in the measurement of each core had different injection systems, and there was no calibration between them.

#### 3.3. Total organic carbon measurements

Sample depths for total organic carbon (TOC) are shown in Tables 2–4. Samples for TOC measurements were first homogenised and decarbonated by adding 10 $\mu$ l of HCl (1 M) until mineral carbon has been completely removed. TOC was determined with a Carlo-Erba (NA 1500) Elemental Analyser. Combustion of the decarbonated samples was performed in tin foil at 1020°C under a helium flow (120 ml/min) using V<sub>2</sub>O<sub>5</sub> as catalyst. A standard (atropine) and a blank was combusted every 12 samples to quantify carbon concentrations.

# 4. Results

## 4.1. Chronological assessment

In order to establish a stratigraphic framework for the cores, constant sedimentation rates were assumed between age control points, and linear interpolation was used to obtain ages between the control points. Age models of cores KS8230 and KS8231 were constructed from a combination of <sup>14</sup>C dates, <sup>18</sup>O dates, and foraminiferal data as reported by Pujol and Vergnaud-Grazzini (1989) (Table 1). The age model in core TG-5 was based on four <sup>14</sup>C dates (Table 1). The age model of MD 95-2043 was based on 17 AMS-<sup>14</sup>C measurements on monospecific foraminiferal samples (Table 1).

To correct natural variations and to improve time resolution, <sup>14</sup>C dates were calibrated to calendar years

Table 1 Age control points for each core

. age BP)
357
\$38
SO2
/50
556
187
515
332
365
)50
487
515
502
332
000
365
)90
350
128
527
)11
396
)18
)41
976
327
315
502
)21
)81
399
289
762
517
371
116

using the CALIB 4.1 program (Stuiver and Reimer, 1993). The approximate time recorded in our cores is 20 ky for KS8230, 20.5 ky for KS8231, 23 ky for TG-5 and 26.5 ky for MD 95-2043.

Sedimentation rates varied among the different cores. The highest sedimentation rates were obtained



Fig. 2. Age model and sedimentation rates. Age models were constructed from a combination of <sup>14</sup>C dates, <sup>18</sup>O dates and foraminiferal data (see text and Table 1). Models were constructed by testing and comparing SST, TOC and diatom abundance data. For sedimentation rates we have assumed that these are constant between the dated points. <sup>14</sup>C dates were calibrated to calendar years with the CALIB 4.1 program (Stuiver and Reimer, 1993).

in core MD 95-2043, with an average value of 35.85 cm/ky for the time interval studied. The average sedimentation rates for cores KS8230 and KS8231 were 17.2 and 14.4 cm/ky, respectively. Core TG-5 had the lowest sedimentation rate, with an average value of 8.85 cm/ky (Fig. 2).

# 4.2. SSTs

The results of all parameters measured on the cores (UK'37 index, diatoms and TOC) are showed in Figs. 3–6.

SST derived from the unsaturation index UK'37 were measured in two cores, TG-5 and MD 95-2043 (Figs. 3 and 6). Both cores showed the same general trend: a warming of about 9°C for the last deglaciation distributed in two phases; a maximum SST in the early Holocene ( $\sim$ 10 ky BP) and a stabilisation of temperatures along the Holocene. The main differences between the two SST records lie in the absolute

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Fig. 4. <sup>18</sup>O dates (after Pujol and Vergnaud-Grazzini, 1989), TOC and diatom valves per gram of dry sediment for core KS8230. All ages are given in calendar ky BP after calibration with the CALIB 4.1 program. See also explanation in Fig. 3.

values. Core TG-5 always shows values  $1-2^{\circ}$ C warmer than core MD 95-2043. The different subbasin positions of the cores could explain this difference in SST values. However, this divergence can probably be accounted for in terms of analytical considerations: since there was no calibration between the gas chromatographs used in the measurement of each core, this cannot be excluded as an important factor. Furthermore, alkenone concentrations were very low in core TG-5, especially for the Holocene period, and this can also introduce important deviations in SST estimations (Villanueva and Grimalt, 1997).

The glacial SSTs from core MD 95-2043 have already been discussed in a previous study (Cacho et al., 1999) reporting the strong correlation of this record with the  $\delta^{18}$ O from the GISP2 ice core (Messe et al., 1997). Heinrich events are also well defined in this core due to the concurrence of both rapid SST drops and high concentrations of *Neogloboquadrina pachyderma* (s). For the time period addressed in the present study (30 ky) H1 and H2 can be clearly observed by SST drops of about 2.5°C (Fig. 6).

#### 4.3. Diatoms

Diatoms were present in all cores. The range of variations in diatom numbers fluctuated from an absence of diatom valves to an abundance of  $3.9 \times 10^6$  valves/g of dry sediment (Tables 2–5) in core MD 95-2043. In the Alboran Basin, abundances



Fig. 5. <sup>18</sup>O dates (after Pujol and Vergnaud-Grazzini, 1989), TOC and diatom valves per gram of dry sediment for core KS8231. All ages are given in calendar ky BP after calibration with the CALIB 4.1 program. See also explanation in Fig. 3.



