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Temperature and salinity variations of Mediterranean Sea surface waters over the last 16,000 years from records of planktonic stable oxygen isotopes and alkenone unsaturation ratios

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Abstract

Alkenone unsaturation ratios and planktonic $\delta^{18}\text{O}$ records from sediment cores of the Alboran, Ionian and Levantine basins in the Mediterranean Sea show pronounced variations in paleo-temperatures and -salinities of surface waters over the last 16,000 years. Average sea surface temperatures (SSTs) are low during the last glacial (averages prior to 13,000 years: 11–15°C), vary rapidly at the beginning of the Holocene, and increase to 17–18°C at all sites during S1 formation (dated between 9500 and 6600 calendar years). The modern temperature gradient (2–3°C) between the Mediterranean sub-basins is maintained during formation of sapropel S1 in the Eastern Mediterranean Sea. After S1, SSTs have remained uniform in the Alboran Sea at 18°C and have fluctuated around 20°C in the Ionian and Levantine Basin sites. The $\delta^{18}\text{O}$ of planktonic foraminifer calcite decreases by 2‰ from the late glacial to S1 sediments in the Ionian Basin and by 2.8‰ in the Levantine Basin. In the Alboran Sea, the decrease is 1.7‰. Of the 2.8‰ decrease in the Levantine Basin, the effect of global ice volume accounts for a maximum of 1.05‰ and the temperature increase explains only a maximum of 1.3‰. The remainder is attributed to salinity changes. We use the temperature and salinity estimates to calculate seawater density changes. They indicate that a reversal of water mass circulation is not a likely explanation for increased carbon burial during S1 time. Instead, it appears that intermediate and deep water formation may have shifted to the Ionian Sea approximately 2000 years before onset of S1 deposition, because surface waters were as cold, but saltier than surface water in the Levantine Basin during the Younger Dryas. Sapropel S1 began to form at the same time, when a significant density decrease also occurred in the Ionian Sea. © 2000 Elsevier Science B.V. All rights reserved.

Keywords: Mediterranean Sea; sapropel S1; last glacial maximum; temperature; salinity; density

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1. Introduction

The Mediterranean Sea is the source of an important intermediate water mass in the Atlantic Ocean. The density of outflowing water is affected by temperature and salinity variations in the Mediterranean Sea, and variations in the amount and character of outflowing water may influence the global thermohaline circulation system (Bigg, 1994; Zahn et al., 1987). Sediment records in the Eastern Mediterranean Sea show that such variations induced by periodic climatic changes have occurred since at least the Pliocene: they resulted in recurring anoxic conditions in the Mediterranean Sea, during which dark-colored and organic carbon-rich sediment layers were formed (Cita et al., 1977; Thunell et al., 1984). These sapropels may be well-constrained and young analogs of organic carbon-rich sediments found in older geological periods (Ryan and Cita, 1977) and their occurrence signals profound changes in the biogeochemical system of the Mediterranean Sea under natural conditions (Olausson, 1961). The last of these events (S1) is only approximately 9000 years old and was deposited when civilizations emerged around the Mediterranean Sea after the climatic amelioration following the last glacial. What were climate patterns at that time, and how did the Mediterranean Sea react to the changing climatic conditions (Stanley and Galili, 1996)? Study of the sapropels and their climatic and hydrographic causes also permits us to identify the range of natural variability in the Mediterranean environment; it is then possible to evaluate the added impact of anthropogenic influence (Sarmiento et al., 1988; Béthoux, 1989; Rohling and Bryden, 1992; Roether et al., 1996).

Our aim here is to examine changes in the character of sea surface water masses in the Mediterranean Sea that may illustrate the change in the Mediterranean Sea following the last glaciation. Switches of the depositional mode between the sapropel (S1 and its predecessors) and non-sapropel state have most likely resulted from changes in temperature and salinity of surface waters, which are in turn sources for intermediate and deep waters (Olausson, 1961; Vergnaud-

Grazzini et al., 1977; Thunell and Williams, 1989). Most of the hypothetical scenarios for sapropel deposition may be directly linked to a change in the water balance and specifically to a change in the ratio of evaporation (e) over precipitation plus river runoff (p) (Béthoux, 1993; Rohling, 1994). Increased precipitation, river runoff, or melting of ice during warm and moist climatic conditions are all caused by maximal insolation (Cita et al., 1977; Rossignol-Strick et al., 1982; Hilgen, 1991) and are known from land records to accompany sapropels (Rossignol-Strick, 1993; Mommersteeg et al., 1995).

In a previous paper, we have shown that sea surface conditions (temperature T and salinity S) were not uniform for all sapropels (Emeis et al., 1998), whereas the external forcing mechanism (insolation) is rhythmic and may have been the same for all events (Rossignol-Strick, 1985). To clarify some aspects of the switch from non-sapropel to sapropel state in the Mediterranean Sea, we attempt here to reconstruct gradients in the physical properties of the sea surface waters for time slices bracketing sapropel events. Because it is easiest to start from modern conditions, sapropel S1 is a logical first candidate to validate this approach. In the course of this validation, we test some of the hypotheses outlined below. They have been formulated by early workers in the field and are aimed at explaining paleoenvironmental change in the Mediterranean Sea leading to sapropels. The hypotheses, which in some respects are interrelated, are schematically shown in Fig. 1.

