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Hydrological changes in the Mediterranean Sea over the last 30,000 years

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[1] Sea surface temperatures were reconstructed over the last 30,000 years from alkenone paleothermometry (SST-$_{alk}$) and planktonic foraminifera assemblages using the Modern Analog Technique (MAT) (SST-$_{foram}$) along two cores of the Mediterranean Sea: MD84-632 (Levantine basin) and MD04-2797 (Siculo-Tunisian Strait). Oxygen isotope of planktonic foraminifera *G. bulloides* for core MD04-2797 and *G. ruber* for core MD84-632 were also determined. SST-$_{alk}$ in the Levantine basin indicate colder values at the Last Glacial Maximum (LGM) ($\sim$14°C) than earlier established from MAT, and a cooling amplitude of 6°–7°C, comparable to the central Mediterranean Sea. Climatic events such as the Younger Dryas (YD) and Heinrich events 1 and 2 (H1 and H2) were times of significant cooling in the two cores. In the Eastern basin, values of local seawater oxygen isotope, δw, indicate relatively saltier waters during the LGM and deglaciation than today, with increasing δw values (higher salinity) in the Eastern basin and decreasing ones (lower salinity) in the central Mediterranean Sea, during cold stadials. The observed alterations of surface water properties (T and δw) in the central and eastern Mediterranean at the LGM are consistent with model experiments showing slightly lower evaporation in the Mediterranean than today, except for the Eastern basin.

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1. Introduction

[2] Precise reconstruction of past temperature and hydrological changes in the Mediterranean Sea area has been the target of numerous studies. This region is of particular interest for understanding the linkages between climate changes and the progressive establishment of urban settlements and the Neolithic cultures. Marine sediments provide continuous and spatially integrated records of the surface hydrology which can be compared to other continuous continental paleorecords such as lake level reconstructions or oxygen isotope records in speleothems [*Bar-Matthews et al.*, 2003; *Robinson et al.*, 2006]. Several approaches based on micropaleotological and isotope analyses of planktonic foraminifera have been applied along marine cores to estimate paleotemperatures and paleosalinities. However, accurate prediction of glacial Sea Surface Temperatures (SSTs) has been hampered by the absence of modern analogs to be compared with the fossil foraminifera fauna found in glacial Mediterranean sediments, in particular in the Eastern basin [*Kallel et al.*, 1997a; *Hayes et al.*, 2005]. The co-occurrence of warm (*Globigerinoides ruber*) and polar (*Neogloboquarina pachydrerma dextra*) foraminifera species in glacial fauna assemblages, which cannot be explained by seasonality changes today, introduces a strong bias toward warmer SSTs. In the few locations where the Modern Analog Technique (MAT) has been applied, SST reconstitutions for the Last Glacial Maximum (LGM) suggest a large W-E temperature gradient resulting from the strong contrast between the cold western Mediterranean Sea and the relatively warmer Levantine basin waters, the latter being only 1°–2°C cooler than today [*Kallel et al.*, 1997a; *Hayes et al.*, 2005]. However, using alkenone paleothermometry, *Emeis et al.* [2000, 2003] calculated a stronger glacial cooling (5°–6°C) in the Levantine basin (OPD 967), though relying on few measurements of glacial (but not LGM) sediment horizons. Furthermore, land temperatures calculated from Soreq cave (Israel) speleothem fluid inclusions suggest that LGM were even colder than 6°C [*McGarry et al.*, 2004].

[3] To improve our understanding of Mediterranean glacial climate, we reconstructed the SST record of two cores, in the Levantine basin (MD84-632) and Siculo-Tunisian Strait, central Mediterranean Sea (MD04-2797). Oxygen isotopes were also determined in planktonic foraminifera in the two cores to derive changes in the oxygen isotope composition of seawater (δ_w), a *proxy* for salinity, to examine the alteration of freshwater input/evaporation budget. This study specifically addresses the effects of rapid climate changes associated with cold millennial-scale events of the last deglaciation, i.e., Heinrich events (H) and Younger Dryas (YD) on the Mediterranean climate.

2. Methods

[4] The two sediment cores, MD04-2797 (36°57N, 11°40E, 771 m water depth) and MD84-632 (32°47N, 34°22E, 1425 m water depth) have been analyzed and compared to several other cores along a west-east transect which includes the Atlantic Ocean and the western and eastern Mediterranean basins (Figure 1).

