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Hydrological relationship between the North Atlantic Ocean and the Mediterranean Sea during the past 15 - 75 kyr

Martine Paterne,[1] Nejib Kallel,[1,2] Laurent Labeyrie,[1] Maryline Vautravers,[3]
Jean-Claude Duplessy,[1] Martine Rossignol-Strick,[4] Elsa Cortijo,[1]
Maurice Arnold,[1] and Michel Fontugne[1]

Abstract. The Mediterranean Sea hydrology at the time of the Heinrich formation in the North Atlantic Ocean was analyzed by comparing sea surface temperatures (SSTs) and oxygen isotope composition of seawater (δw) changes during the past 75 kyr in two marine cores. These were compared to the palynological variations derived in the Mediterranean Sea core. During the last glacial the two oceanic SST records show similar and synchronous patterns, with several long-term cooling periods, ending by abrupt SST increases. At the time of the Heinrich events, cold SSTs and low salinity prevailed in the Mediterranean Sea. The freshwater budget was similar to the modern one, permitting the presence of a mixed forest on the Mediterranean borderlands. The post-Heinrich periods are marked by a freshwater budget decrease, limiting oak and fir tree growth in the Mediterranean region. Increase of precipitation or reduction of evaporation is observed before the Heinrich episode, and is associated with a well-developed mixed Mediterranean forest.

1. Introduction

Recent paleoceanographic investigations have shown that the hydrological changes in the North Atlantic Ocean and in the Mediterranean Sea were directly connected over the past 18 kyr, except at ~8500 ka during the formation of Sapropel S1 [*Kallel et al.*, 1997a, b]. During the last glacial period the North Atlantic area experienced several abrupt climatic changes known as Heinrich events in the marine sediments [*Heinrich*, 1988; *Broecker et al.*, 1992; *Bond et al.*, 1992, 1993] and Dansgaard-Oeschger events in marine and ice cores [*Dansgaard et al.*, 1993; *Bond and Lotti*, 1995; *Rasmussen et al.*, 1996]. Associated detritals in the marine environment are restricted to an ice-rafted belt between 40° and 55°N [*Ruddiman*, 1977]. They are related to massive discharges of icebergs originating from the Baffin Bay and/or the Arctic/northern Eurasian areas [*Grousset et al.*, 1993; *Gwiazda et al.*, 1996; *Revel et al.*, 1996].

The resulting low-salinity anomalies of the surface waters subsequently affected the North Atlantic Deep Water production [*Oppo and Lehman*, 1995; *Sarnthein et al.*, 1994; *Vidal et al.*, 1997; *Zahn et al.*, 1997; *Vidal et al.*, 1998], and consequently, the North Atlantic heat pump weakened. In the same time, the climate of the surrounding continental areas would be modified by changes of the atmospheric and associated moisture transports during these events [*Grimm et al.*, 1993; *Guiot et al.*, 1993; *Watts et al.*, 1996a].

The aim of this paper is to analyze the hydrological behavior of the Mediterranean area at the time of the huge ice sheet calvings in the North Atlantic Ocean. Detailed oxygen isotope analyses of planktonic foraminifera ($\delta^{18}O$), sea surface temperatures (SSTs) and seawater (δw) records were developed on two well-dated deep-sea sediment cores collected in the ice-rafted belt in the North Atlantic Ocean and in the Tyrrhenian Sea (Figure 1). Moreover, the Tyrrhenian Sea core yielded a pollen record [*Rossignol-Strick and Planchais*, 1989], which allows direct comparison of the continental and marine climatic conditions prevailing outside the immediate North Atlantic region during the last glacial.

2. Methods

We derived an abolute chronology by using ^{14}C age calibration [*Stuiver and Reimer*, 1993; *Mazaud et al.*, 1991] in order to compare the hydrological changes in the North Atlantic Ocean and in the Tyrrhenian Sea. Procedure to derive seawater $\delta^{18}O$ (δw) changes has already been described [*Labeyrie et al.*, 1986; *Duplessy et al.*, 1991; *Maslin et al.*, 1995; *Kallel et al.*, 1997a]. It is based on SST estimates and on measurements of oxygen isotope compositions of planktonic foraminiferal shells which reflect changes of 1) the global ocean water δw composition due to ice volume fluctuations, 2) the local sea surface δw due to the freshwater budget (precipitation **P** + runoff **R**/evaporation **E**), and 3) the isotopic fractionation between calcium carbonate and water, which depends upon the temperature **T*** at which foraminifera form their tests [*Epstein*, 1953; *Shackleton*, 1974].

[1]Laboratoire des Sciences du Climat et de l'Environnement, Unité mixte CNRS-CEA, Gif sur Yvette, France.
[2]Now at Département des Sciences de la Terre, Université de Sfax, Sfax, Tunisia.
[3]Département de Géologie et Océanographie, Université de Bordeaux I, Talence, France.
[4]Laboratoire de Paléontologie et Palynologie, Université Pierre et Marie Curie, Paris, France.

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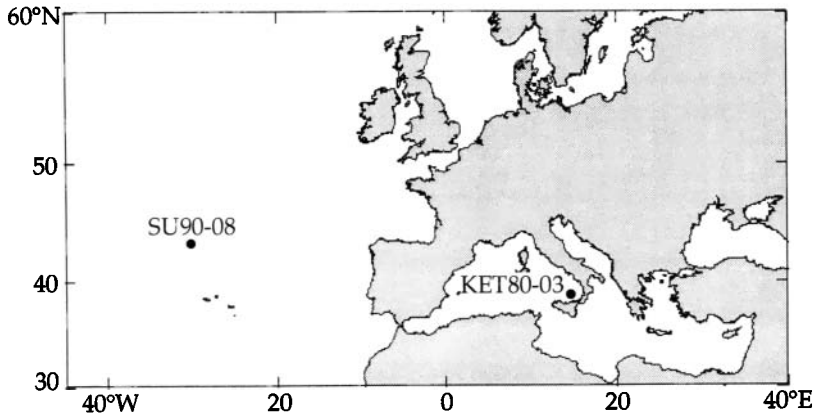

Figure 1. Core locations: SU 90-08: 43°03.1'N, 30°02.5W (water depth, 3080 m; length, 12.27 m) and KET 80-03: 38°49'2 N, 14°29'5 E (water depth, 1900 m; length, 10.10 m).