(1) *Initial requirement for sapropel formation is a density decrease in the sea surface of the Eastern Mediterranean Sea; sapropel S1 is triggered by warming combined with an input of freshwater. These two effects lowered the surface water density in the areas of intermediate and deep water formation to an extent which precluded deep sinking of oxygenated surface waters.*

This hypothesis links sapropels to anoxia that results from a lack of oxygen recharge of the sub-thermocline water masses. The deep and intermediate water masses are formed when water denser than surface water sinks below the zone of photo-

synthesis and gas exchange with the atmosphere. If no new dense water is formed, oxygen in the old intermediate and deep waters will be quickly depleted (estimated as approximately 600–700 years; Rohling, 1994) and the depth range which they occupy (depending on their volume and their relative densities to each other and to surface water) will become anoxic. The rearrangement of water masses, of surface areas occupied by relatively dense or light surface waters, and depth intervals occupied by waters of a particular density is a physical phenomenon. However, such a rearrangement may carry a variety of consequences for the chemical and biological environment (discussed by many authors, e.g. Rohling and Gieskes, 1989; Calvert et al., 1992; Fontugne and Calvert, 1992; Howell and Thunell, 1992) and may have contributed to sapropel formation by storing nutrients below the euphotic zone, by increasing the productivity or by enhancing preservation of organic matter. For the sake of simplicity, our treatment of Hypothesis 1 explicitly neglects the important chemical and biological aspects of sapropel deposition.

The density anomaly of water, expressed as σ_t (kg/m^3), is dependent on salinity S and temperature T of a given body of seawater. It is calculated from T and S according to the formula given by UNESCO (1981). Besides faunal assemblages indicating low salinity conditions (Ryan, 1972; Vergnaud-Grazzini et al., 1977; Cita et al., 1977), pronounced decreases of $\delta^{18}\text{O}$ values of planktonic foraminifer calcite at the bases of and in sapropels (Cita et al., 1977; Williams and Thunell, 1979) suggest a decrease in surface salinity immediately prior to sapropel deposition. Decreasing salinity requires a change in e/p from modern conditions, most likely by increased discharge from land or by increased advection from the Atlantic Ocean.

The second factor determining the density of seawater is temperature. Most sapropels of late Pleistocene age occur during climatic warming phases or interglacials (Cita et al., 1977) and a good correlation exists between orbitally-forced insolation maxima and sapropels (Rossignol-Strick, 1985). Quantifying the temperature change associated with or preceding sapropel formation is possible either by micropaleontological, isotopic,

or organic geochemical methods. In the present study, we make use of the alkenone unsaturation ratio that reflects temperature during growth of prymnesiophyte algae (Brassell et al., 1986).

(2) *The combined effects of salinity and temperature change on surface water density prior to and during sapropel deposition are larger in the Eastern than in the Central and Western Mediterranean. The spatial variation in the contribution of salinity to this change may be indicative of a point source of freshwater.*

This hypothesis follows observations by numerous authors that the decrease in $\delta^{18}\text{O}$ of planktonic foraminifers associated with sapropel formation is larger in the Eastern than in the Western Mediterranean. Under discussion as sources for the freshwater are point sources such as the Nile River (Rossignol-Strick et al., 1982, 1983; Fontugne et al., 1994), the Black Sea catchment basin via the Aegean (Kullenberg, 1952; Olausson, 1961; Ryan, 1972; Lane-Serff et al., 1997), or diffuse sources in the northern watershed of the Mediterranean (Cramp et al., 1988; Rohling and Hilgen, 1991; Fontugne et al., 1994). Hypothesis 2 examines the hypothesis of Rossignol-Strick et al. (1982), who stated that precipitation of the African monsoon (channeled by the Nile River) is the most important agent in e/p change in the Eastern Mediterranean Sea. If the freshwater source was a point source located in the east (such as the Nile River) rather than a source in the northern watershed, then the surface water in the Eastern Mediterranean Sea should have experienced a larger salinity change than surface water farther from the source (Fig. 1B).

(3) *Temperature records in time slices corresponding to S1 time from different locations have gradients indicative of a circulation reversal.*

The positive water balance of an estuarine circulation system, which has been proposed as one model of sapropel formation (e.g. Sarmiento et al., 1988; Thunell and Williams, 1989, among many others) requires a compensatory flow into the basin for reasons of salt and mass conservation. The

inflowing water is usually from sub-thermocline levels in the adjacent basin; it is thus colder and nutrient-rich. Such colder and nutrient-rich water may have enriched the nutrients and drawn down the temperatures in the Eastern Mediterranean by mixing with the upper waters during S1 time (Calvert et al., 1992). If a circulation reversal occurred, this must be reflected in a change of the temperature gradient just prior to sapropel deposition. The temperature change may be expected to

be larger near the source of the upwelling water (Fig. 1C).

2. Materials and methods

Samples of core KS 8230 were obtained from C. Vergnaud-Grazzini and we converted the published stratigraphy to calendar years using the program CALIB 3.0.3c (Stuiver and Reimer,