[5] Detailed oxygen isotope records, expressed in ‰ versus VPDB (Vienna Pee Dee Belemnite standard) were obtained on planktonic foraminifera *Globigerinoides ruber* on MD84-632, and *Globigerina bulloides* on MD04-2797. VPDB is defined with respect to NBS19 calcite standard [*Coplen*, 1988]. For both species, between 6 and 20 shells were picked in the 250–315 μm size range. Analyses were performed at LSCE on Finnigan Delta+ and MAT251 mass-spectrometers. Linearity was corrected following *Ostermann and Curry* [2000]. The mean external reproducibility (1σ) of carbonate standards is

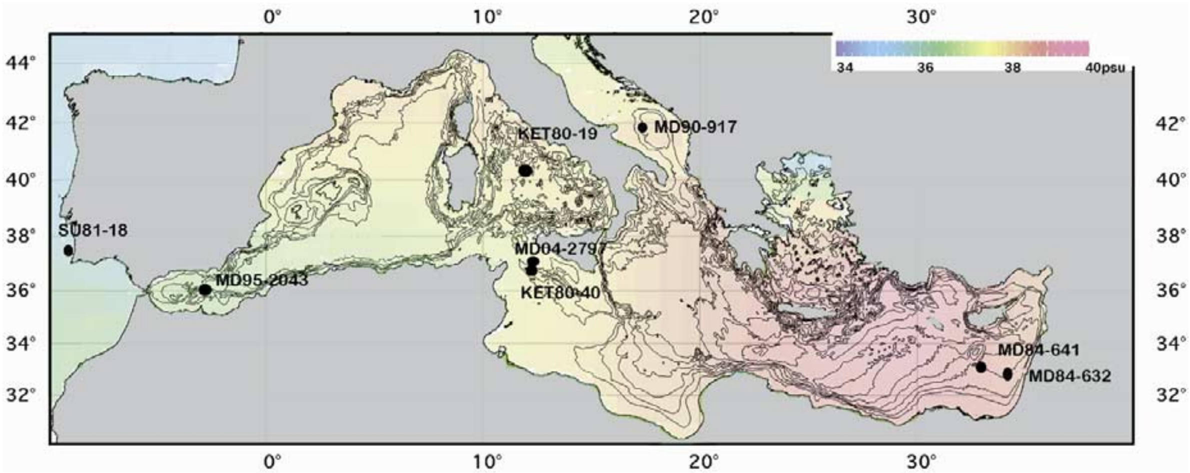

Figure 1. Map of the Mediterranean Sea showing the locations of the different cores used in this study. Background colors index modern surface salinity distribution (in PSU).

±0.05‰, measured NBS18 δ^{18}O is −23.2 ± 0.2‰ VPDB.

[6] SSTs were first determined using planktonic foraminifera assemblages. Each foraminifera sample (>150 μm fraction) was split into 300–1000 individuals for identification and counting. We applied the modern analogue technique (MAT) [*Huston*, 1979; *Prentice*, 1980; *Prell*, 1985; *Kallel et al.*, 1997a] to derive SSTs and associated dissimilarity coefficients. The latter parameter measures the mean distance between each fossil assemblage and the ten best modern analogues, which were selected by statistical analysis to generate SST estimates. The reference database used for the study is composed of 253 core-top sediments, 130 from the Mediterranean Sea and 123 from the Atlantic Ocean [*Kallel et al.*, 1997a, 1997b]. For fossil samples with good modern analogues in the reference database, the dissimilarity is generally <0.25 [*Prell*, 1985]. Above this value, the dissimilarity coefficient indicates that an insufficient number of close modern analogues has been found in the database and SST estimates are discarded. The mean SST error is estimated at ±1°C.

[7] SSTs were also estimated using the C_{37} alkenone unsaturation index, $U^{K'}_{37}$, as defined by *Prahl et al.* [1988], following the experimental procedure described by *Ternois et al.* [1997]. The $U^{K'}_{37}$ index values were converted into temperature using the following calibration published by *Conte et al.* [2006], which include the Mediterranean data set from *Ternois et al.* [1997]:

$$T(^\circ C) = -0.957 + 54.293\left(U^{K'}_{37}\right) - 52.894\left(U^{K'}_{37}\right)^2 + 28.321\left(U^{K'}_{37}\right)^3$$

[8] Briefly, sediment extraction was performed in a mixture of 3:1 dichloromethane: methanol (v/v) in an ultrasonic bath for 15 min and centrifuged for 10 min at about 1500 rpm. Extraction was repeated twice. The three extracts were combined and concentrated to about 1–2 mL by evaporation with a rotary evaporator. The concentrated extracts were then transferred in a 4 mL vial and evaporated to dryness under a nitrogen stream. Alkenones were isolated for the total lipid extract by silica gel chromatography. They were then analyzed on Varian Star 3400CX gas chromatograph equipped with a fused CP-Sil-5CB silica capillary column (50 m × 0.32 mm i.d., 0.25 μm film thickness, Chrompack), a Septum Programmable Injector (SPI) and flame ionization detector (FID). The oven temperature was programmed from 100°C to 300°C at 20°C min^{-1}. Helium was used as the carrier gas (25 mL min^{-1}). Analytical precision obtained after duplicate injections was calculated to be less than 0.01 $U^{K'}_{37}$ unit ratio.