2.1. Chronology

Accelerator mass spectrometry (AMS) ^{14}C datings were performed on monospecific foraminifera in the two cores (Figures 2, 3, and 4 and Table 1). An additionnal dating in the North Atlantic Ocean core SU 90-08 was provided by the identification of the well-dated Ash Zone I [*Bard et al.*, 1993]. Furthermore, Ash Zone II was also recognized in this core, the age of which was estimated at ~57.5 ka [*Smythe et al.*, 1985]. In the Tyrrhenian Sea core KET 80-03, additional datings were provided by the K/Ar ages of some Italian pyroclastic eruptions, geochemically identified in the central Mediterranean Sea sediments [*Paterne et al.*, 1986; 1988]. Because of the presence of such absolute ages, the ^{14}C ages were thus expressed in calendar ages for the past 18 kyr using *Stuiver and Reimer*'s [1993] calib 3.0 program. Below 18 ka a magnetic correction has been applied to the ^{14}C ages [*Mazaud et al.*, 1991]. We assumed a constant reservoir age of the surface waters (400 years) through time in the two oceanic areas, though such an assumption may generate errors of up to 800 years in the North Atlantic Ocean [*Bard et al., 1993; Austin et al.*, 1994]. The age of the oxygen isotope subevent 4.2 was fixed at 65 ka [*Imbrie et al.*, 1982] and of the oxygen isotope transition 4.0 was fixed at 62 ka, as an average of the ages obtained by linear interpolation (Figures 2, 3 and 4). Chronology was then assessed by linear interpolations between two dated levels in each core. Time resolution of the sampling interval is ~550 years in both cores, with a 2 cm sampling interval in core SU 90-08 and a 5 cm one in core KET 80-03. In the last core, resolution increased to some 300 years from 500 to 680 cm.

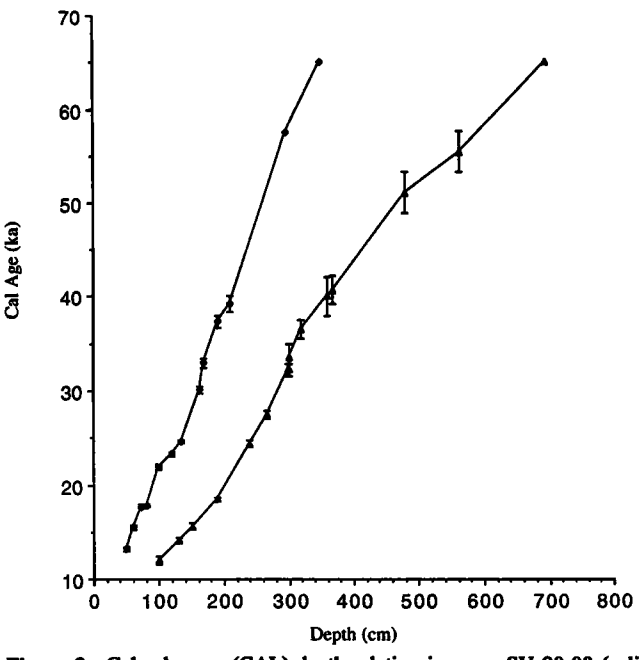

Figure 2. Calendar age (CAL)-depth relation in cores SU 90-08 (solid diamonds) and KET 80-03 (solid triangles).

2.2. Sea Surface Temperature Estimates

The North Atlantic Ocean and Tyrrhenian Sea SSTs were estimated using the modern analog technique [*Hutson*, 1979; *Prell*, 1985] (Figures 3 and 4). Method and derived SST estimates were previously presented for core SU 90-08 [*Cortijo et al.*, 1997] and for the past 18 kyr for core KET 80-03 [*Kallel et al.*, 1997a, b].

In core SU 90-08, SST estimates are based on good modern analogs with dissimilarity coefficients mainly lower than 0.15 [*Overpeck et al.*, 1985]. Nevertheless, SST values with standard deviations larger than 2.5°C have not been taken into account in further seawater $\delta^{18}O$ calculations. In core KET 80-03, modern analogs are mainly found in the North Atlantic Ocean during glacial conditions. Dissimilarity coefficients are larger than in core SU 90-08 but rarely exceed 0.25 (Figure 3) because of the coexistence of foraminifera characteristic of the North Atlantic Ocean and the Mediterranean Sea. This feature is enhanced by bioturbation effects, as dissimilarity coefficients increase when sea surface temperatures vary abruptly (Figure 4). Thus we considered that SST estimates are reliable along the Tyrrhenian Sea core, except when standard deviations are larger than 2°C.