Table 1
Core locations and present-day conditions at the sea surface

Core	Position	Water depth (m)	Jan–Jun, 0–30 m ^a , SST (°C)	Annual averages at 5 m (Levitus and Boyer, 1994)		Alkenone SST (surface sediment, °C)
				Salinity	σ_t	
KS 8230	26°27.16'N/003°53.18'W	795	15.7	36.82	26.36	17.4
RL 11	36°44.75'N/017°43.05'E	3376	16.6	38.24	27.30	16.4
ODP Site 967	34°04.25'N/032°43.53'E	2551	18.7	38.73	27.11	18.5
Surface sediment samples ^b				U_{37}^k	SST M ^c (°C)	SST T ^d (°C)
M40/86-2	41°12.34'N/002°49.99'E		14.3	0.574	16.1	19.1
M40/85-4	41°21.01'N/003°01.34'E		14.4	0.555	15.5	18.7
M40/88-2	38°56.26'N/004°36.16'E		16.0	0.626	17.6	20.4
M40/87-6	38°59.28'N/004°01.38'E		15.8	0.568	15.9	19.0
M40/89-3	38°45.02'N/005°20.27'E		16.2	0.613	17.2	20.1
M40/80	40°57.30'N/011°00.24'E		15.2	0.555	15.5	18.7
M40/79-3	40°21.48'N/012°06.44'E		15.5	0.561	15.7	18.8
M40/78-2	37°02.17'N/013°11.38'E		15.7	0.568	15.9	19.0
M40/77B	37°00.21'N/016°09.38'E		16.3	0.594	16.7	19.6
M40/77	37°00.56'N/016°17.80'E		16.3	0.585	16.4	19.4
M40/73	39°31.70'N/018°58.27'E		16.0	0.579	16.2	19.2
M40/76-4	35°13.76'N/021°29.96'E		16.9	0.585	16.4	19.4
M40/76-2	35°13.83'N/021°28.29'E		16.9	0.629	17.7	20.5
M40/71	34°48.67'N/023°11.63'E		16.9	0.567	15.8	19.0
M40/70	33°42.82'N/024°42.11'E		17.5	0.590	16.5	19.5
M40/69	33°51.53'N/024°51.46'E		17.5	0.590	16.5	19.5
S20	34°02.74'N/027°46.80'E		17.1	0.601	16.9	19.8
S21	33°36.22'N/028°34.22'E		17.3	0.631	17.8	20.5
S23	32°40.83'N/030°35.73'E		17.9	0.646	18.2	20.9
BC4	35°35.81'N/030°34.54'E		17.7	0.611	17.2	20.0
BC5	35°00.26'N/031°09.53'E		17.8	0.594	16.7	19.6
BC7	33°39.39'N/032°40.04'E		18.8	0.652	18.4	21.0
S39	34°02.11'N/032°37.57'E		18.7	0.654	18.5	21.1

^a Brasseur et al. (1996).

^b S=Meteor 25/1, 1993, provided by M. Krom (box corer). M40=Meteor 40/4, 1998, own samples (multicorer). BC=Marion Dufresne MD 81, 1995, provided by I. Cacho (box corer).

^c SST M=calculated from Müller et al. (1998).

^d SST T=calculated from Ternois et al. (1997).

1993); the $\delta^{18}\text{O}$ curve has been established on the planktonic foraminifer *Globigerina bulloides* (Vergnaud-Grazzini and Pierre, 1991). A short (75 cm) trigger core of a longer piston core from the Ionian Sea (Meteor 25/1 RL 11) was made available by J. Keller (Freiburg). Samples of Hole 967 E were collected during Ocean Drilling Program Leg 160 (Emeis et al., 1996). Table 1 lists locations of cores analyzed here, as well as modern surface water properties (Levitus and Boyer, 1994); Fig. 2 shows their position.

For carbon and oxygen isotope analysis in core Meteor 25/1 RL 11, approximately 20 specimens of *Globigerinoides ruber* (white variety) were separated from the $>63\ \mu\text{m}$ fraction and ultrasonically cleaned in methanol. The samples were reacted at 90°C with 100% phosphoric acid on an automated carbonate device connected to a VG Prism mass spectrometer calibrated with NBS 19, NBS 28 and NBS 20. The results are reported in the conventional ‰ notation with reference to Vienna PDB. Some of the samples of the RL 11 core contained only the pink variety of *G. ruber*. Based on three samples from core RL 11, where both white and

pink species coexisted, and the data of Tang and Stott (1993), the oxygen isotope composition of *G. ruber* (pink) is $0.58 \pm 0.25\text{‰}$ more negative than that of *G. ruber* (white). For this reason the values of the samples which contained only the pink variety have been corrected by 0.58‰ .

From four intervals of core Meteor 25/4 RL 11 we extracted samples of mixed planktonic foraminifers ($>63\ \mu\text{m}$) for AMS ^{14}C datings (Table 2) and use interpolated corrected sample ages (using the program CALIB 3.0; Stuiver and Reimer, 1993) between these tie points in our age model (reservoir age assumed to be 400 years). The stratigraphy of Site 967 is based on visual correlation of the stable isotope and alkenone temperature curves from RL 11 and 967, and on assuming an isochronous age for the bases of S1 at both locations.

Lipid extracts from core samples were analyzed for alkenone unsaturation ratios according to the following procedures. Surface sediment and core samples were freeze-dried and homogenized. Subsamples of 1–3 g sediment were extracted by ultrasonic agitation with 35 ml CH_2Cl_2 ($2 \times 10\ \text{min}$). After each of the two extraction