Fig. 6. SST, TOC and diatom valves per gram of dry sediment for core MD 95-2043. Lower right: relative abundance of the most significant diatom taxa for the YD interval. H2:

Table 2

Sampling depth, calendar age, estimated SST (°C), TOC content, diatoms/mm<sup>2</sup> and absolute abundance of diatom valves of core TG-5

Depth (cm)	Cal. age $(\times 10^3)$	TOC (%)	Depth (cm)	Cal. age $(\times 10^3)$	TOC (%)	Depth (cm)	Cal. age $(\times 10^3)$	SST (°C)	Depth (cm)	Cal. age $(\times 10^3)$	Valves/ mm <sup>2</sup>	Valves/g $(\times 10^4)$
6	0.58	0.87	167	18.31	0.70	3	0.29	17.4	0	0.00		Frag
14	1.36	0.86	169	18.56	0.78	6	0.58	18.9	14	1.36	0.08	2.44
19	1.92	0.87	170	18.69	0.83	8	0.78	17.7	29	3.04		Frag
29	3.04	0.91	172	18.94	0.82	14	1.36	17.9	34	3.60		Frag
34	3.60	0.79	175	19.32	0.74	19	1.92	16.4	52	5.62		Frag
37	3.94	0.77	177	19.58	0.90	29	3.04		59	6.41		Frag
49	5.29	0.90	180	19.96	0.90	34	3.60		64	6.97		Frag
52	5.62	0.88	183	20.34	0.64	52	5.62	18.7	74	8.09		Frag
54	5.85	0.93	187	20.85	0.71	59	6.41		84	9.18		Frag
57	6.18	0.94	191	21.35	0.73	64	6.97	19.1	94	10.27	0.06	0.79
59	6.41	0.81	193	21.61	0.76	69	7.53	25.0	104	11.36	0.05	0.75
62	6.75	0.79	196	21.99	0.73	74	8.09	19.2	114	12.45	0.06	0.89
64	6.97	0.83	199	22.37	0.70	79	8.63	18.9	127	13.87	0.08	1.14
67	7.31	0.75	202	22.75	0.61	84	9.18	20.3	137	14.96	0.19	2.68
69	7.53	0.83	204	23.00	0.68	89	9.72	20.8	148	16.16	0.06	0.89
72	7.87	1.23	206	23.26	0.70	94	10.27	21.5	163	17.80	0.03	0.45
74	8.09	0.97	208	23.51	0.73	97	10.60	21.1	170	18.69	0.03	0.45
79	8.63	0.71				104	11.36	18.8	180	19.96	0.17	2.33
81	8.85	0.81				109	11.91	15.5	193	21.61	0.06	0.89
84	9.18	0.86				114	12.45	17.5	199	22.37	0.39	9.00
89	9.72	0.95				122	13.33	16.1	208	23.51	0.27	3.80
92	10.05	1.07				127	13.87	16.7				
94	10.27	0.95				137	14.96	14.2				
97	10.60	0.80				142	15.51	12.4				
99	10.82	0.89				152	16.60	14.1				
104	11.36	0.82				154	16.82	12.7				
109	11.91	1.09				160	17.47	13.7				
114	12.45	1.05				163	17.80	12.3				
122	13.33	1.20				169	18.56	12.4				
127	13.87	1.17				175	19.32	12.1				
133	14.53	1.12				180	19.96	12.3				
137	14.96	0.99				183	20.34	16.1				
142	15.51	0.91				185	20.59	15.3				
152	16.60	0.79				191	21.35	12.4				
154	16.82	0.82				199	22.37	11.2				
160	17.47	0.67				204	23.00	12.8				
163	17.80	0.80				208	23.51	14.5				

were lower than those normally observed in strong upwelling areas, where the number of diatom valves in the sediments may reach several millions. Nevertheless, the quantitative variations are similar in all cores (Figs. 3–6).

During marine isotopic stage 2 (MIS2) diatoms are present in all four Alboran cores. MD 95-2043 has the highest values and from the MIS3/MIS2 boundary to Termination 1a (T1a) several peaks can be seen (Fig. 6). This feature was also observed in the other cores, and allows a good correlation. The small differences observed in core correlations could fall within the error of the <sup>14</sup>C chronology. Additionally, the dating methods used could also induce chronological errors, and the fact that we assumed constant sedimentation rates between the points dated, which may not necessarily be true, could also account for the differences in age misalignments.

During deglaciation, diatom abundances show strong fluctuations. A diatom peak is recorded in cores TG-5, KS8230, KS8231 and MD 95-2043 during T1a and immediately after (Figs. 3–6). The

Table 4

Table 3

Sampling depth, calendar age, TOC content, diatoms/mm<sup>2</sup> and absolute abundance of diatom valves of core KS8230. For diatom countings, see also, Abrantes (1988a,b)

Depth	Cal. age	TOC	Valves/	Valves/g
(cm)	$(\times 10^{3})$	(%)	$\mathrm{mm}^2$	$(\times 10^4)$
0	0.00	0.75		5.60
2	0.09	0.77	0.53	3.78
10	0.44		0.46	3.27
50	2.75			Frag
100	6.13	0.84		Frag
120	7.49			Frag
150	9.00			Frag
180	11.83			Frag
200	12.84	0.93		Frag
208	13.24	0.81	0.13	0.95
220	13.84			5.60
231	14.40	0.66	0.46	3.27
238	14.75	0.57	0.25	1.79
250	15.35			6.10
258	15.76	0.50	0.59	4.17
268	16.26	0.59	0.58	4.07
278	16.76	0.50	0.53	3.77
290	17.36			5.00
294	17.57	0.50	0.35	2.48
300	17.87			8.40
310	18.37	0.56	0.36	2.52
320	18.87			3.00
330	19.38	0.51	0.72	5.06
340	19.88			1.20
350	20.38	0.64	0.37	2.62

Bølling-Allerød period (B-A), which spans about 2000 yr, is characterised by an important reduction in diatom numbers, a feature clearly visible in all cores. In core MD 95-2043, where the time resolution is better, the B-A period is easy to recognise and the observed minimum in diatom abundance shows a tendency towards lowest values at the Younger Dryas (YD) boundary. Following the B-A, diatoms reach their highest abundances in core MD 95-2043. A peak was also recorded in core KS8231 but was not found in cores KS8230 and TG-5. This peak was related to the YD. The diatom assemblage found at this peak in core MD 95-2043 is dominated by resting spores of Leptocylindrus danicus (63-89%) (Fig. 6), while in core KS8231 the major component of the diatom assemblage for this time interval is Thalassionema nitzschiodes var. nitzschiodes (30-50% of the assemblage) (Abrantes, 1988a). Moreover, in core

MD 95-2043, *Paralia sulcata* accounts for 4–16% of the diatom assemblage during the YD, but *L. danicus* and *P. sulcata* do not peak at the same time (Fig. 6).