[9] Sea surface salinity as expressed by the local seawater δ^{18}O, δ_w, was determined following the method proposed by *Duplessy et al.* [1991]. δ_w depends on changes of E-P and the mean ocean δ^{18}O resulting from changes in continental ice volume. δ_w variations were calculated by solving the paleotemperature equation [*Shackleton*, 1974] using the SSTs values derived from foraminifera and alkenones.

$$T(^\circ C) = 16.9 - 4.38\left(\delta^{18}O_{calcite} - \delta_w + 0.27\right) + 0.1\left(\delta^{18}O_{foraminifere} - \delta_w + 0.27\right)^2$$

[10] Local $\Delta\delta_w$ changes were then obtained by subtracting the effect of continental ice melting

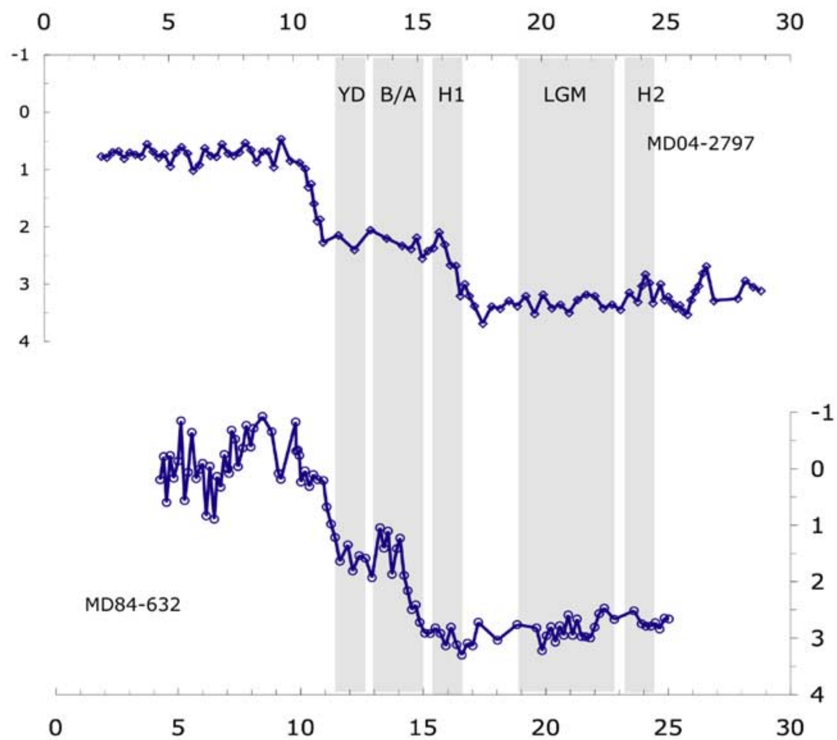

Figure 2. Oxygen isotope values (in ‰VPDB) plotted against age in cal ka B.P. measured in *G. bulloides* along MD04-2797 core in the Siculo-Tunisian Strait (top graph) and *G. ruber* along MD84-632 core in the Levantine basin (bottom graph). Gray vertical bars indicate cold stadials (Younger Dryas, Heinrich 1 and 2 events), the Last Glacial Maximum, and the warm Bølling/Allerød.

on global seawater δ^{18}O. The latter is assumed to be equal to the deglacial sea level curve of *Lambeck and Chappell* [2001] multiplied by a constant coefficient of 1.1‰/130 m from *Waelbroeck et al.* [2002]. We did not convert the $\Delta\delta_w$ values into salinity units because of uncertainty resulting from possible changes in the slope of the δ_w/salinity relationship.

[11] The δ^{18}O value of planktonic foraminifera calcite depends on both SST and δ_w at the time of foraminifera growth. It has been shown that *G. bulloides* main season of production in the Mediterranean Sea is April–May, whereas *G. ruber* develops preferentially in October–November [*Kallel et al.*, 1997a]. *Sicre et al.* [1999] have shown that in the western Mediterranean Sea, the alkenone production peaks in spring and fall, while in the Eastern basin, maximum coccolith fluxes in sediment traps occur in early spring [*Ziveri et al.*, 2000]. Thus alkenone-derived SST values were not used directly to derive δ_w. δ_w was calculated using SSTs derived from MAT and then from alkenones after applying a correction taking into account different seasonality between alkenone production and *G. bulloides* for MD04-2797 core and *G. ruber* for MD84-632, prior to substitution into the paleotemperature equation (see discussion below).

[12] The age model developed for MD04-2797 is based on correlations between micropaleontological abundance curves and foraminiferal isotopic composition with the nearby ^{14}C dated cores KET80-40 (36°47N; 11°49°E, 715 m water depth) (G. Siani, personal communication, 2007), KET80-19 (40°33N; 13°21E, 1920 m water depth) [*Kallel et al.*, 1997a, 1997b] and MD95-2043 (36°08N; 02°37W, 1841 m water depth) [*Cacho et al.*, 1999, 2001]. For MD84-632, the age model was obtained from correlation with planktonic foraminifera abundance curves of the neighboring core MD84-641 (33°02N; 32°38E; 1375 m water depth) [*Fontugne et al.*, 1994]. ^{14}C ages have been translated in calendar years using Calib 5. The detailed methodology will be described by L. Essallami (manuscript in preparation, 2007). Mean uncertainty on ages is estimated to about 500 years. For core MD04-2797, sedimentation rate decreases from 37 cm/1000 years during glacial to 32 cm/1000 years for the Holocene, while for core MD84-632, sedimentation rate increases