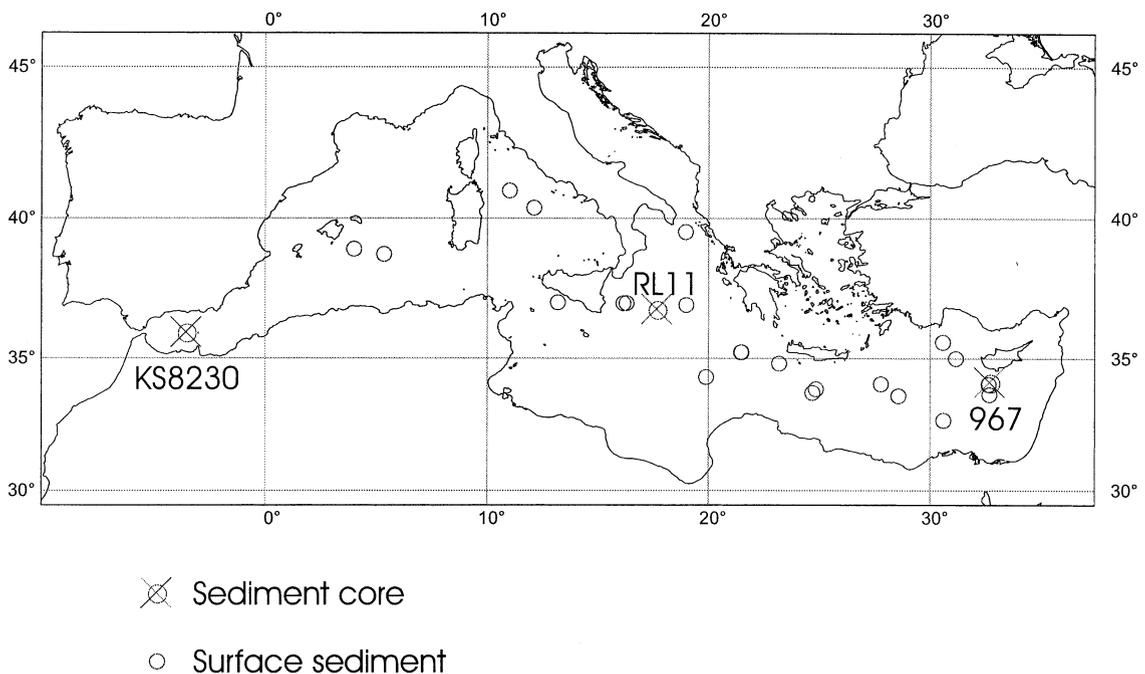


Fig. 2. Location map of cores and core top sediments analyzed.

Table 2
Results of AMS ^{14}C datations on mixed planktonic foraminifer tests ($>63\ \mu\text{m}$) from core RL 11

Sample depth (cm)	^{14}C age	$\pm 1\sigma$	$\delta^{13}\text{C}$	$\pm 1\sigma$	Calendar years ^a
12–13	2235	55	0.5	1.2	124–249 aC
31–32	6590	70	0.7	1.2	5224–5053 bC
42–43	9250	80	2	1.2	7885–8010 bC
77–78	14910	110	2	1.2	15282–15580 bC

^a Reservoir effect of 400 years subtracted.

steps, the samples were centrifuged and the solvent was collected by pipette and the combined extracts were dried in a rotary evaporator. Subsequently, the combined extracts were fractionated using high pressure liquid chromatography (HPLC) on a column packed with silica gel (MERCK LiChrosphere Si 100-5) with four elution steps (0.5 ml *n*-hexane, 14 ml *n*-hexane/dichloromethane [90:10; v/v], 9 ml dichloromethane and finally 9 ml acetone). Upon recovery checks, the ketones, including long-chain alkenones, were isolated in the second fraction; the first fraction contained aliphatic hydrocarbons, the third and fourth fractions yielded the more polar compounds. Gas chromatographic analyses were carried out on a HR 8000 Fisons gas chromatograph fitted with a cold on-column injection system, a 30 m (0.32 mm inner diameter) glass capillary column (DB5HT) and a flame ionization detector. Hydrogen was used as carrier gas, flow rate 2 ml/min. Oven temperature was run from 35 to 300°C at 15°C/min followed by an isothermal period of 15 min duration at 300°C, during which the alkenones elute, and from 300 to 330°C at 15°C/min followed by an isothermal period of 10 min at 330°C. Alkenones were identified based on their retention times and those of standards that have been examined by gas chromatography/mass spectrometry. Peak areas were used to calculate the $U_{37}^{k'}$ index: $U_{37}^{k'} = C_{37:2}/(C_{37:2} + C_{37:3})$ according to Brassell et al. (1986). Alkenone concentrations were determined by comparing FID responses to known external standard concentrations (5 α -cholestane).

2.1. SST reconstructions

To derive sea surface temperatures from alkenone unsaturation ratios, we used the global rela-

tionship between $U_{37}^{k'}$ in surface sediments and sea surface temperatures found by Müller et al. (1998):

$$\text{SST } (^{\circ}\text{C}) = (U_{37}^{k'} - 0.044)/0.033, \quad (1)$$

which is virtually identical to the original equation by Prahl and Wakeham (1987) based on cultures of *Emiliania huxleyi*.

The limited data on Eastern Mediterranean surface sediment samples (Table 1 and plotted in Fig. 3) suggest that temperatures derived from Eq. (1) agree well with the surface water (0–30 m) winter and spring SST averages of the core locations taken from the Mediterranean Oceanographic Database (Brasseur et al., 1996) and fall within the global trend.

The previously published relationship of Ternois et al. (1997):

$$\text{SST } (^{\circ}\text{C}) = (U_{37}^{k'} + 0.21)/0.041 \quad (2)$$

for suspended matter in the Northwestern Mediterranean Sea was found to give a less satisfactory fit for Mediterranean surface sediments (Fig. 3), but has been calibrated against actual in situ temperatures and does not suffer from smoothing and edge-artifacts in the climatological atlases. The alkenone flux data given by Ternois et al. (1997) suggest that the alkenone unsaturation ratios must be weighted by the seasonal production cycles of haptophytes in the modern oligotrophic Eastern Mediterranean, where seasonally different fluxes are indicated: because stratification breaks up in cold winter months, this is the time of nutrient recharge to the euphotic zone (Krom et al., 1992) and spring and winter are the dominant production seasons of coccolithophorids in the Mediterranean Sea (Knappertsbusch, 1993).