T1b is marked by a continuous and rapid decrease in diatom abundance, with the final disappearance of diatoms in the early Holocene. Very few samples contained entire diatom valves and in most cases only fragments were found. At the surface, diatoms reappear, but in lower abundances than the values found downcore for MIS2 and the Termination.

Fresh-water diatoms and phytoliths were counted in core MD 95-2043 (Fig. 7). The main fresh-water

Sampling depth, calendar age, TOC content, diatoms/mm <sup>2</sup> and										
absolute abundance of diatom valves of core KS8231. For diatom										
countings, see also, Abrantes (1988a,b)										

Depth	Cal. age	TOC	Valves/	Valves/g
(cm)	$(\times 10^{3})$	%)	mm <sup>2</sup>	$(\times 10^4)$
0	0.00	0.66		2.70
6	0.64	0.50	0.45	3.17
10	1.07			Frag
25	2.68	0.39		Frag
30	3.26			Frag
40	4.67			Frag
48	0.00	0.49		Frag
50	6.08			Frag
75	8.29	0.52		Frag
100	9.62	0.43		Frag
110	10.28			3.50
120	10.94	0.64	0.32	2.28
130	11.83			7.50
140	12.42	0.46	0.59	4.17
160	14.09			10.80
162	14.31	0.40	0.46	3.26
170	15.18			4.50
175	15.73			4.20
180	16.27	0.47	0.58	4.07
192	17.40	0.52	0.56	3.97
200	17.52			1.50
210	17.68		0.36	2.53
220	17.83		0.62	4.40
230	17.99			6.40
240	18.29	0.76	0.43	3.08
250	18.87			5.60
260	19.45	0.56	0.46	3.24
280	20.38	0.83	0.29	2.08
290	20.78	0.93	0.25	1.79
300	21.19			0.72
312	21.68	0.76	0.53	3.75

Table 5

Core MD 95-2043: sampling depth, calendar age, estimated SST (°C), TOC content, diatoms/mm<sup>2</sup>, absolute abundance of diatom valves, fresh-water diatoms (FD) and opal phytoliths (Phy)

Depth (cm)	Cal. age $(\times 10^3)$	SST (°C)	Depth (cm)	Cal. age $(\times 10^3)$	SST (°C)	Depth (cm)	Cal. age $(\times 10^3)$	SST (°C)	Depth (cm)	Cal. age $(\times 10^3)$	Valves/ mm <sup>2</sup>	Valves/g $(\times 10^5)$	$\frac{\text{FD/g}}{(\times 10^5)}$	Depth (cm)	$\frac{\text{Phy/g}}{(\times 10^5)}$	Depth (cm)	TOC (%)
0	1.01	18.8	324	9.42	19.8	660	14.06	15.0	0	1.01	0.19	0.63	0.21	0	0.63	0	0.57
2	1.08	18.3	330	9.52	19.1	664	14.10	15.6	10	1.38	0.08	0.28	0.00	10	0.83	80	0.55
4	1.16	18.3	332	9.56	19.6	670	14.17	14.8	20	1.75	0.00	0.00	0.00	20	0.00	110	0.52
10	1.38	17.9	334	9.59	19.2	674	14.21	14.9	30	2.12	0.08	0.14	0.00	30	0.00	170	0.55
14	1.53	18.1	340	9.69	19.1	680	14.27	15.6	40	2.49	0.00	0.00	0.00	40	0.00	240	0.45
20	1.75	18.4	344	9.76	19.9	684	14.32	14.9	80	3.87	0.00	0.00	0.00	80	0.00	280	0.48
24	1.90	19.0	350	9.85	19.4	686	14.36	14.0	120	4.87	0.08	0.28	0.00	120	0.00	332	0.82
30	2.12	18.4	354	9.91	19.8	690	14.44	14.3	160	5.66	0.00	0.00	0.00	160	0.00	370	0.71
34	2.27	18.5	360	10.00	19.0	694	14.51	14.6	200	6.39	0.00	0.00	0.00	200	0.00	400	0.61
40	2.49	18.3	364	10.05	18.7	700	14.62	13.9	240	7.11	0.08	0.28	0.00	240	0.00	410	0.73
44	2.64	18.5	370	10.14	18.3	704	14.69	13.2	280	8.40	0.00	0.00	0.00	280	0.00	430	0.66
50	2.86	18.3	374	10.19	18.2	710	14.84	12.7	320	9.35	0.00	0.00	0.00	320	0.00	440	0.67
54	3.01	18.7	380	10.28	18.1	714	14.98	12.0	360	10.00	0.00	0.00	0.00	360	0.56	460	0.65
60	3.21	18.2	384	10.34	19.2	720	15.21	13.4	390	10.42	0.00	0.00	0.00	390	0.83	480	0.67
64	3.34	19.1	390	10.42	18.6	724	15.36	12.9	420	10.84	0.08	0.28	0.00	420	0.00	500	0.72
66	3.41	18.5	394	10.48	19.1	730	15.58	11.6	440	11.03	0.25	0.83	0.28	440	0.83	520	0.74
70	3.54	18.6	400	10.56	18.9	734	15.73	12.5	450	11.13	0.25	0.83	0.42	450	2.50	530	0.80
74	3.67	19.0	404	10.62	18.4	742	16.02	12.2	460	11.23	0.08	0.28	0.00	460	0.83	550	0.90
80	3.87	18.9	410	10.70	18.4	744	16.10	11.8	470	11.33	0.17	0.56	0.00	470	0.56	560	0.84
84	4.00	19.0	414	10.76	18.6	750	16.32	10.9	480	11.43	0.25	0.83	0.00	480	1.67	580	0.92
90	4.20	18.7	420	10.84	18.8	754	16.47	10.6	490	11.56	1.52	5.00	0.00	490	1.94	610	0.77
94	4.33	19.0	424	10.88	18.6	760	16.67	10.6	500	11.77	5.85	19.31	0.42	500	1.94	620	0.74
100	4.47	19.0	430	10.93	18.2	764	16.79	11.9	510	11.98	1.60	5.28	0.00	510	2.22	650	0.83
104	4.55	18.8	434	10.97	17.2	770	16.96	11.6	520	12.13	7.95	26.25	0.42	520	0.42	680	0.73
110	4.67	19.3	440	11.03	17.6	774	17.07	12.0	530	12.27	3.91	12.92	1.25	530	1.25	690	0.58
114	4.75	19.6	444	11.07	17.9	784	17.36	10.6	540	12.41	11.87	39.17	0.00	540	3.33	720	0.61
120	4.87	18.5	450	11.13	18.2	790	17.53	10.2	550	12.55	2.53	8.33	0.00	550	2.08	730	0.72
124	4.95	18.6	454	11.17	17.9	794	17.64	10.4	560	12.69	1.01	3.33	0.00	560	1.39	760	0.38
130	5.07	18.8	460	11.23	18.0	800	17.81	12.6	570	12.83	0.35	1.17	0.17	570	1.00	770	0.39
134	5.15	18.8	470	11.33	17.0	804	17.99	12.7	580	12.97	0.25	0.83	0.00	580	1.11	780	0.57
140	5.27	18.3	474	11.37	16.8	810	18.33	11.3	590	13.17	0.08	0.28	0.00	590	1.39	810	0.32
150	5.46	18.3	480	11.43	16.1	814	18.57	12.9	600	13.45	0.25	0.83	0.17	600	0.83	820	0.38
154	5.54	18.3	484	11.47	16.0	820	18.91	12.9	610	13.55	0.35	1.17	0.17	610	1.33	830	0.33
160	5.66	18.7	490	11.56	14.5	824	19.15	12.8	620	13.65	0.00	0.00	0.00	620	0.33	850	0.43
164	5.74	18.4	494	11.65	15.0	828	19.38	12.8	630	13.76	0.40	1.33	0.00	630	0.17	860	0.37
170	5.86	18.9	500	11.77	14.8	830	19.49	13.0	640	13.86	0.15	0.50	0.00	640	1.17	890	0.49
174	5.94	19.6	504	11.85	15.4	834	19.73	12.7	650	13.96	0.25	0.83	0.33	650	0.50	900	0.37
180	6.05	19.2	506	11.90	15.3	840	20.07	12.7	660	14.06	0.40	1.33	0.50	660	0.83	920	0.51
184	6.12	19.4	510	11.98	15.0	844	20.31	13.0	670	14.17	0.32	1.04	0.21	670	0.83	940	0.32