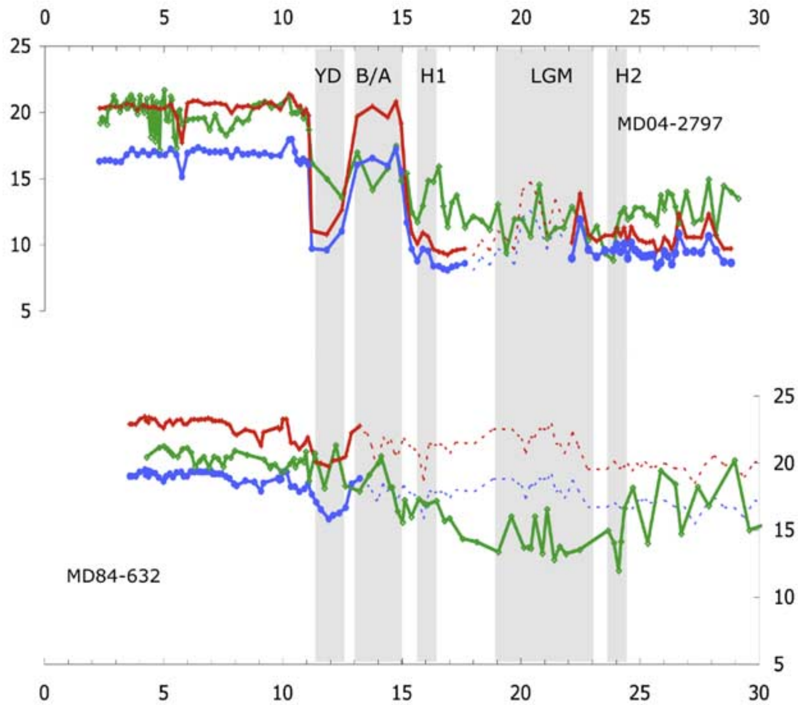

Figure 3. Sea surface temperatures (SSTs), in °C, determined from alkenones (green curve) and planktonic foraminifera assemblages using the Modern Analog Technique (MAT) for April–May (in blue) and October–November (in red) along MD04-2797 (top graph) and MD84-632 (bottom graph). The dotted lines represent time intervals for which the dissimilarity coefficient is >0.25 and SSTs discarded. Gray vertical bars indicate cold stadials (Younger Dryas, Heinrich 1 and 2 events), the Last Glacial Maximum, and the warm Bølling/Allerød.

from a mean glacial value around 7 cm/1000 years to Holocene mean value around 11.5 cm/1000 years.

3. Results

[13] Figure 2 displays the δ^{18}O records of *G. bulloides* and *G. ruber* over the last 30,000 years, respectively along core MD04-2797 (Siculo-Tunisian Strait) and core MD84-632 (Levantine basin). Values are reported in auxiliary material Table S1.[1] The δ^{18}O curves depict a depletion with an amplitude of about 2.5‰ in the Central Mediterranean Sea, and 3.5‰ in the Eastern Basin. These shifts far exceed the single influence of melting continental ice sheets (around 1.05 ± 0.1‰ [*Duplessy et al.*, 2002]), thus indicating large effects due to changes in SST or local salinity. In the Levantine basin, δ^{18}O shows more negative values at the time of last sapropel (S1) formation, consistent with wetter conditions, thus partly accounting for the large isotopic depletion from LGM values. As discussed below, the difference in apparent timing of the deglacial trends is attributed to regional differences in salinity versus temperature changes.

[14] Figure 3 compares SSTs derived from alkenones (SST-$_{alk}$) and MAT for April–May and October–November (SST-$_{foram}$). SST-$_{foram}$ between 22,000–17,600 cal yr B.P. in MD04-2797 core, and prior to 13,000 cal yr B.P. in MD84-632 core were discarded because of high dissimilarity coefficients (dotted lines in Figure 3; values are reported in auxiliary material Table S2). The LGM cooling in MD04-2797 agrees with previous findings for the western and central Mediterranean Sea, around 6°C [*Hayes et al.*, 2005]. In the Levantine basin, glacial SST-$_{alk}$ are also 6°–7°C colder than the upper Holocene (21°C). This cooling amplitude is larger than earlier numbers based on planktonic foraminifera assemblages (1°–2°C), but confirm a previous alkenone estimate from the nearby ODP 967 site (34°04N; 32°43E) [*Emeis et al.*, 2000, 2003]. It is also larger than reported by *Bard et al.* [2000] in the Iberic margin SU81-18 core (37°46N; 10°11W, 3135 m water depth) using alkenones (~5°C).

[15] Known North Atlantic and western European climatic events are recognizable in MD04-2797

[1]Auxiliary materials are available at ftp://ftp.agu.org/apend/gc/2007gc001587.