By using the first relationship, we run the risk of overestimating the variability in SST: the range

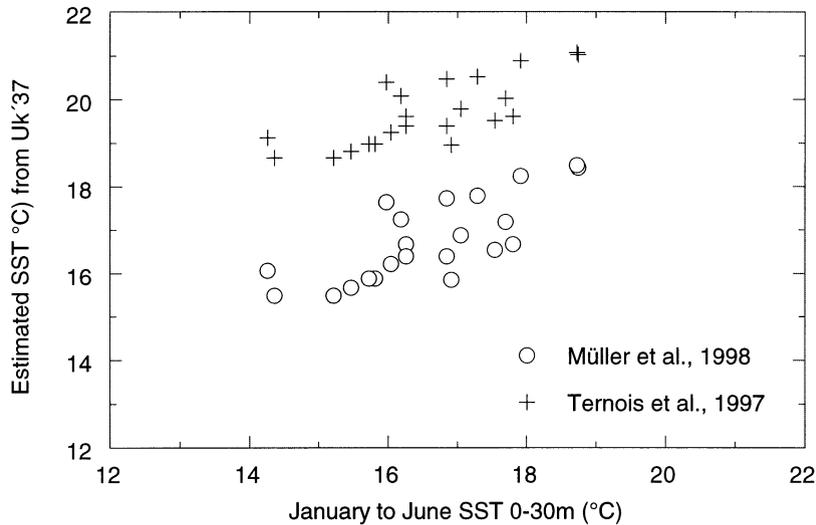


Fig. 3. Average winter and spring sea surface temperatures in 0–30 m water depth at the locations of surface sediment samples (locations in Fig. 2) from the Mediterranean Sea (calculated from seasonal averages given by Brousseau et al., 1996) plotted against the temperature estimates derived from alkenone unsaturation measurements. The circles are based on temperatures according to the equation given by Müller et al. (1998); crosses denote estimates based on the regression equation by Ternois et al. (1997).

of U_{37}^k values in our samples is from 0.42 to 0.824, and the corresponding range of SST estimates is 11.4°C (a sample deposited during last glacial maximum) and 23.6°C (a sample deposited during the climatic optimum around 6 kyr) using the Müller et al. (1998) equation. Calculated with the Ternois et al. (1997) equation, the range is between 15.4 and 25.2°C. The coldest estimate using the Ternois et al. (1997) relationship would imply that at no time in our records have SST been cooler than the present-day deep water (13.2°C) in the Eastern Mediterranean Sea, which is unlikely. A more important argument for using the Müller et al. (1998) equation is the temperature levels calculated for those parts of cold sapropels S6 and S8, in which the planktonic foraminifer *Globigerinoides ruber* (white) is absent (Emeis et al., in preparation). This species has a lower temperature limit of 14°C (Bijma et al., 1990). It was not encountered in coarse fractions (prepared for oxygen isotope measurements given in Emeis et al., 1998) of those samples of S6, for which the average SST estimate according to Müller et al. (1998) was $15.2 \pm 1.6^\circ\text{C}$ ($n=25$). The corresponding average SST estimate from the Ternois et al. (1997) equation is $18.4 \pm 1.3^\circ\text{C}$, which is well

within the range tolerated in culture experiments. For samples of S8 barren of *G. ruber*, the estimate based on Müller et al. (1998) was $16.4 \pm 1.5^\circ\text{C}$ ($n=11$); the Ternois estimate was $19.4 \pm 1.2^\circ\text{C}$. That this absence of *G. ruber* is not an effect of salinity change is indicated by the presence of abundant tests in samples from the warm sapropels S7 and S5 of the same location and a salinity tolerance range of 22 to 49 PSU (Bijma et al., 1990).

The dataset of Brousseau et al. (1996) gives a range of 3.6°C (from 16.6 to 20.2°C) for spring and 3.6°C (from 13.9 to 17.3°C) for winter sea surface temperatures in the water column 0–30 m at the core locations; these two seasons are the dominant production seasons of coccolithophorids in the Mediterranean Sea (Knappertsbusch, 1993). Using the (warmer) annual average values may overestimate the effect of temperature on variations in isotopic fractionation and on salinity. However, our main objective in this work is to investigate gradients between records that have been calculated using the same set of regression equations.

Organic carbon concentrations were determined on subsamples by measuring total carbon evolving

as CO₂ after combustion at 1200°C on a Heraeus elemental analyzer and by determining the amounts of inorganic carbon after treatment with 17% phosphoric acid in an Eltra carbon analyzer.

2.2. Salinity reconstructions

The record of salinity at each of the sites is a composite of regional, global, and local effects that need to be considered in order to obtain local variations. To calculate salinity from alkenone and δ¹⁸O data, we use the expression given by Rostek et al. (1993):

$$S = \Delta S_0 + S^* + (\Delta \delta^{18}\text{O}_F - a - b\Delta T)/c. \quad (3)$$