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Table 5 (continued)

Depth	Cal. age $(\times 10^3)$	SST	Depth	Cal. age $(\chi 10^3)$	SST	Depth	Cal. age $(\times 10^3)$	SST	Depth	Cal. age $(\times 10^3)$	Valves/	Valves/g $(\times 10^5)$	FD/g	Depth	Phy/g	Depth	TOC
(cm)	( X 10 )	( )	(cm)	( X 10 )	( )	(cm)	( X 10 )	( )	(cm)	( X 10 )	mm	( X 10 )	( x 10 )	(cm)	( x 10 )	(cm)	(%)
190	6.22	19.4	514	12.05	14.3	850	20.65	12.1	680	14.27	0.40	1.78	0.89	680	0.44	960	0.39
194	6.29	19.5	520	12.13	14.0	854	20.88	12.7	690	14.44	0.19	0.63	0.21	690	0.42		
200	6.39	19.5	524	12.19	13.3	860	21.23	12.0	700	14.62	0.15	0.50	0.33	700	0.50		
204	6.46	19.4	530	12.27	12.5	864	21.46	12.2	710	14.84	0.95	3.13	0.42	710	0.21		
210	6.56	19.1	534	12.33	12.5	870	21.81	11.4	720	15.21	0.91	3.00	0.83	720	1.00		
214	6.63	18.8	540	12.41	12.2	874	22.04	12.4	730	15.58	0.66	2.17	0.67	730	0.50		
220	6.73	18.8	544	12.47	12.1	880	22.39	12.1	738	15.88	0.61	2.00	0.67	738	1.33		
224	6.80	19.4	554	12.61	12.4	884	22.62	12.1	750	16.32	0.88	2.92	1.04	750	0.42		
230	6.91	19.7	560	12.69	12.0	890	22.97	11.6	760	16.67	0.86	2.83	1.17	760	0.67		
234	6.97	19.2	564	12.75	12.4	894	23.20	12.5	770	16.96	0.44	1.46	0.63	770	0.21		
240	7.11	19.3	570	12.83	12.7	900	23.55	11.3	780	17.24	0.71	2.33	0.17	780	1.00		
244	7.24	19.5	572	12.86	13.2	904	23.78	10.3	790	17.53	0.76	2.50	0.42	790	0.83		
250	7.43	18.9	580	12.97	13.8	910	24.13	9.9	800	17.81	0.45	1.50	0.33	800	2.50		
254	7.56	18.8	584	13.02	14.3	914	24.36	9.5	810	18.33	0.38	1.25	0.63	810	0.42		
260	7.75	18.8	590	13.17	14.7	924	24.94	12.6	820	18.91	0.51	1.67	0.56	820	1.39		
264	7.88	19.0	594	13.35	14.5	930	25.29	12.1	830	19.49	0.44	1.46	0.83	830	0.21		
270	8.07	18.0	600	13.45	14.8	934	25.52	12.2	840	20.07	0.96	3.17	1.67	840	1.50		
274	8.20	18.9	604	13.49	15.3	940	25.87	12.0	850	20.65	0.76	2.50	1.00	850	1.17		
280	8.40	18.2	610	13.55	14.7	944	26.10	13.6	860	21.23	0.45	1.50	0.67	860	1.67		
284	8.52	18.5	614	13.59	15.8	950	26.45	11.3	870	21.81	0.25	0.83	0.21	870	0.42		
290	8.72	18.5	620	13.65	15.6				880	22.39	0.34	0.56	0.42	880	0.14		
294	8.85	18.8	624	13.70	15.5				890	22.97	0.33	0.55	0.09	900	0.31		
300	9.01	19.1	630	13.76	15.1				900	23.55	0.42	0.34	0.08	920	0.76		
304	9.08	19.6	634	13.80	15.6				910	24.13	0.61	0.67	0.43	940	0.18		
310	9.18	18.9	640	13.86	15.5				920	24.71	0.46	0.51	0.05				
314	9.25	19.6	644	13.90	15.3				930	25.29	0.17	0.18	0.12				
320	9.35	19.1	650	13.96	15.2				940	25.87	0.67	0.73	0.31				
324	9.42	19.8	654	14.00	15.5				950	26.45	0.28	0.31	0.12				