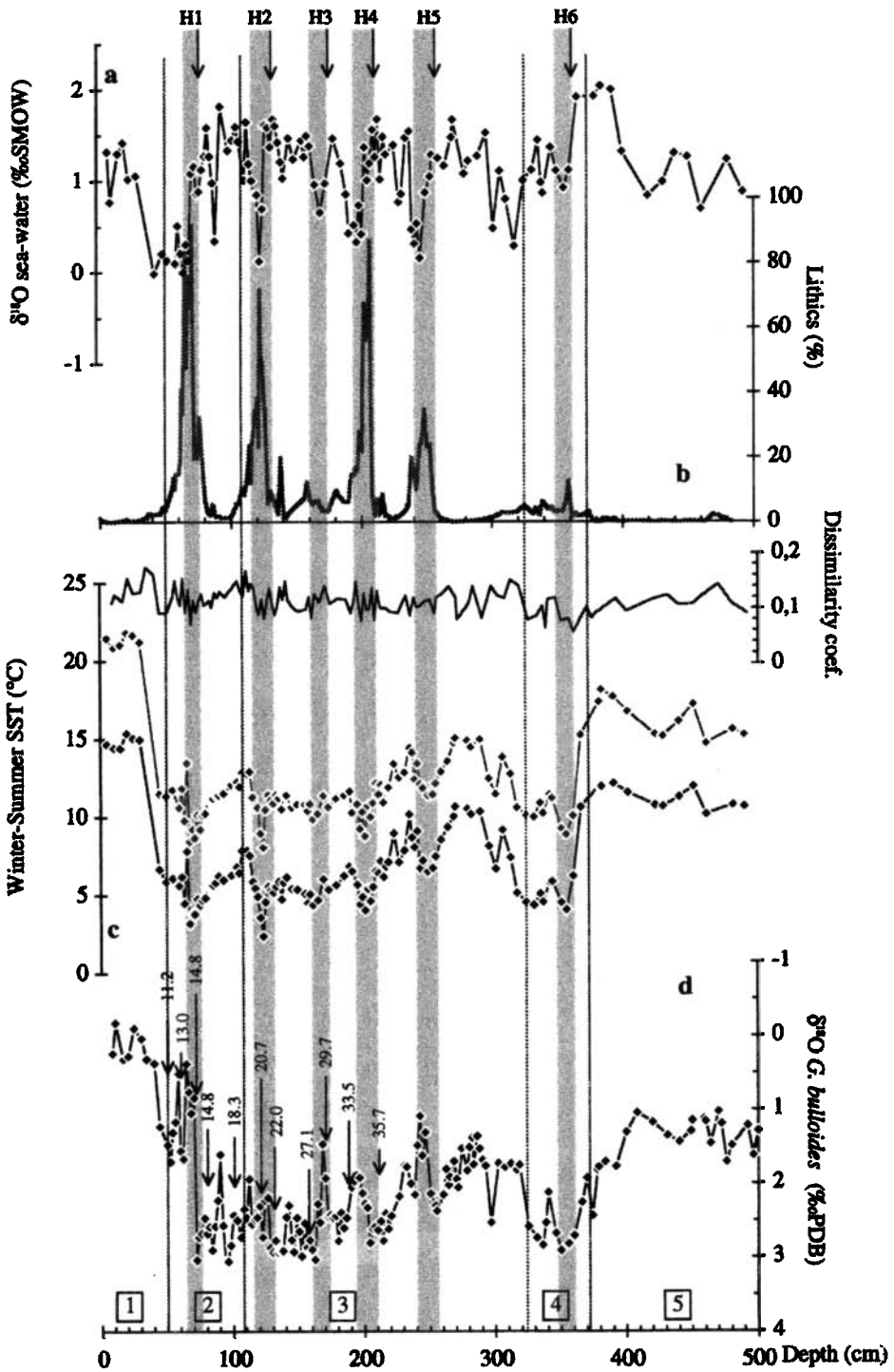

Figure 3. Plot of (a) seawater δw (‰ SMOW) (diamond line, left axis) and the lithics abundance (per cent) (shaded line, right axis), (b) winter and summer sea surface temperatures SSTs (degrees celsius) (solid diamond and shaded diamond lines, left axis) and dissimilarity coefficients (solid line, right axis), and (c) $\delta^{18}O$ (‰ Pee Dee belemnite (PDB) record of *Globigerina bulloides* as a function of depth in North Atlantic deep-sea core SU 90-08. Sampling interval for SST data is 2 cm until 260 cm and 10 cm below. Oxygen isotope stage boundaries are also shown. Dashed areas represent the Heinrich events from H1 to H6, defined by the increase of the magnetic susceptibility and of ice-rafted detritals [*Grousset et al.*, 1993; *Vautravers*, 1997], except for H3 marked by a δw low (see text). The ^{14}C datings are reported.

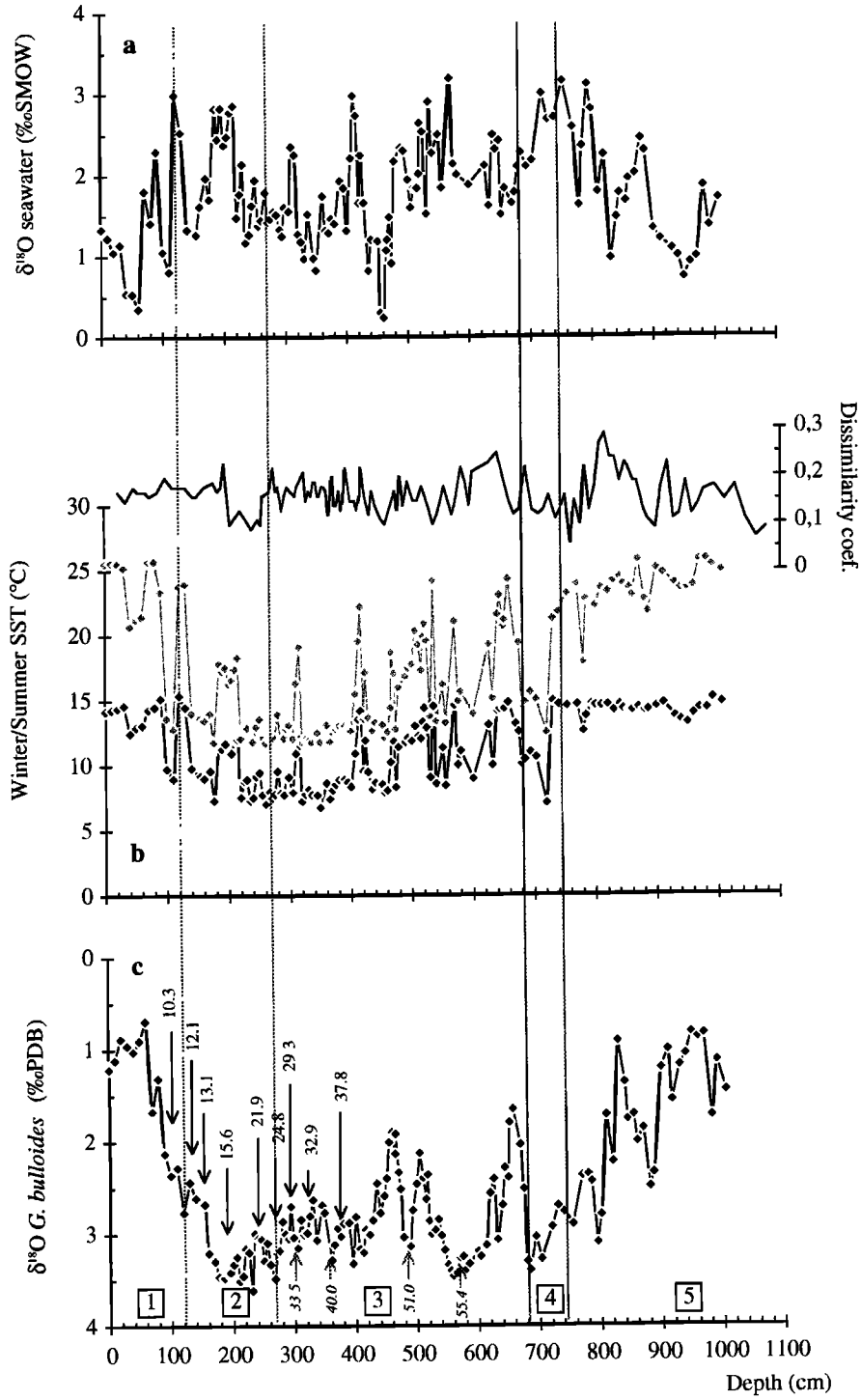

Figure 4. Plot of (a) seawater δw (‰ SMOW), (b) winter and summer SST (degrees celsius) (solid diamonds and shaded diamonds lines, left axis) and dissimilarity coefficient (solid line, right axis), and (c)$\delta^{18}O$ (‰ PDB) record of *G. bulloides* as a function of depth in the Tyrrhenian deep-sea core KET 80-03. Oxygen isotope stage boundaries and the ^{14}C and K/Ar (in italics) datings are also reported.

Table 1. AMS ^{14}C Datings on *Neogloboquadrina pachyderma left* in Core SU 90-08 and *Globigerina bulloides* in Core KET 80-03, Picked in the Peaks of Abundance

SU 90-08					KET 80-03				
Depth	AMS ^{14}C	1σ	Cal Ages	1σ	Depth	AMS ^{14}C	1σ	Cal Ages	1σ
50[a]	11.20[a]	0.15[a]	13.20[a]	0.16[a]	100	10.27	0.16	11.99	+0.25/-0.43
60	13.04	0.11	15.50	0.22	130	12.14	0.18	14.17	0.27
72	14.83	0.11	17.73	0.15	150	13.10	0.20	15.58	0.35
80	14.88	0.10	17.79	0.14	190	15.59	0.19	18.49	0.20
100	18.33	0.15	21.89	0.25	240	21.89	0.39	24.39	
120	20.70	0.21	23.33		267	24.84	0.42	27.46	
134	22.05	0.20	24.60		298	29.30	0.59	32.20	
162	27.08	0.33	30.10		300	29.30	0.70	32.20	
170	29.69	0.51	32.95		300[a]		1.50[a]	33.50[a]	
192	33.45	0.66	37.33		320	32.87	0.97	36.49	
210	35.73	0.88	39.20		360[a]		2.00[a]	40.00[a]	
297[a]			57.50[a]		370	37.80	1.50	40.68	
350[b]			65.00[b]		480[a]		2.20[a]	51.00[a]	
					562[a]		2.20[a]	55.40[a]	
					694[b]			65.00[b]	

AMS, accelerator mass spectrometry; Cal, calendar ages. The AMS ^{14}C dating at depth 72 cm in core SU 90-08 has been considered too old and has not been not taken into account.