Here, S is past local salinity, S^* is the present-day local salinity (taken from Levitus and Boyer, 1994), and $\Delta \delta^{18}\text{O}_F$ is the difference in ¹⁸O/¹⁶O composition of planktonic foraminifer calcite at time t and the present. ΔS_0 means the global change of ocean salinity due to ice shield growth; because this change affects all locations in the same manner, we neglect its effect in our comparison of the three sites in the Mediterranean Sea. The correction parameter a represents time-dependent global variations of seawater δ¹⁸O attributable to ice volume changes (Vogelsang, 1990). The constant b quantifies the temperature effect on δ¹⁸O_F [−0.2‰/K in the case of *Globigerinoides ruber* in the Indian Ocean (Duplessy et al., 1981) and approximately for carbonate water isotopic equilibrium in general (Epstein et al., 1953)]. This coefficient relates δ¹⁸O_{calcite} to δ¹⁸O_{seawater}. To estimate this isotopic change related to temperature change, we use the independently estimated temperature from alkenone unsaturation ratios. ΔT stands for the temperature change from modern to past. We used alkenone temperature estimates based on surface sediment measurements as our modern end-member; these values are given in Table 1 for all three sites. At 10°C colder temperatures during the last glacial compared with modern, the effect is $b \cdot T = +2‰$ δ¹⁸O which must be subtracted from Δδ¹⁸O_{glacial–interglacial}. The maximum ice effect in our data set is $a = 0.93‰$ (at 16,000 years, taken from the Vogelsang, 1990 data set), which is also subtracted from Δδ¹⁸O to

yield a local isotopic difference of seawater compared with modern. Finally, c is the coefficient relating salinity and δ¹⁸O_{seawater}. We use two ratios. The first is $c = 0.25‰$ δ¹⁸O/p.s.u. (Pierre, 1999) and is based on a recent and comprehensive survey of surface waters that shows a linear correspondence between salinity and δ¹⁸O. The ratio may have been different in the past, when different moisture sources were providing freshwater to the Mediterranean (Rohling and Bigg, 1998; Rohling and de Rijk, 1999). To account for such possible changes during S1 time, we also use an alternative ratio of $c = 0.45‰$ δ¹⁸O/p.s.u. This ratio is based on the assumption that the freshwater end-member in this period may have been much lighter in isotopic composition (δ¹⁸O = −16‰ compared with −8.2‰ today; Pierre, 1999). A value of 0.41‰ δ¹⁸O/p.s.u. was previously used by Thunell and Williams (1989) in an attempt to reconstruct Mediterranean salinity changes. With these two values for the freshwater end-member signature, we bracket a considerable range of relationships between salinity and δ¹⁸O_{seawater} such as may have occurred in the past.

A possible source of error is our normalization-to-modern approach, which ties all past variations to the isotopic values and temperature reconstructions measured in the modern surface sediment. Any error would propagate downcore and distort the real signal of salinity. On the other hand, such an error in the modern tie point would result in a systematic offset in the paleoceanographic reconstructions. In this case, the relative change between surface water properties in the Ionian Sea and the Levantine Sea, which we consider to be crucial for our intent, would still reflect real phenomena.

A second problem in our method is the possible seasonal offset in the formation of the isotopic and the temperature signal. We ignore the possibility that *Globigerinoides ruber* has a different growth season compared with the coccolithophores which form the alkenone signal. We also ignore that both planktonic organisms may reside in different depth habitats. Our justification is that *G. ruber* reproduces in a lunar cycle and thus is present in the to 30 m water column over the entire year, even though maxima in its abundance appear

to lag coccolithophore blooms (Hemleben, personal communication, 1998) by some weeks. Its depth habitat is comparable with that of *Emiliania huxleyi*, the dominant alkenone producer (Knappertsbusch, 1993; Winter and Siesser, 1994).

3. Results

All analytical data may be obtained upon request to the senior author. Values of TOC are very uniform and vary between 0.5 and 1% in the Alboran Sea core over the time interval since 16,000 years. The record lacks any indication of an increase related to contemporaneous sapropel deposition in the Ionian Sea (RL 11) or the Levantine Basin (Site 967 north of Eratosthenes Seamount; Fig. 4A). Alkenone concentration levels in core RL 11 increase at around 9500 years and are high throughout the sapropel interval. It is dated to begin at 9450 and to end at 6630 calendar years in the Ionian Sea core. Steeply rising alkenone concentrations also mark the onset and full sapropel interval of S1 at Site 967. According to shipboard TOC determinations performed during ODP Leg 160, S1 contains 2.56% organic carbon at Site 967.

3.1. Stable oxygen isotope composition of planktonic foraminifers

Oxygen isotope composition of planktonic foraminifer tests is plotted versus age in Fig. 4B. The maximum glacial–interglacial amplitude of *Globigerina bulloides* $\delta^{18}\text{O}_{\text{calcite}}$ in the Alboran Sea core KS8230 is from 2.99‰ at 15,000 years to 0.46‰ at 8000 years, amounting to a $\Delta\delta^{18}\text{O}_{\text{glacial–interglacial}}$ of 2.5‰. In core RL 11, we find that the lowest sample (interpolated age 15,900 years) contains tests of *Globigerina ruber* with a $\delta^{18}\text{O}_{\text{calcite}}$ of 2.61‰, whereas $\delta^{18}\text{O}$ in the sample of 8700 years (within S1) is much lighter at -0.26‰ . This gives a $\Delta\delta^{18}\text{O}$ of approximately 2.9‰ in the Ionian Sea. The isotope value at Site 967 is 3.11‰ for $\delta^{18}\text{O}$ of *Globigerina ruber* at 13,700 years and -0.66‰ for a sample of 9000 years in S1; the $\Delta\delta^{18}\text{O}$ in the Levantine Basin is thus around 3.8‰.