Fig. 7. Core MD 95-2043. Absolute abundance of fresh-water diatoms, opal phytoliths and windblown particles per gram of dry sediment. Windblown particles = fresh-water diatoms + opal phytoliths. Reinterpretation of the PhFD index (Jansen and Van Iperen, 1991; see text for redefinition). Thicker line indicates a three-point smoothing. See also explanation in Fig. 3.

diatom in the Alboran Sea is *Aulacoseira granulata*, together with other limnobiontic forms (*Cyclotella ocellata* and *Stephanodiscus astrea*). The abundance pattern of these groups parallels that of the total diatom assemblage (Figs. 6 and 7). The maximum abundance of both groups occurs during the Last Glacial Maximum (LGM) and the YD; they then disappear after 8.5 ky BP, reappearing again in surface samples. Phytoliths are constantly present over 24–11 ky, with a double peak during the YD, and diatoms show the highest values at 21–19, 17–15 ky, and a smaller peak at the YD (Fig. 7).

#### 4.4. TOC content

The TOC accumulated in the Alboran Sea sediments during the time span considered in this work shows a similar pattern in all the cores studied. During MIS2 minimum values were recorded in all cores. Cores TG-5, KS8230 and KS8231 showed average values in weight percentages of around 0.75, 0.6 and 0.67, respectively, while in MD 95-2043 the values were around 0.5% (Figs. 3–6).

TOC increased during deglaciation, reaching a maximum at the YD event. The highest average

values (1.0 and 1.2%, respectively) were observed in cores TG-5 and KS-8230; KS8231 and MD 95-2043 reached 0.55 and 0.9%, respectively (Figs. 3–6).

For core MD 95-2043, TOC values are high after T1b to about 9 ky BP. During the Holocene, TOC values are lower but slightly higher than during the glacial stage. For core KS8230 resolution is too low for the Holocene, and for core TG-5 the TOC values are extremely variable.

# 5. Discussion

## 5.1. Preservation versus paleoproductivity

The diatom record in the sediments is controlled by two factors: primary productivity and preservation during diatom settling and sedimentation. Although the preservation of diatom valves in the Mediterranean Sea is very poor, when the silica supply from the surface water is sufficient, a deposition of diatom valves in the sediments is favoured. Previous studies by Abrantes (1988a) and Bárcena and Abrantes (1998) have already related the presence of diatoms in Alboran Sea sediments to upwelling conditions in the overlying waters. Further, high sedimentation rates also favour the preservation of diatom valves in the sediments.

In this work, the highest number of diatom valves per gram of dry sediment was recorded in core MD 95-2043 (Fig. 6), located in the deepest eastern Alboran Basin; this core also has the highest sedimentation rate. Currently, MD 95-2043 is outside the influence of both productive systems and the record of diatoms at the core top is low in comparison with the record at the YD. Furthermore, the diatom species found are robust forms and preferential dissolution cannot be discarded. Despite all these effects, if downcore variations in diatom abundance were related only to preservation, the number of diatoms would be much lower than that found today. Therefore, we consider that the high abundance of diatoms in the sediment must be related to the silica supply at the sea surface and hence linked to the productivity of the surface waters. Consequently, the downcore variations in diatom abundance would be a result of changes in paleoproductivity.

However, as pointed out by Abrantes (1988a) and

Bárcena and Abrantes (1998), abundance is much lower than in other strong upwelling areas (Abrantes, 1991). Furthermore, Abrantes (1988a) published a study on the diatom assemblages preserved in two of the cores discussed here, KS8230 and KS8231, and considers those diatom assemblages to be autochtonous (Abrantes, 1988a).