[a] Ages referred to tephra layers.

[a] Ages correspond to Milankovitch ages of oxygen isotope subevent 4.2 [*Imbrie et al.*, 1982].

2.3. Planktonic ($\delta^{18}O$) and Seawater (δw) Oxygen Isotope Compositions

Oxygen isotope records are developed from *Globigerina bulloides* picked in the 250-315 μm size in both cores. The North Atlantic Ocean record was previously presented [*Grousset et al.*, 1993; *Cortijo et al.*, 1997] and is here restricted to the interval from Holocene to oxygen isotope stage transition 4/5 (Figure 3).

The modern seawater (δw) oxygen isotope values are estimated using the modern salinity values [*Levitus*, 1982] and the modern δw/salinity relationship [*Duplessy et al.*, 1991; *Kallel et al.*, 1997a]. They are estimated at 0.7‰ for a salinity of S = 35.9 and at 1.4‰ for S = 38.03 at core sites in the North Atlantic Ocean and in the Mediterranean Sea, respectively. A 0.5‰ δw change corresponds to a 1 and a 1.22 salinity change, respectively. Past δw values were estimated by solving the paleotemperature equation [*Shackleton*, 1974], using the optimal growth temperature **T*** [*Duplessy et al.*, 1991]. For the species *G. bulloides* it is equivalent to **T*** = summer SST - 1°C in the range 7°-22°C in the North Atlantic Ocean [*Duplessy et al.*, 1991] and to **T*** = April-May SST in the range 14°-16°C in the Mediterranean Sea [*Kallel et al.*, 1997a].

During the past 75 kyr, summer SST fluctuations vary from 8° to 22°C in core SU 90-08 (Figure 2). They cover the whole range of the SST calibration of **T*** for *G. bulloides* in the North Atlantic Ocean [*Duplessy et al.*, 1991]. Thus the isotopic temperature **T*** = summer SST - 1°C will be used to calculate the δw compositions in the North Atlantic Ocean core (Figure 3). In the Mediterranean Sea core, April-May SSTs vary from 8° to 17°C (Figure 4), mainly below the calibration temperature range of **T*** in the Mediterranean Sea (14°-16°C) [*Kallel et al.*, 1997a]. Nevertheless, we assumed that the seasonal bloom in the Mediterranean Sea occurred within the optimal temperature range determined in the North Atlantic Ocean [*Duplessy et al.*, 1991]. Taking into account the 0.07‰ error due to mass spectrometer measurements and the mean standard deviation on SSTs, the averaged errors on the δw estimates are 0.45‰ for the Tyrrhenian Sea (σ_{SSTs} = 1.14°C) and 0.60‰ for the North Atlantic Ocean (σ_{SSTs} = 1.79°C).

3. Results

Within the absence of any sedimentological signature of the Heinrich events in the Mediterranean Sea the study of the climatic conditions during these events in this area basically rests on chronology. We briefly present the chronology of the North Atlantic Ocean paleohydrological records as they were previously discussed in detail [*Grousset et al.*, 1993; *Cortijo*, 1995; *Cortijo et al.*, 1997; *Vidal et al.*, 1997; *Elliot et al.*, 1998]. The stratigraphical position of the Heinrich events in the Mediterranean Sea was then assessed by comparing the North Atlantic Ocean and Mediterranean Sea paleohydrological records before analyzing the climatic response of the Mediterranean area to the short-term climatic variability in the North Atlantic Ocean.

3.1. Heinrich Events in the North Atlantic Ocean Core SU 90-08

In the North Atlantic Ocean the Heinrich events are characterized by an increase of the abundance of coarse lithics, by cold SSTs, by depleted $\delta^{18}O$ values of the planktic foraminifera, and by low δw values [*Heinrich*, 1988; *Broecker et al.*, 1992; *Bond et al.*, 1992, 1993; *Maslin et al.*, 1995].

Such features are recognizable in core SU 90-08 (Figures 3 and 5). Peaks of abundant lithic concentrations were identified and