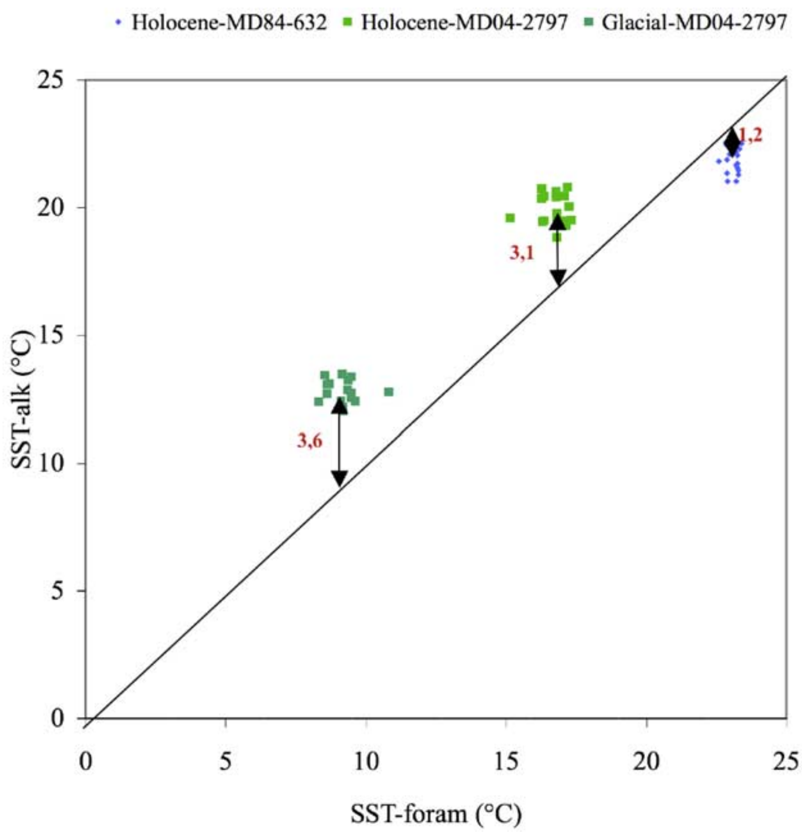

Figure 4. Correction plot for SST-$_{alk}$ (see text). Reported are the differences between SST_{alk} and April–May SST-$_{foram}$ for the Holocene and LGM in core MD04-2797 (central Mediterranean Sea). Also reported are the differences between SST-$_{alk}$ and October–November SST-$_{foram}$ for the Holocene in core MD 84-632 (Levantine Basin).

in the two temperature proxy records: the cold interval Younger Dryas (YD) (centered at about 12,000 cal yr B.P.) and Heinrich events (H1 ~ 16,000 cal yr B.P. and H2 ~ 24,000 cal yr B.P.) as well as the warmer Bølling/Allerød (B/A, 13,000–15,000 cal yr B.P). The two sites indicate that during the B/A, SST-$_{foram}$ rise to values comparable to the Holocene, while SST-$_{alk}$ remain cooler. Also different is the warming of the YD termination starting earlier in the SST-$_{alk}$ than in the SST-$_{foram}$ record (~11,500 cal yr B.P.). Lower than Holocene SST values during the B/A have been documented by alkenones in the Alboran and Tyrrhenian Sea, as well as the shorter YD cooling phase than found in the North Atlantic (~1,200 years) [*Cacho et al.*, 2001]. Except for the duration of the YD, these events are broadly synchronous to the Atlantic Ocean records [*Duplessy et al.*, 1986; *Bard et al.*, 2000].

[16] In order to make precise δ_w estimates in the Mediterranean Sea, SST-$_{alk}$ were substituted in the paleotemperature equation because (1) there was no reliable glacial SST-$_{foram}$ estimates for MD84-632 and for MD04-2797 between 22,000–17,600 cal yr B.P. and (2) SST-$_{alk}$ were acquired at higher resolution than SST-$_{foram}$. A temperature correction was applied to SST-$_{alk}$ to account for different productivity seasons between foraminifera and *Emiliania huxleyi* in each basin. Since $\delta^{18}O$ was determined on *G. bulloides* in the central Mediterranean Sea, SST-$_{alk}$ were compared to April–May SST-$_{foram}$, the main production season of this species. Average SST-$_{foram}$ values of 16.7°C for the Holocene (21 samples) and 9.2°C for the glacial period (17 samples) were calculated for core MD04-2797. Average SST-$_{alk}$ of 20°C over the Holocene and 12.8°C for the LGM were calculated along the same core. Thus Holocene and glacial SST-$_{alk}$ were both corrected by −3.5°C to predict spring SSTs (Figure 4). Comparison has been made with the Iberic margin core SU81-18 and Adriatic Sea core MD90-917 (41°17N; 17°37E, 1010 m water depth). In core SU81-18, the mean difference between SST-$_{alk}$ from *Bard et al.* [2000] recalculated using the calibration of *Conte et al.* [2006] and April–May SST-$_{foram}$ for the upper Holocene and LGM values is +0.5°C, as maximum productivity is spring in both cases. For

core MD90-917, the mean LGM difference between SST-$_{alk}$ and spring SST-$_{foram}$ [*Siani et al.*, 2001] is around 3.0°C, thus close to the correction calculated for MD04-2797, within the methodology uncertainty (±1°C). Our data were also compared to SST-$_{alk}$ in core MD95-2043, from the Alboran Sea, and corrected from the same value as in core MD04-2797.