# 5.2. Changes in paleoproductivity

The persistence of the double gyre of the Alboran Sea is variable and its permanence depends on several factors (thermohaline circulation, meteorological factors, etc.) (Parrilla and Kinder, 1987). The present-day Mediterranean climate type is due to the position of the ITCZ in the upper atmosphere, its seasonal migration and the influence that it bears upon more localised pressure systems. In summer, a very stable high-pressure system is established across the Mediterranean, resulting in hot dry conditions. As the high-pressure system decays (movement of the ITCZ), the passage of southwesterly depressions influences the Mediterranean Basin through the winter months, resulting in prevailing westerlies in the Strait of Gibraltar. Therefore, the position of the ITCZ does produce distinct weather patterns in the Mediterranean area. Moreover, it seems likely that a more pronounced migration of the ITCZ could have largescale effects on regional climate and oceanographic circulation (Cramp and O'Sullivan, 1999). Parrilla (1984) observed that the permanence of atmospheric low pressure over the western Mediterranean affects the oceanography in the Alboran Sea. The dominance of the westerlies induces maximum stream flux; the stream enters the centre of the basin directly, and the WAG is displaced to the east (Parrilla, 1984). During glacial periods the ITCZ migrated to the south favouring the localisation of a low-pressure system over the Mediterranean with prevailing westerlies (COHMAP Members, 1988). Vergnaud-Grazzini and Pierre (1991) suggested that before 16 ky, prevailing westerlies and low pressures over the Mediterranean favoured the eastward displacement of the WAG of the Alboran Basin. Therefore, we infer that during MIS2 meteorological conditions would have induced an intensified eastward flow, which would have displaced the WAG to the east. The North Alboran upwelling system would have also moved and hence

maximum fertility in surficial waters could have occurred in our easternmost cores. In general, we suggest that for the time interval considered in this study, the highest productivity was displaced towards the east; the easternmost cores record the highest diatom abundance, while cores TG-5 and KS8230 have the lowest values (Figs. 3–6).

Based on the diatom record, paleoceanographic conditions favoured increases in paleoproductivity from 22.5 to 16 ky during the LGM with a maximum at ca. T1a. During the B-A event, a significant reduction in paleoproductivity occurred, followed by a new increase in paleoproductivity during the YD (Figs. 5 and 6). Several paleoproductivity indicators (coccolithophorids, dinoflagellate cysts,  $\delta^{13}$ C and benthic foraminifera) point to changes in paleoproductivity during the last 25 ky in the Alboran Sea. These have been attributed to variations in oceanographic conditions (Abrantes, 1988a; Weaver and Pujol, 1988; Turon and Londeix, 1988; Targarona et al., 1997a; Vergnaud-Grazzini and Pierre, 1991). However, our diatom record does not reveal evidence of an intensification in sea surface productivity during the YD in the cores from the upwelling region (TG-5 and KS8230) (Figs. 3 and 4), which is in good agreement with the findings of Targarona et al. (1997a). During the Holocene the inferred paleoproductivity decreased to the lowest values of the last 23 ky; this was also deduced by Abrantes (1988a,b). Paleoproductivity was slightly reestablished during the Recent.

## 5.3. Deep water conditions

On comparing the micropaleontological data with the TOC content, several discrepancies are seen (Figs. 3–6). Core MD 95-2043 has previously been studied by Sierro et al. (1998). Based on the high TOC content, high alkenone concentration, low magnetic susceptibility, high fragmentation of planktonic foraminifera and a marked reduction in the abundance and diversity of benthic foraminifera with a characteristic fauna of Mediterranean sapropels, these authors identified the youngest ORL of the Western Mediterranean between 16.5 and 9 cal ky BP; that is, including the YD. The initiation of this ORL is therefore about 4 ky older than the base of sapropel 1, while the top of the ORL is 1 or 2 ky older than that of sapropel 1. Alkenone concentrations reach very high values (1000-1800 ng/g) in the Alboran ORL (Sierro et al., 1998). These values are even higher than those recorded in the S1 layer (300-800 ng/g)from the Levantine Basin (ODP site 976, Emeis et al., 1998). All the foregoing evidence suggests that strong changes in deep-ventilation/surface-productivity occurred in both Eastern and Western Mediterranean basins but in different periods. The possible connection between these two events and the hydrological pattern associated with them remains a complex puzzle. On comparing the UK'37-SST from the Alboran, Levantine and Ionian basins during the S1 deposition time (8.6–10.2 ky BP; Emeis et al., 1998) similar values can be observed (18-19°C), indicating that the present thermal gradient between these basins was drastically reduced. This observation confirms previous results on  $\delta^{18}$ O gradients across the Mediterranean Sea, which have been interpreted to be roughly half those of present-day times (Rohling and De Rijk, 1999). Therefore, our results also support the hypothesis that during the S1 the Mediterranean Sea functioned as a less active concentration basin than today (Rohling and De Rijk, 1999). On the other hand, the low planktonic foraminifera fragmentation and the benthic microfauna clearly suggest the existence of well-oxygenated water during the LGM. After T1a, conditions changed to carbon-rich and oxygen-poor waters. Sierro et al. (1998) consider that during the YD bottom ventilation seemed to recover slightly. These new data confirm Caralp's inferences based on benthic foraminifers (Caralp, 1988).

The TOC content in cores KS8230 and TG-5 follows the same pattern as that observed for MD 95-2043 (Figs. 3, 4 and 6), and we therefore consider that for the time span between 16.5 and 9 cal ky BP bottom ventilation was reduced. The existence of poorly oxygenated bottom waters in the Western Mediterranean may be explained by increasing surface productivity or slower rates in deep water renewal. The diatom record, a proxy for surface productivity, is to some extent independent of deep water properties and indicates that during the LGM, T1a and the YD, sea surface productivity was higher, while during the B-A and especially after T1b it was lower. The opposite behaviour between the planktonic and benthic proxies (Sierro et al., 1998; Caralp, 1988) suggests that although surface productivity plays a role in increasing organic fluxes to the bottom, an important reduction in deep water renewal should also be considered.

## 5.4. Surface water conditions

Recent SST studies on core MD 95-2043 (Cacho et al., 1999) have revealed evidence of a clear connection between the Greenland  $\delta^{18}$ O record in GISP2 (Messe et al., 1997) and its D-O events and the Alboran SST, as well as good coherence among the temperature minima in both records. The study by Cacho et al. (1999) also recognises the influence of the Heinrich events in the Alboran Sea and, based on the presence of N. pachyderma (s), the authors conclude that important invasions of polar waters occurred. Pujol and Vergnaud-Grazzini (1989) observed a peak in N. pachyderma (s) in core KS8230 during the time span corresponding to the H1, which, at the time, was interpreted as a North Atlantic cool water invasion. Weaver and Pujol (1988) recognised a relationship between SST and coccolitophorid assemblages, where the absence of Gephyrocapsa oceanica (considered by the authors to be a warm water species) agrees with low temperatures. For the time span under study, there is therefore evidence that our cores show a clear relationship between North Atlantic and Mediterranean Sea climatology. Do diatom assemblages also respond to these cool water invasions?