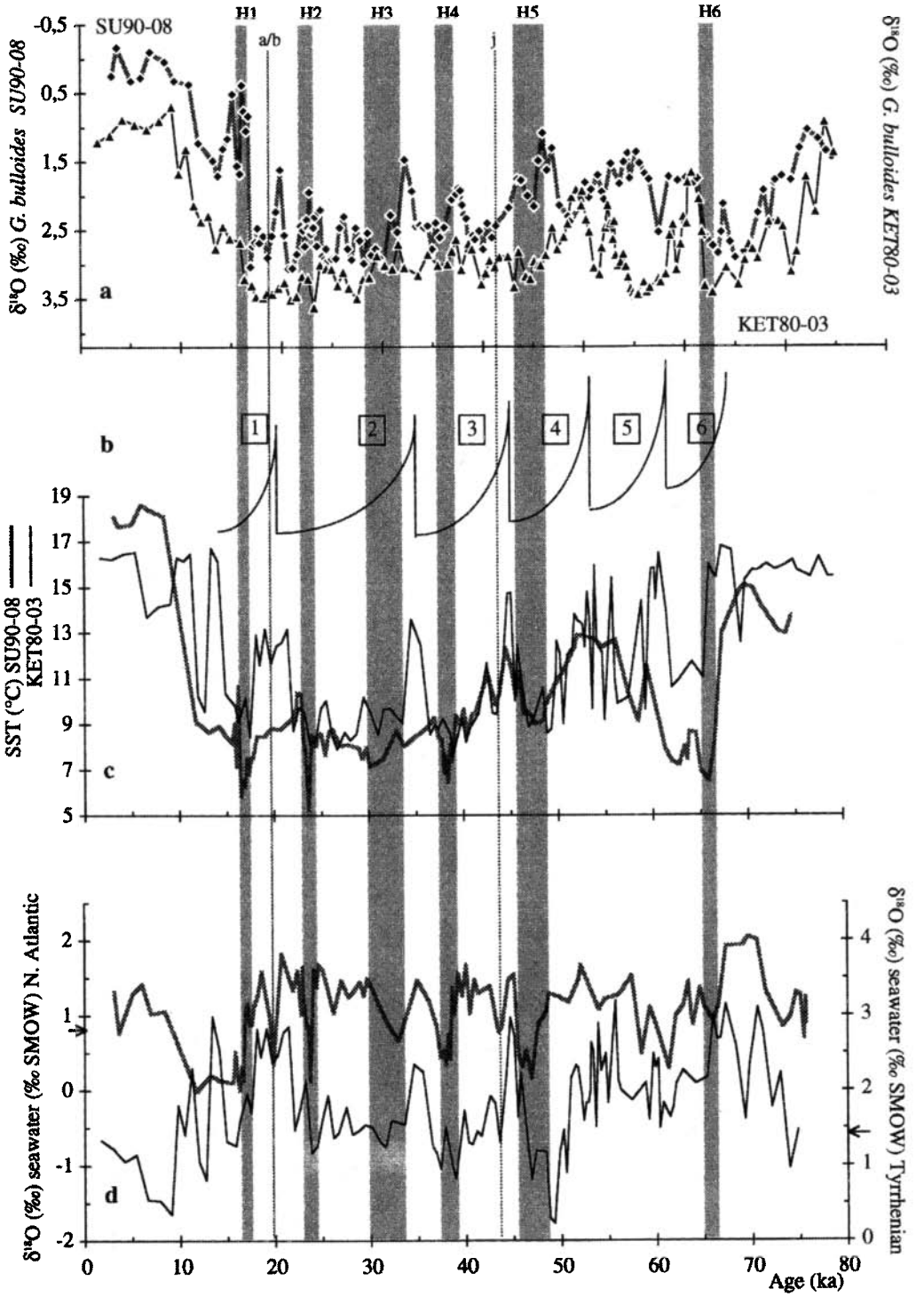

Figure 5. Plot of (a) the $\delta^{18}O$ (‰ PDB) record of *G. bulloides* in cores SU 90-08 (dotted line, solid diamonds, left axis) and KET 80-03 (solid line and solid triangles), (b) the long-term cooling cycles from 1 to 6 (see text), (c) the North Atlantic Ocean mean winter-summer SSTs and Tyrrhenian Sea April-May SSTs (degrees celsius), and (d) theδw (‰ SMOW) in cores SU 90-08 (dotted line) and KET 80-03 (solid line) as a function of age in ka. Positions of the Heinrich events are presented (vertical dashed areas). Arrows on the y axis represent the modern δw values (see text).

Table 2. AMS ^{14}C and Calendar ka Ages of the Onset and End of the Heinrich Events in Core SU 90-08 and of the Main δw Changes in Core KET 80-03

	SU 90-08				KET 80-03			
	Depth, cm	AMS ^{14}C (-400 years)	1 σ	Cal Age	Depth, cm	AMS ^{14}C (-400 years)	1 σ	Cal Age
	65	13.47	0.15	16.15	120	11.50	0.24	13.44
H1	75	14.65	0.15	17.30	180	15.00	0.28	17.76
a/b ?	90	16.60	0.18	19.84				
	115	20.10	0.26	22.96	220	19.40	0.43	22.03
H2	130	21.66	0.26	24.22	245	22.40	0.57	24.96
	160	26.70	0.55	29.70				
H3	175	30.40	0.51	33.75	305	30.20	0.72	33.57
	191	32.50	1.00	37.10	320	32.87	0.97	36.49
H4	210	35.73	0.88	39.20	355	35.30	1.80	39.30
j?	232			43.80	407		1.80	44.20
	240			45.50	430		2.70	46.40
H5/S2	255			48.70	475		2.70	50.72
					500		3.00	52.11
MED					520		3.00	53.12
H6	360			66.50	710			65.60

related to the Heinrich events, except for H3 [*Grousset et al.*, 1993; *Cortijo*, 1995]. They are related to cold SSTs and low δw values. They ended with a rapid warming of the surface waters. The ^{14}C ages of H1, H2, and H4 correspond well with previous datings (Table 2) [*Bond et al.*, 1992, 1993; *Andrews and Tedesco*, 1992; *Andrews et al.*, 1994; *Manighetti et al.*, 1995; *Cortijo et al.*, 1997; *Vidal et* al., 1997; *Zahn et al.*, 1997]. The calibrated ages also agree with the ages obtained by the correlation of Greenland Ice Sheet Project 2 (GISP) and North Atlantic Ocean records [*Bond and Lotti*, 1995]. Although comparable considering dating errors, the ages of the H3 boundaries show a systematic aging in core SU 90-08 with respect to datings in other North Atlantic Ocean cores [*Bond et al.*, 1992, 1993]. However, in this core, H3 is defined neither by a peak of susceptibility nor by an increase of ice-rafted debris [*Grousset et al.*, 1993] but by a δw low. Since their discovery several pulses of ice-rafted debris have been recognized in addition to H1-H6 events in both the North Atlantic Ocean [*Bond and Lotti*, 1995; *Elliot et al.*, 1998] and the Norwegian Sea [*Bauman et al.*, 1995; *Fronval et al.*, 1995; *Rasmussen et al.*, 1996]. Thus the δw depletion we noticed at H3 could be synchronous to other previous ice-rafted discharges (f, g, and h events given by *Bond and Lotti* [1995]). Similar δw decreases, unrelated to detritic material, occurred in core SU 90-08 at ~19 and 44 calendar ka (Figures 2, 4 and 5 and Table 2). We tentatively related them to the a/b and j events, respectively, observed in the North Atlantic Ocean cores [*Bond et al.*, 1993; *Bond and Lotti*, 1995].

3.2. Chronological Comparison of the Tyrrhenian Sea and North Atlantic Ocean Hydrological Records

We compared the timing of the SST, foraminiferal $\delta^{18}O$, and δw changes in cores SU 90-08 and KET 80-03 without any stratigraphical adjustment between the two cores. First-order changes of SST and foraminiferal $\delta^{18}O$ values are well marked and nearly synchronous in the two marine areas (Figure 5). By contrast, the timing of second-order $\delta^{18}O$ changes in the two cores is markedly different, for example, from 40 to 50 ka, whereas the associated SST changes are almost synchronous. As a result, the calculated short-term δw changes show a different pattern in the two oceanic regions. The Mediterranean Sea record reveals several long-term δw decreases followed by rapid δw increases. By contrast, the North Atlantic Ocean δw record exibits rapid fluctuations of short duration, as observed in the northeast Atlantic Ocean [*Maslin et al.*, 1995].