$\delta^{18}\text{O}$ immediately before the onset of sapropel S1 deposition decrease gradually to 0.94‰ at 9700 years from 2.09‰ at 13,500 years in the Ionian Sea core. In the period between 10,400 and 8700 years, when high alkenone concentrations indicate that full sapropel conditions have been established, $\delta^{18}\text{O}$ has decreased by 1.2‰ to a value of -0.26‰ . At Site 967, pre-sapropel $\delta^{18}\text{O}$ is 1.21‰ at 11,000 years and drops to -0.66‰ at 8980 years, when sharply increasing alkenone abundance signals the establishment of sapropel conditions approximately 2000 years later.

3.2. SST estimates

Alkenone SST estimates show a high amplitude of temperature change in the Mediterranean Sea over the last glacial–interglacial cycle with highest variability in the record from the Ionian Sea (Fig. 4C). In the Alboran Sea core, the paleo-SST derived from the molecular thermometer have been very uniform at $18.3 \pm 0.4^\circ\text{C}$ over the last 9000 years. Before that time, SST decrease rather steadily to values around $13.7 \pm 1.2^\circ\text{C}$ between 14,000 and 16,000 years; glacial–interglacial temperature change is thus of order 5°C in the Alboran Sea.

The Ionian Sea record has a higher amplitude of glacial–interglacial temperature change and displays more variability. Minimum temperatures as low as 13°C occur before 14,000 years. SST rise to values around 21°C in the Alleröd and decline again to 14°C during the Younger Dryas. During S1, the values of SST vary between 16 and 19°C and reach a plateau of 22°C between 6000 and 4000 years. The amplitude between the glacial and the post-glacial interval is thus in excess of 8°C . Average temperatures at this site are $18.2 \pm 0.1^\circ\text{C}$ for the three youngest samples, $20.9 \pm 1.8^\circ\text{C}$ for post-S1 samples, $17.0 \pm 0.9^\circ\text{C}$ for S1, and $15.4 \pm 1.4^\circ\text{C}$ for samples from before 13,000 years. The variability is low during S1 deposition and much higher before and after the sapropel event.

At Site 967 in the Levantine Basin, the average temperature in the period from 16,000 to 13,000 years is $14.0 \pm 2.0^\circ\text{C}$, rising to an average temperature $18.8 \pm 0.8^\circ\text{C}$ within the S1 time period. Samples above S1 have an average of 20°C with a