[17] For the Levantine basin, δ^{18}O was measured on *G. ruber*, which develops mainly in October–November. The difference between Holocene average SST-$_{foram}$ ~ 22°C (17 samples) and SST-$_{alk}$ ~ 20.8°C was smaller (+1.2°C) (Figure 4). In absence of reliable glacial SST-$_{foram}$, we assumed the same seasonal shift for the LGM as for the Holocene and applied a correction of +1.2°C to all SST-$_{alk}$ values. This is a reasonable assumption, as coccolith production occurs during early spring, thus at SST slightly colder than fall, *G. ruber* growth season. δ_w was then estimated combining the corrected SST-$_{alk}$ with the δ^{18}O of *G. bulloides* for MD04-2797 and *G. ruber* for MD84-632. For comparison, when possible, the same calculations were performed using SST-$_{foram}$. The accuracy of such reconstruction depends primarily on that of SST estimates. A 1°C error for SST estimates result in a 0.23‰ in the calculated δ_w. An additional error of 0.05‰ is associated with the mass spectrometry measurements of foraminiferal δ^{18}O. This leads to a total analytical error on past δ_w estimates of about +0.3‰ (equivalent to a salinity uncertainty of about 1 PSU).

[18] As can be seen from Figures 5a and 5b, the $\Delta\delta_w$ reconstructions are quite robust during periods of small SST changes, as for the Holocene and the LGM. During rapid SST changes such as deglaciation, results must be interpreted with caution, especially when SST-$_{alk}$ and SST-$_{foram}$ do not match, such as during the B/A. In the Levantine basin, the $\Delta\delta_w$ records reconstructed from SST-$_{foram}$ and SST-$_{alk}$ display similar trends over the last 13,000 cal yr B.P. (Figure 5b). Both $\Delta\delta_w$ curves show a decrease between 10,500 and 6,500 cal yr B.P., which includes the deposition of sapropel S1. Higher $\Delta\delta_w$ values occur ~16,000 cal yr B.P., at the time of H1, and decreased at the onset of the B/A to remain relatively stable during the YD. The central Mediterranean $\Delta\delta_w$ record (Figure 5a) strongly contrasts with the Eastern Basin one, and correlates better with the North Atlantic reconstructions using foraminifera off Portugal [*Duplessy et al.*, 1992]. H1 and H2 are characterized by lowest $\Delta\delta_w$ values suggesting a salinity decrease, while the YD exhibits values comparable to the B/A when calculated from SST-$_{alk}$. During the B/A, higher $\Delta\delta_w$ values were obtained from SST-$_{foram}$ as a result of warmer SST-$_{alk}$.

4. Discussion

[19] The SST and $\Delta\delta_w$ records emphasize the different influence of the North Atlantic climate on the hydrology of the central and eastern Mediterranean areas during the last 25,000 years. Large temperature cooling during cold Heinrich (H1 and H2) and YD events, and the warm B/A and transition to Holocene are recorded in both basins. However, salinity changes, as recorded by $\Delta\delta$w, show opposite temporal variations in the two cores. Cold events coincide with a decreased δw/salinity in the central Mediterranean, similar as the North Atlantic core SU81-18 [*Duplessy et al.*, 1992], whereas in the Eastern basin we observe an apparent δw/salinity increase. Surface water cooling events during H1, H2 and YD can be explained by the reduction of the Atlantic heat transport to the high latitudes which had a great impact over the European continent, whereas the associated depleted δw values in the central Mediterranean Sea likely result from the southward migration of the polar front of the North Atlantic allowing less saline waters to enter the Mediterranean Sea.

[20] Opposite temporal evolution of δw between eastern and central Mediterranean during rapid climate changes of the deglacial suggest that, coeval to the North Atlantic cooling, the eastern Mediterranean underwent higher excess evaporation over precipitation, than today. Our δw reconstruction is consistent with existing continental paleoclimatic record in the Levant region. In a recent compilation of lake level curves, *Robinson et al.* [2006] have shown a lowering of Lake Lisan/Dead Sea system and Lake Tiberias (Jourdan Valley) during H1 and H2. During the YD, halite deposit in the Dead Sea has been interpreted as reflecting extreme arid conditions, while according to the Levantine basin δw curve, H1 was more arid that the YD. The speleothem δ^{18}O record [*Bar-Matthews et al.*, 2003] agrees with the lake levels fluctuations with most positive values during the YD indicating lowest rainfall. On the basis of the elemental ratio of marine sediments, *Calvert and Fontugne* [2001] suggested that intensified winds and dust transport over the Levantine basin could have contributed to glacial aridity.