In the time interval 15-75 ka the Tyrrhenian Sea sediments recorded long progressive cooling phases, ending by a steep SST increase (cycles 1-6 in Figure 4). Such a pattern is strongly similar to that described in the North Atlantic Ocean as the longer-term cooling cycles [*Bond et al.*, 1993] (Figure 5). The North Atlantic Ocean SST warmings systematically lead the Tyrrhenian Sea ones after H4, H2, and H1. Nevertheless, the time offsets are ~1.2 ± 1.5, 1.2 ± 0.55, and 1.0 ± 0.48 ka, respectively, (at 1σ), which lie within uncertainties linked to dating, time resolution of the sampling interval, and bioturbation effects. Moreover, these offsets could also be due to variations of the

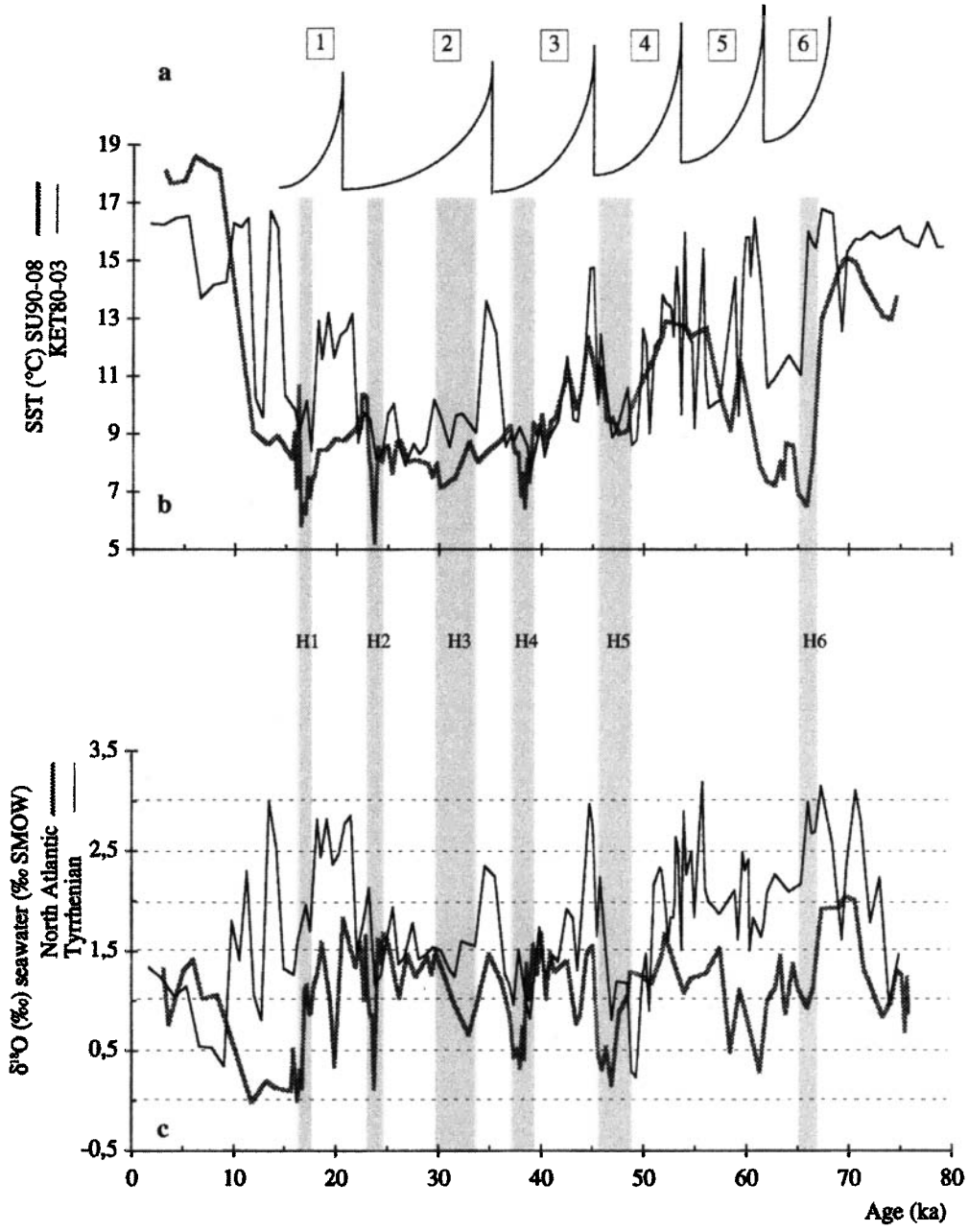

Figure 6. Plot of (a) the long-term cooling cycles from 1 to 6 (see text), (b) the North Atlantic Ocean mean winter-summer SSTs and Tyrrhenian Sea April-May SSTs (degrees celsius), and (c) the δw (‰ SMOW) in cores SU 90-08 (dotted line) and KET 80-03 (solid line) as a function of age in ka. Positions of the Heinrich events are presented (vertical dashed areas).

reservoir age of the North Atlantic Ocean surface waters [*Bard et al.*, 1993]. Considering the overall dating uncertainties, we estimated that the observed chronological shifts between the two SST records are not significant. Furthermore, we did not perform any stratigraphical adjustment between the two cores according to the good similarity of the two SST patterns. We therefore assume that the end of the Heinrich events in both the Mediterranean Sea and the North Atlantic Ocean coincides with the rapid step of SST increase, and the Heinrich event corresponds to the preceeding cold time interval.

3.3. Hydrology of the Mediterranean Sea at the Time of the North Atlantic Ocean Heinrich Events

In the Mediterranean Sea the SST cycles match long-term sea surface water freshening phases by contrast with the North Atlantic Ocean δw patterns (Figure 5). During the past 75 kyr the δw values in core SU 90-08 increased to a mean δw value of ~1.5‰. Short-term excursions toward depleted δw values occur during the Heinrich events with a longer-term one during the deglaciation. The Mediterranean Sea record is rather