The increase in paleoproductivity, associated with the easternmost cores, during the YD is inferred from a diatom peak dominated by L. danicus RS in core MD 95-2043 (Figs. 5 and 6) and T. nitzschioides var. nitzschioides in core KS8231 (Abrantes, 1988a). L. danicus RS is a common neritic species distributed world-wide and associated with cold and temperate waters. This species dwells preferentially in North Atlantic waters (Hargraves, 1990). It has been related to the stratified phase of a turbulent environment (Hobson and McQuoid, 1997). Recently, blooms of L. danicus RS have been recorded in water lenses with relatively low salinity in relation to the Ebro river delta (Mediterranean Sea) (Estrada, personal communication). Moreover, the observation of plankton in the Alboran Sea points to increases in biomass in an N-S transect from the Spanish coast to the Alboran Island at 3°W during January, L. danicus

being the main component of the assemblage. This increase has been related to the presence of ASW, which contributes to increased stability in the upper part of the water column (Delgado, 1990). In the Portuguese and NW African margins, this species is common during spring and summer and this could be related to nutrient depletion (T. Moita, personal communication). French (1982) concluded that spore production by L. danicus is a response to nitrogen depletion. Core top countings in the Almeria-Oran Front area (Bárcena, in preparation) demonstrate that this species accounts for 10-20% of the total diatom assemblage, but it does not occur at the high abundances found in this core; the rest of the surface sediment diatom assemblage is composed by Chaetoceros RS, T. nitzschioides, T. bacillaris, Thalassiothix sp., Thalassiosira spp., Coscinodicus spp. and P. sulcata, which is present at low abundances (2%). On the other hand, a similar maximum in L. danicus has been observed in ODP core 977 for the same time interval as the one observed in MD 95-2043 core (Bárcena, in preparation). T. nitzschioides var. nitzschioides is a planktonic species associated with occasional upwelling, or with nutritive waters transported away from upwelling centres (Blasco et al., 1981; Pokras and Molfino, 1986; Pokras, 1987; Abrantes, 1988a,b, 1990; Bárcena and Abrantes, 1998). Furthermore, P. sulcata, a bottom taxon which can survive in a planktic mode, has been used as an indicator of high littoral production (Margalef, 1978; Margalef et al., 1979; Abrantes, 1988b, 1990, 1991). Moreover, Sancetta et al. (1992) related the presence of P. sulcata in deep-sea sediments to lateral transport off the continental shelf. The deepest Alboran Basin could act as a giant sediment trap, collecting material advected from the shelf, and P. sulcata, recorded in the depeest core MD 95-2043, could have been transported from shallower areas of the Alboran Basin.

The YD is the last of a long series of alternations between colder and warmer conditions that occurred during the Last Glacial Period, the D–O events (Broecker, 1992). As proposed by Broecker (1992), the YD cold event was triggered by the diversions of meltwaters from the ice sheets surrounding the Atlantic, which spread fresh water across the North Atlantic. Accordingly, Mediterranean circulation patterns would also have been affected by this. Therefore, a salinity decrease in the ASW entering the Alboran Sea would explain the L. danicus peak in core MD 95-2043 (Fig. 6) as resulting from relatively cooler and low-salinity waters and stratification in the upper part of the water column. Moreover, it is also possible that local rivers could have brought increased amounts of fresh water during the YD event as a result of glacier melting from the mountains surrounding the Alboran Sea. In this sense, Sancetta (1994) proposed enhanced production associated with the low-salinity plume from river discharges. Both possible mechanisms of introducing fresh water into the Mediterranean could act simultaneously and they are not mutually exclusive. Meltwater pulses or surges entering the Alboran Sea would change its density structure, and would affect deep oxygen renewal, avoiding the oxidation of the organic matter at depth.

Meteorological forcing would have affected the paleoceanography of the region; prevailing westerlies and low pressures over the Mediterranean during this event would have favoured the eastward displacement of the WAG of the Alboran Basin, and the coastal upwelling would have been placed at around  $3^{\circ}$ W. The presence of a *T. nitzschioides* peak in core KS8231 (Abrantes, 1988a) together with the absence of diatoms in the westernmost cores agree with this inference.

On the other hand, during Heinrich events H2 and H1 the diatom assemblage did not respond to the North Atlantic cool water invasions in the same way as during the YD. Diatom assemblage is composed of planktonic taxa such as *T. nitzschioides* var. *nitzschioides*, *Thalassiosira eccentrica*, *Rhizosolenia* spp. and *Alveus marinus*. Cacho et al. (1999) have reported that during the Heinrich events the hydrodynamic response of the Alboran sea to the climatic changes was more intense than for the D–O stadials. This evidence is in agreement with the diatom record, which reveals different surface water conditions during the Heinrich events.