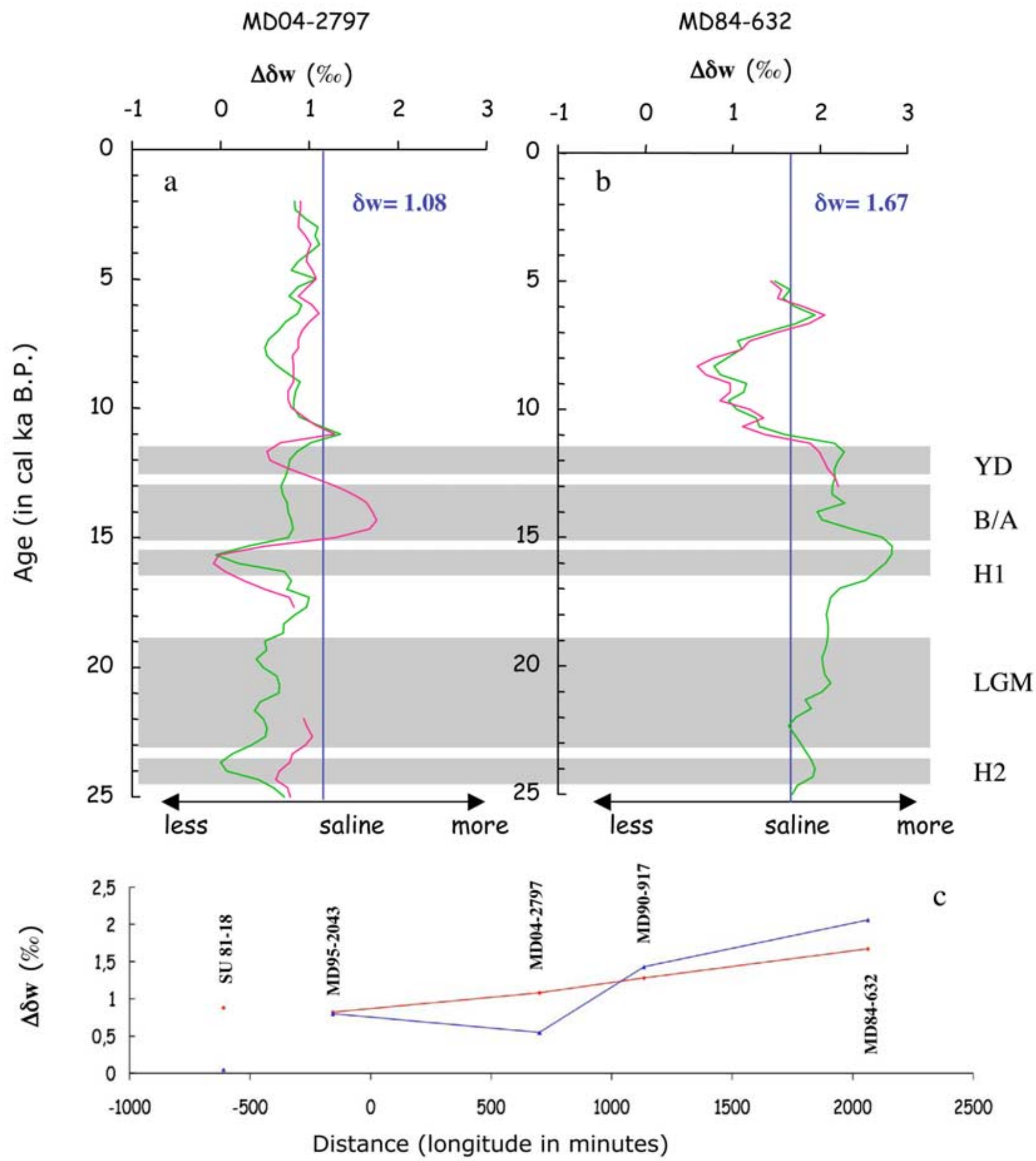

Figure 5. $\Delta\delta$w values calculated for (a) MD04-2797 and (b) MD 84-632 using SSTs derived from MAT (red curves) and alkenones (green curves) after applying a correction taking into account different production season between foraminifera and *E. huxleyi* (shown in Figure 4). Gray vertical bars indicate cold stadials (Younger Dryas, Heinrich 1 and 2 events), the Last Glacial Maximum, and the warm Bølling/Allerød. (c) Longitudinal $\Delta\delta$w changes from the North Atlantic to the Levantine basin at the LGM using five cores, including those presented in this study. Distance between cores is expressed in minutes longitude at 35°N. From west to east: SU81-18, MD95-2043, MD04-2797, MD90-917, and MD84-632. Mean analytical uncertainty on $\Delta\delta$w is ±0.3‰ (approximately equivalent to 1 psu). The red line presents the Holocene gradient; the blue line presents the LGM gradient.

[21] To provide a comprehensive basin-scale view of hydrological changes ($\Delta\delta$w) during the LGM, we reconstructed W-E transects starting in the Atlantic Ocean off Portugal, ending at the MD84-632 site (Figure 5c). The $\Delta\delta$w transect is built with existing records of SST-$_{alk}$ from the North Atlantic (core SU81-18 [*Bard et al.*, 2000], corrected by +0.5°C as discussed earlier), the Alboran Sea (core MD95-2043 [*Cacho et al.*, 2001], corrected by −3.5°C as for core MD04-2797), the Adriatic Sea (core MD90-917 (G. Siani, unpublished data, 2007), corrected by 3.0°C as discussed earlier). The distance between cores along the W-E transect was calculated from their longitude coordinate and