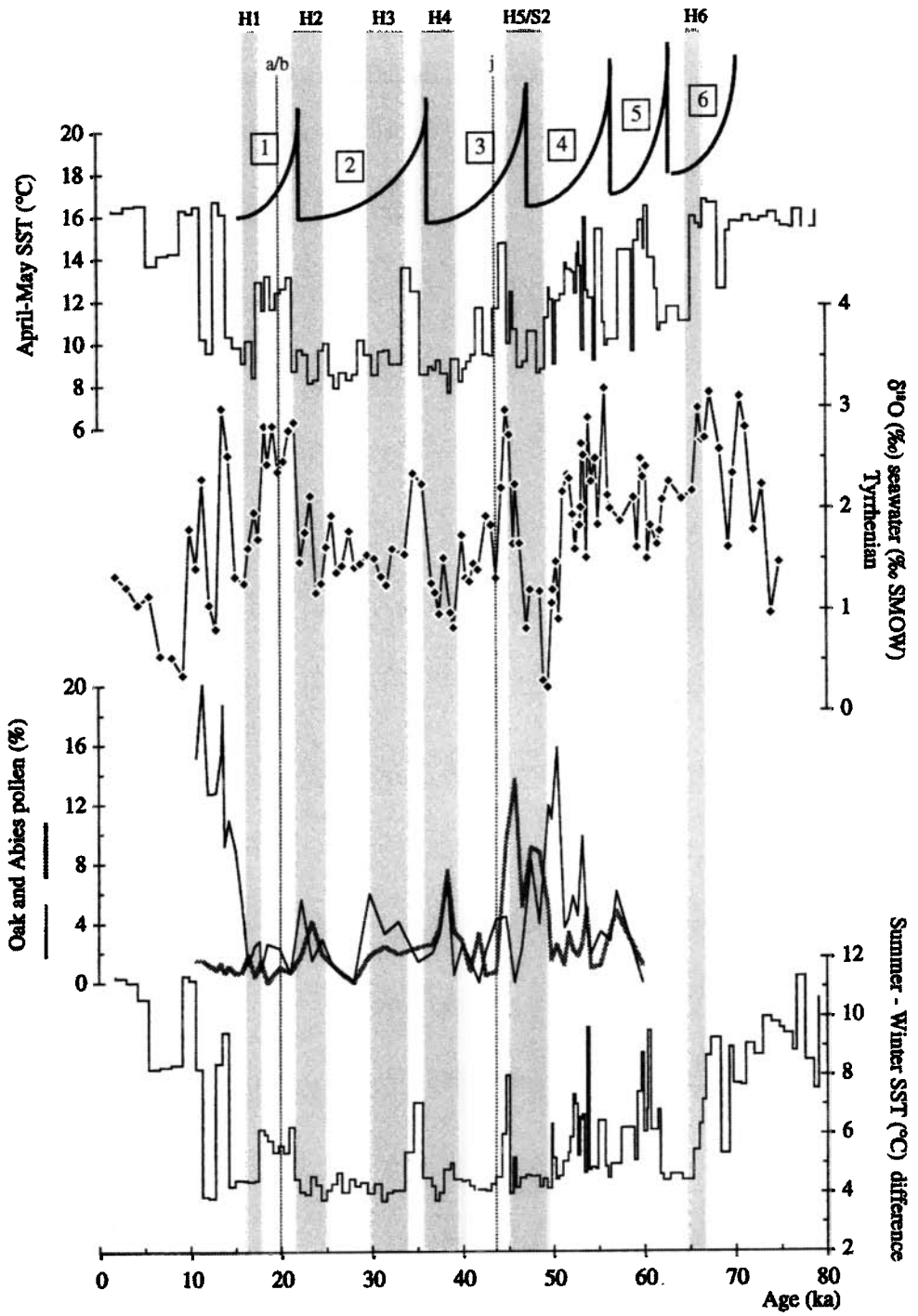

Figure 7. Plot of (a) the long-term cooling cycles from 1 to 6 (see text), (b) April-May SST (degrees celsius), (c)δw (‰ SMOW), (d) Oak (shaded line) and fir tree (solid thin line) pollen abundance (per cent). (Fir tree pollen abundance (per cent) was divided by 3 for graphic convenience), (e) summer to winter SST difference (degrees celsius) as a function of time (ka) in Tyrrhenian Sea core KET 80-03. Position of the Heinrich events, as defined in the text, is also presented.

characterized by several δw excursions of large increasing amplitude from the modern value, except at ~8 and 50 ka. The warm to cold SST transitions are marked by gradual δw lowerings of the Mediterranean Sea surface waters whereas these periods correspond to short excursions of low δw in the North Atlantic Ocean surface waters. Nevertheless, the Mediterranean Sea surface waters recorded low δw values at the time of the Heinrich events and therafter a rapid δw increase, except after H3, similar to that observed in the North Atlantic Ocean.

The δw value was much lower in the Tyrrhenian Sea than in

the North Atlantic Ocean at ~50 ka. The δw difference between both basins was similar to that observed at ~9 ka during sapropel S1 [*Kallel et al.*, 1997b]. In the Tyrrhenian Sea the δw decrease slightly preceeded the H5 one by some 2 kyr and could thus be considered as fully contemporaneous taking into account age uncertainties (Figure 5 and Table 2). Nevertheless, several authors reported the presence of a discrete sapropelic layer S2 at the beginning of stage 3 [*Ryan*, 1971; *Cita et al.*, 1977; *Vergnaud-Grazzini et al.*, 1977, 1986]. Moreover, in core KET 80-03, pollen concentration was augmented at least by a factor of 10 at 50 ka [*Rossignol-Strick and Planchais*, 1989], which would indicate a reduction of bottom water oxygenation as was observed during sapropel S1 [*Rossignol et al.*, 1992; *Combourieu-Nebout et al.*, 1998]. At the time of S1 the δw departure from the modern value in the Mediterranean Sea is ~0.8‰, and salinity may thus be estimated at ~36, close to the salinity estimate at 36.5 in core SU 90-08 at the same time. Such a salinity pattern would not imply an antiestuarine circulation of the Mediterranean Sea waters, which is neither deduced from benthic $\delta^{13}C$ records in cores off Portugal [*Zahn and Sarnthein*, 1987; *Zahn et al.*, 1997] nor from salinity and density gradients between the Mediterranean Sea and the North Atlantic Ocean off the Gibraltar Strait [*Duplessy et al.*, 1992; *Kallel et al.*, 1997a, b]. Hydological changes, observed in the central North Atlantic Ocean core SU 90-08, may differ from those affecting surface waters off the Gibraltar Strait, as observed during the Last Glacial Maximum [*Duplessy et al.*, 1991]. Thus part of the large seawater freshening in the Tyrrhenian Sea at 50 ka may be related to S2 preceeding H5.

For the modern conditions the Mediterranean Sea acts as a concentration basin (**P + R < E**), and the modern δw difference between the two marine sites is estimated at ~0.6 ‰. At the time of Heinrich events H2-H6 (20-70 ka) the positive difference between the Mediterranean Sea and the North Atlantic Ocean was nearly the same as today (Figure 6). The Mediterranean Sea δw changes may thus be considered as a direct response to the inflowing North Atlantic surface waters. Consequently, the modern Mediterranean Sea freshwater balance was not affected at the time of the Heinrich events. During the warm intervals, which follow these events, except after H3, the δw difference was larger than the modern one because of a larger δw increase in the Mediterranean Sea than in the North Atlantic Ocean. Thus the freshwater budget **P + R - E** decreased as a result of either enhanced evaporation or decreased precipitation and runoff after the Heinrich events. During the transitional periods from warm to cold climatic conditions (cycles 4, 3 and 2) the δw difference between the two marine regions becomes close to zero. As the δw values in the North Atlantic Ocean at core site SU 90-08 seat quasisteadily above the modern δw value, the Mediterranean Sea freshwater budget increased either by enhanced precipitation plus runoff or by decrease of the evaporation. During long-term cooling cycle 1 the δw difference was nearly constant from 18 to 16 ka and higher than the modern one at ~1.3‰. During this period the North Atlantic hydrological changes were almost transmitted to the Mediterranean Sea, the freshwater budget of which was unchanged, as previously noted [*Kallel et al.*, 1997b].

4. Discussion

Pollen analyses in core KET 80-03 [*Rossignol-Strick and Planchais*, 1989] permit direct comparison of the continental and oceanic climatic changes at the time of the Heinrich events in the Mediterranean area (Figure 7). Among the tree species the dominant types in core KET 80-03, except *Pinus*, are the deciduous oak, *Quercus robur pubescens*, and the fir tree, *Abies*, over the 10-60 ka time span. The modern analogues of the pollen spectra in core KET 80-03 are located in southeastern Europe and middle Asia, and they primarily respond to moisture availability [*Rossignol-Strick and Planchais*, 1989]. Oak is characteristic of the temperate cool forests of the subhumid and humid Mediterranean altitudinal zones at 200-1300 m, with precipitation ranging from some 600 to 1200 mm yr^{-1}. Fir trees belong to the cool humid mountainous forest above the deciduous forest.