## 5.5. Atmospheric conditions

The fresh-water diatom group is common in surface sediment samples from the Alboran Sea (Abrantes, 1988a; Bárcena, in preparation). Abrantes (1988a) postulated two possible sources: the windblown diatoms from NW Africa or introduction of diatombearing sediments from the Guadalquivir River into the Alboran Basin. In the present work, fresh-water diatoms (A. granulata as the major component of the group, together with limnobiotic taxa) as well as phytoliths were observed to follow a distribution pattern synchronous to that of the marine diatoms (Figs. 6 and 7). Opal phytoliths, silica-rods from epidermic cells in grasses, are injected into the atmosphere during dry-season brush fires. Freshwater diatoms (A. granulata) come from the deflation of diatomaceous deposits in dry lake beds (Pokras and Mix, 1985). Thus, the presence of a mixture of microfossils with distinct ecological requirements - terrestrial and lacustrine environments - allows us to interpret this fact as a wind-transported assemblage rather than as being of a run-off origin. Moreover, there are no important river systems in the Alboran Basin (Fig. 1). Dust storms are common in North Africa during the arid and dry seasons, and satellite images point to southern Algeria (Great Western Erg region) as a possible source area (Bücher and Lucas, 1984; Seyve and Fourtanier, 1985). Additionally, the presence of "red snow" over the Spanish coast has been widely described. This "red snow" is related to southwesterlies generated by low pressures over the Gulf of Cadiz (Parrilla and Kinder, 1987). Therefore, NW Africa is the most feasible source area for this windblown group, possibly indicating more arid conditions and intensified winds. The African climate is clearly linked to that of high-latitude ice sheets, with glacial aridity and interglacial humidity (Sarnthein et al., 1982; Stabell, 1986). During glacial periods the ITCZ migrated to the south, favouring the localisation of a low-pressure system over the Mediterranean and prevailing westerlies and therefore favouring "red snow" over the Mediterranean. Previous studies by Bárcena et al. (1997) and Flores et al. (2000) have reported different types of behaviour between the two types of siliceous windtransported microfossil. The abundance of fresh-water diatoms was used as an aridity indicator, while opal phytoliths were linked to relatively wet conditions on land.

Downcore variations in total windblown particles (fresh-water diatoms + opal phytoliths) (Fig. 7) are evidence of changes in wind conditions. Maximum wind intensity occurred during the LGM, with a decreasing tendency towards the glacial-interglacial transition. During the B–A period wind intensities were minimum, being re-established during the YD.

An aridity index was calculated based on the following equation: phytoliths/phytoliths + freshwater diatoms (Jansen and Van Iperen, 1991). These authors consider river transport for fresh-water diatoms in the Zaire fan region and interpret maxima in the group as originating in humid periods. Since our siliceous microfossils are wind-transported, our interpretation of the aridity index (PhFD) is opposite to that of Jansen and Van Iperen. Consequently, low values in the PhFD index are interpreted here as indicators of arid conditions. According to our PhFD index interpretation, MIS2 was characterised by arid conditions in the area, but from 18.5 to 17.3 cal ky BP a relatively humid period can be identified, which is related to a phytolith peak (Fig. 7). Fresh-water diatoms would indicate arid conditions in North Africa, with lake desiccation during the MIS2 until 15 cal ky BP. The progressive decline in phytoliths would also indicate a gradual climatic deterioration (Fig. 7). The progressive reduction of fresh-water diatoms together with the increase in phytoliths during the B-A period would correspond to the reestablishment of vegetation belts and the replenishment of lakes and this would have continued throughout the rest of the deglaciation and would have prevailed until ca. 8 cal ky BP. During this period the PhFD index has higher values, which have been interpreted as humid conditions. Nevertheless, during the YD climatic conditions could have been relatively dryer, as indicated by the lower values of the PhFD index (Fig. 7).

These observations are in agreement with the palynological data obtained in the Alboran Sea by Turon and Londeix (1988), Cheddadi et al. (1991) and Targarona et al. (1997b). Our observations also agree with an intensification of the African monsoons. Humid periods in the Mediterranean Sea area are related to an intensification of the African monsoons, which are heavier during maxima in Northern Hemisphere summer insolation (Cramp and O'Sullivan, 1999). During the last deglaciation, the Northern Hemisphere supported maximum summer insolation (at ca. 520 W/ m<sup>2</sup>), together with an increase in precipitation in winter as well as during summer, and a well-developed herbaceous belt, suggesting warm and relatively moist climatic conditions during the deglaciation.

#### 6. Conclusions

The high abundance of diatoms in the sediment is related to the productivity of surface waters. Downcore variations in diatom abundances result from changes in paleoproductivity. Based on the diatom record, we can estimate that paleoceanographic conditions would have favoured increases in paleoproductivity during the LGM, with a maximum at ca. T1a, and during the YD event, especially in the easternmost area studied.

During the LGM period, meteorological conditions, with increased westerlies, induced an intensified eastward flow which would have displaced the WAG to the east. A possible eastward displacement of the North Alboran upwelling system would also have been involved.

We consider that for the time span between 16.5 and 9 cal ky BP bottom ventilation was reduced. This interpretation is based on opposite behaviour between the planktonic and benthic proxies. Although surface productivity plays a role in increasing organic fluxes to the bottom, an important reduction in deep water renewal should also be considered.

A decrease in salinity in the incoming ASW accounts for the *L. danicus* peak in core MD 95-2043 as resulting from relatively cooler and less saline waters and stratification in the upper water column. Two possible mechanisms for the introduction of fresh water into the Mediterranean during the YD, favouring *L. danicus* development, are considered. Cooler and less saline ASW triggered by the diversions of meltwaters from the ice sheets surrounding the Atlantic, and fresh-water lenses from the runoff of local rivers are both possible and not mutually exclusive mechanisms for introducing fresh water into the Alboran Sea.

The abundance of fresh-water diatoms is used as an aridity indicator, while opal phytoliths are linked to relatively wet conditions on land. During MIS2 up to 14 cal ky BP the region supported a gradual climatic deterioration with a loss of the grass-belt. A moister episode is interpreted from 18.5 to 17.3 cal ky BP. During the B–A, climatic conditions would have been warmer and moister, prevailing up to ca. 8 cal ky BP. During the YD, climatic conditions could have been dryer. Moreover, wind intensity was maximum during the LGM and the YD.

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