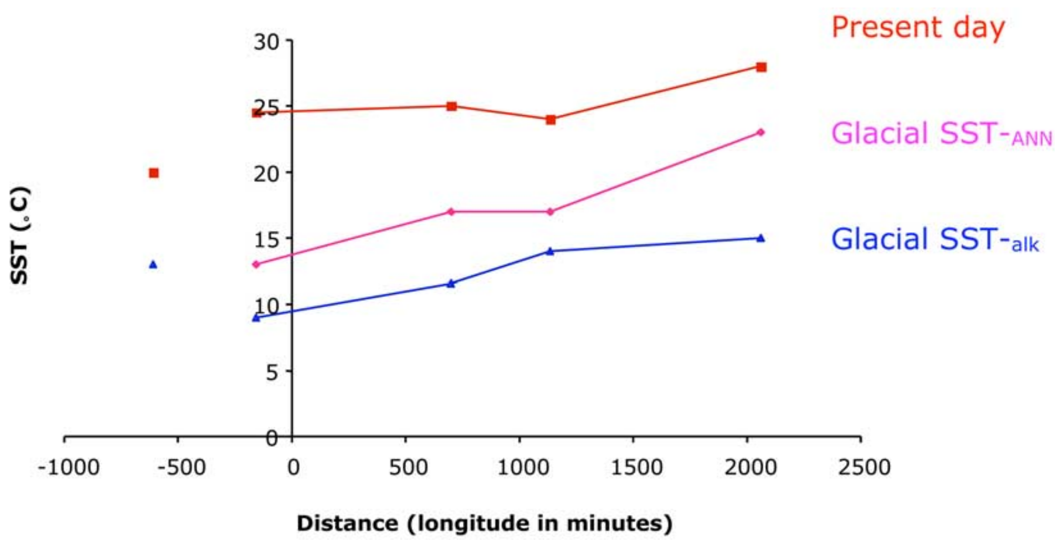

Figure 6. Plot showing west-east gradients of the present-day SSTs (Present day), glacial SSTs derived from artificial neuronal networks (Glacial SST$_{\text{-ANN}}$), and alkenones corrected from seasonality (Glacial SST-$_{\text{alk}}$).

expressed in minute longitude at 35°N. The zero reference corresponds to zero longitude, east of the Gibraltar Strait. The $\Delta\delta$w transect reveals a progressive isotopic enrichment of the surface water from the North Atlantic to the Eastern basin, except for a slight depletion at the MD04-2797. The gradient at the LGM is steeper than today and indicates drastic changes from the central part of the Mediterranean, eastward. While $\Delta\delta$w values in the central Mediterranean basin are slightly lighter than today, the sharp $\Delta\delta$w increase east of the Siculo-Tunisian Strait is clearly indicative of different climatic conditions in the Levantine basin during the LGM. As shown in Figure 6, modern SSTs, glacial SST-$_{\text{alk}}$, and artificial neuronal network summer SSTs (ANN-summer) produced for the whole Mediterranean Sea [*Hayes et al.*, 2005], all indicate an increase eastward. Glacial SST-$_{\text{alk}}$ corrected from seasonality are lower than those determined by *Hayes et al.* [2005], especially in the eastern Mediterranean Sea.

[22] Models of LGM paleoatmospheric circulation [*Kutzbach and Guetter*, 1986; *Jousseaume*, 1993] tend to predict slightly reduced rainfall over most of the Mediterranean Sea. However, attempts to model the circulation of the Mediterranean Sea at the LGM using ocean general circulation models [*Bigg*, 1994; *Myers et al.*, 1998] indicate lower evaporation except for a substantial increase of the E-P balance in the Eastern basin, mainly the southern and eastern Levantine basin. Therefore, even though colder temperatures may account for less evaporation at the LGM, evaporation would have still exceeded precipitation. These simulations are in agreement with our results indicating slightly drier conditions during the LGM in the Levantine basin. However, further modeling experiments would be necessary to explain the contrasting δ_w evolution during the last glacial period between the two core sites, in particular during Heinrich event 1. Glacial surface salinity increase has been observed in other low latitude records from the Atlantic and Indian Ocean, and may be related to changes in atmospheric water vapor gradients between the equator and the poles forced by the large glacial continental ice sheets [*Labeyrie*, 2000].

5. Conclusion

[23] Mediterranean SSTs have been reconstructed along two cores recovered in the Siculo-Tunisian Strait and the Levantine Basin using MAT and alkenone paleothermometry. Our results indicate a general agreement in each core between both SST records when SST-$_{\text{foram}}$ are associated with low dissimilarity coefficients. In the Levantine basin, where no reliable glacial SST-$_{\text{foram}}$ could be reconstructed, SST-$_{\text{alk}}$ indicates that the LGM was colder than today by about 6°–7°C. This cooling amplitude is similar to that obtained in the western and central Mediterranean and significantly larger than that reconstructed using foraminiferal fauna. SST-$_{\text{alk}}$ have therefore been used to estimates the past δw evolution. During the LGM, δw increased from west to east, indicating that the Mediterranean Sea remained, as today, an evaporation basin but the gradient was steeper than present time. Our findings are coherent with model simulations of the Mediterranean climate at the LGM, but the

enhanced aridity reconstructed during cold stadials in MD84-632 and continental records of the Levant region contrasting with the δw evolution in the central basin, remains unexplained.

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