During the deglaciation, from 10 to ~18 ka, oak pollen percentages increased more rapidly than the fir tree percentages. Oak increase is correlated to enhanced SSTs, to high seasonnal contrast, expressed as summer and winter SST difference, and to high sea surface salinity (Figure 7). Inversely, below 18 ka, the maximum abundances of oak and fir tree pollen are associated with the warm to cold transitions and to the Heinrich cold climatic events. They thus coïncide with periods of a larger freshwater budget than the modern budget and with a modern-type freshwater budget, respectively. The warm post-Heinrich intervals are marked by tree pollen disappearance, when the Mediterranean Sea surface waters recorded a decrease of the freshwater budget. Thus moisture availability, rather than temperature improvement, governs the development of these tree species, as was previously suggested [*Rossignol-Strick and Planchais*, 1989]. During cold or cooler climatic conditions, model experiments reproduce persistance of the mixed forest (coniferous/temperate deciduous) by a combination of cold winters and cool cloudy summers [*Prentice et al.*, 1992]. Such a model is consistent with the Tyrrhenian Sea palynological and hydrological patterns and with the association of oak and fir tree pollen to periods of low seasonal thermal contrast.

In North America, pollen analyses in a well-dated sequence of Florida revealed moisture increase at the time of the Heinrich events [*Grimm et al.*, 1993]. In the western and southern Europe pollen series, chronology of the climatic changes linked to these events is less constrained [*Guiot et al.*, 1993; *Watts et al.*, 1996a, b]. Rapid increase of arboreal pollen, as seen in the Monticchio Lake in southern Italy, is related to a climatic improvement with temperature and moisture increases and correlated to the post-Heinrich SST warming in the North Atlantic Ocean [*Watts et al.*, 1996a]. Although exhibiting similar a alternation of high and low abundance of oak pollen, this continental scheme disagrees with the Tyrrhenian Sea palynological pattern since cold and cooler periods coincide with a higher abundance of arboreal pollen and warm intervals coincide with their reduction. In the marine environment such a reduction could be due to changes of pollen preservation. However, pine tree pollen, a very resistant species to degradation, is present all along the core, and oak and fir tree pollen percentages vary independantly of the pine tree pollen sum [*Rossignol and Planchais*, 1989]. It cannot either be related to a change of wind directions as the Tyrrhenian Sea core would receive pollen from the close western and/or eastern continental areas. Finally, peaks of abundance of oak and fir tree could be attributed to increased river pollen transport. However, there are few large rivers around the Tyrrhenian Sea. Consequently, the difference between the continental and marine palynological records is likely related to uncertainties on the chronological position of the Heinrich episodes in the continental areas and to

an overinterpretation in term of temperature of the pollen profiles.

During the Heinrich events the North Atlantic Ocean hydrological changes are directly transferred into the Mediterranean Sea. Massive ice discharges to ~40°N into the North Atlantic Ocean led to cold climatic conditions without significant changes of the modern freshwater budget (**P + R < E**) in the Mediterranean area. During the post-Heinrich times the North Atlantic Ocean hydrological changes are amplified in the Mediterranean Sea. The Mediterranean climate was characterized by warm but more arid conditions than today (**P + R << E**). During the transitions the hydrological changes in the two marine areas are disconnected. The Mediterranean Sea freshwater budget increased (**P + R >> E**), while the North Atlantic Ocean hydrology did not vary greatly. Thus, except during the Heinrich events, oceanic circulation changes in the North Atlantic Ocean cannot solely explain the Mediterranean climatic changes by themselves, and reorganization of the atmospheric circulation may be envisaged. Today, the Mediterranean climate is seasonally balanced from a moderate influence of Atlantic lows in winter to the dominance of subtropical highs in summer, so that precipitation and runoff do not equilibrate the dominant evaporation processes. Similar atmospheric conditions may be envisaged at the time of the Heinrich events in the Mediterranean Sea because of the observed modern-type freshwater buget. The changes toward aridity in the post-Heinrich periods may be due to the predominance of the subtropical highs at the mediterranean latitudes, along with a northward latitudinal shift of the Atlantic depressions tracks. Thereafter, during the transitions the Mediterranean area would have been mainly situated on the track of the Atlantic lows, providing cooler and milder conditions. Thus the increase of precipitation before the Heinrich events, at the very northern latitudes of the North Atlantic Ocean and then at northern latitudes, could contribute to the growth of the different ice sheet involved in the massive iceberg discharge. This could, in turn, provide new massive ice sheet calving due to mass instability. Thus one of the causes of ice sheet instability might be searched in the atmosphere toward air masses and water vapor changes [*Broecker*, 1995].

5. Conclusion

During the past 15-75 kyr the Mediterranean Sea surface waters underwent several long-term cooling cycles, ending by abrupt SST increases. This pattern is similar to the SST changes associated to the massive discharge of icebergs into the North Atlantic Ocean, the Heinrich events. The post-Heinrich SST increases in the two oceanic areas are considered synchronous as the overall dating uncertainties do not permit a check of a possible lead of the North Atlantic Ocean SST changes with respect to the Mediterranean ones. By contrast to the open ocean record, the Mediterranean Sea SST cycles matched coeval long-term periods of salinity lowering. The hydrological and palynological records show a coherent pattern of the climatic changes in the Mediterranean area. At the time of the Heinrich events the North Atlantic Ocean hydrological changes were directly transferred to the Mediterranean Sea. Presence of the ice armada at 40°N in the North Atlantic Ocean did not affect the Mediterranean freshwater budget, and the Mediterranean Sea acted as a concentration basin as today. Despite the cold climatic conditions, persistance of a mixed forest of coniferous and temperate deciduous trees is observed, likely because of cooler and more humid summers as indicated by the low seasonnal thermal contrast in the Mediterranean regions. The end of the Heinrich events was marked by a salinity increase of larger amplitude in the Mediterranean Sea in the two oceanic areas. The Mediterranean Sea freshwater budget (**P + R << E**) consequently decreased, and the resulting aridity limited forest development. During the warm to cold transitions, while the North Atlantic Ocean hydrology at 40°N did not greatly change, an increase of the freshwater budget was observed in the Mediterranean Sea along with forest growth. These climatic changes from dry to humid conditions could be related to latitudinal shifts of the Atlantic depression tracks over the Mediterranean area with respect to ice sheet growth and decay.

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M. Arnold, E. Cortijo, J.C. Duplessy, M. Fontugne, L. Labeyrie, and M. Paterne, Laboratoire des Sciences du Climat et de l'Environnement, Unité mixte CNRS-CEA, 91198 Gif sur Yvette, France. (Martine.Paterne@lsce cnrs-gif.fr)

N. Kallel, Département des Sciences de la Terre, Université de Sfax, Laboratoire E08/C10, Route de Soukra, 3038 Sfax, Tunisia.

M. Rossignol-Strick, Laboratoire de Paléontologie et Palynologie, Box 106, Université Pierre et Marie Curie, 75005 Paris, France.

M. Vautravers, Département de Géologie et Océanographie, Université de Bordeaux I, 33405 Talence, France.

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