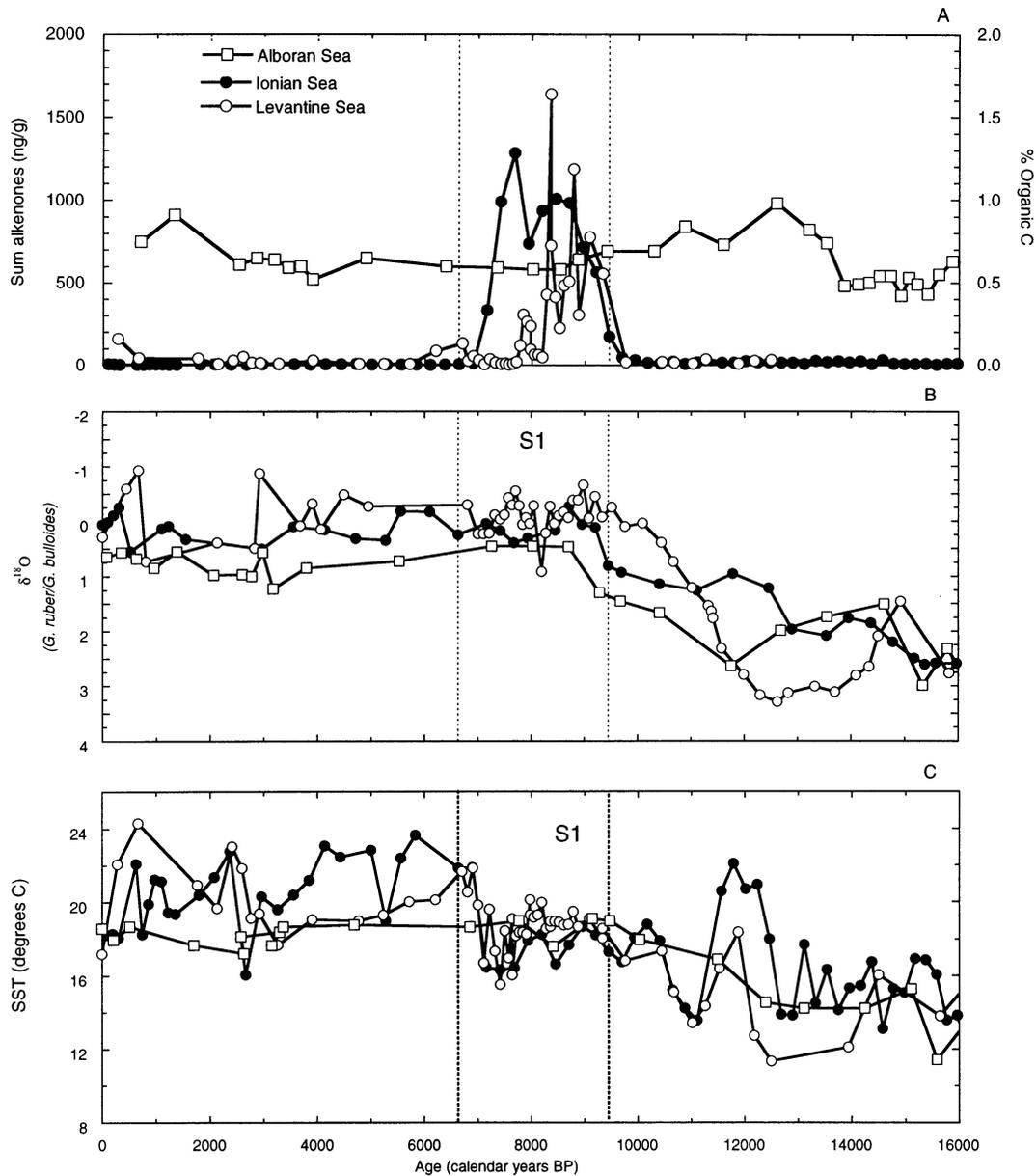


Fig. 4. (A) Organic carbon and alkenone concentrations in the three cores plotted against age mark the onset of S1 (vertical dashed lines) in cores RL 11 (Ionian Sea) and 967 (Levantine Basin). The Alboran Sea core KS8230 shows no sign of increased organic carbon concentrations. (B) Oxygen isotopic composition of planktonic foraminifer calcite in the cores; the data for the Alboran Sea are from Vergnaud-Grazzini and Pierre (1991). (C) Sea surface temperature estimates (from alkenone unsaturation ratios) for the cores.

large scatter ($\pm 1.5^{\circ}\text{C}$). The highest value is 23°C at 2400 years. The youngest sample at 5 cm depth has an alkenone temperature of 22°C , which con-

trasts with values of three surface sediment samples (0–1 cm) from the area that have an average of 17.9°C (Table 1).

3.3. Residual $\Delta\delta^{18}\text{O}$ curves and salinity estimates

The stepwise correction of $\delta^{18}\text{O}$ values for the effects of global ice volume and temperature and the resulting curves of $\delta^{18}\text{O}_{\text{seawater}}$ are depicted in Fig. 5A–C for each of the three cores. The curves of $\delta^{18}\text{O}_{\text{seawater}}$ shown in Fig. 5D (the Alboran Sea is offset by 2‰ for clarity) all show considerable variations that argue for either salinity changes, or for changes in the freshwater end-member determining seawater $\delta^{18}\text{O}$ in the Mediterranean Sea. Even if the salinity-to- $\delta^{18}\text{O}_{\text{seawater}}$ ratios may have changed in the past (e.g. Bigg, 1995; Rohling and Bigg, 1998), these changes will have affected the entire Mediterranean Sea surface water mass and a linear relationship between the two properties will most probably have been maintained. Furthermore, the changes must have affected all three sites simultaneously (within our temporal resolution), and thus the data from the three basins are comparable; the same applies to the global salinity decrease [term ΔS_0 in Eq. (3)]. Assuming — as outlined in the methods sections — two different ratios for the relationship between salinity and $\delta^{18}\text{O}$ of seawater, we calculated curves of salinity change from modern salinity for each of the sites. Minimum (assuming a ratio of 0.45‰/p.s.u.) and maximum (assuming a ratio of 0.25‰/p.s.u.) salinity variations through time are shown in Fig. 6A–C.

The salinity envelopes from the eastern basins show high values (larger than modern) during the end of the glacial, a decrease before or during S1 time, and a return to higher than modern salinity in the Ionian Sea after S1 time; the Alboran Sea and Levantine Basin salinities have remained close to modern values since S1.

In the Levantine Basin at Site 967, the salinity of surface water changed dramatically before sapropel S1 began to form. The salinity increased strongly to approximately 12,000 years and dropped significantly (over 4 to 7 p.s.u.) to values below the modern salinity. The modern salinity threshold is crossed at 11,000 years. During and after S1 deposition, the salt content in the Levantine Basin surface water is lower than or slightly above modern salinity.

In the Ionian Sea, salinity is higher than modern

for the entire period before S1, with the exception of a short minimum after 12,000 years B.P. that saw a decrease in salinity (Cheddadi et al., 1991). The S1 interval is marked by decreased salinity, which was close to modern values, and a subsequent rise at the top of the S1 period. At both Eastern Mediterranean locations, the salinity low during S1 deposition was interrupted by a return to higher values (>1 p.s.u.) centered at approximately 8000 years. However, this increase in salinity did not interrupt the deposition of S1 in either location, because the sediments remain organic-rich.

In the Alboran Sea reconstruction, which is not very resolved in time, salinity appears to have been significantly higher than modern only during the Younger Dryas until 9000 years B.P., which would argue for an enhanced influence of continental climate upon this location that is usually dominated by water exchange with the Atlantic. In the period of S1 deposition, salt concentrations dropped by 1 to 2 p.s.u. some 1000 years after the onset of sapropel deposition in the Eastern Mediterranean and have stayed around modern values since then. The overall amplitude of salinity change in the Alboran Sea is around 3 p.s.u., and shows little variability in the time after S1.

The three curves (Fig. 6D) highlight differences in the details of salinity variations of the Mediterranean sub-basins, in particular between the two locations in the Eastern Mediterranean Sea. At all three locations, salinity peaked at 12,000 calendar years B.P. Sapropel S1 deposition in the Ionian Sea occurred after salinity had dropped by 2 p.s.u. from that maximum around 2500 years later. In contrast to our expectations, the Ionian Sea shows no indication of a strong salinity decrease prior to S1, and during the deposition of S1 salinity never decreased much below the 38.2 found in the Ionian Sea surface waters today. Instead, salt content in the surface waters remained approximately one unit higher than at present within the basal sapropel, and the increase in TOC concentrations preceded the salinity drop over 2 p.s.u. between 9500 and 8600 years. At the top of the sapropel (present-day surface of S1 in core RL 11 dated at 6600 years), the salinity increased abruptly by three units. In contrast to

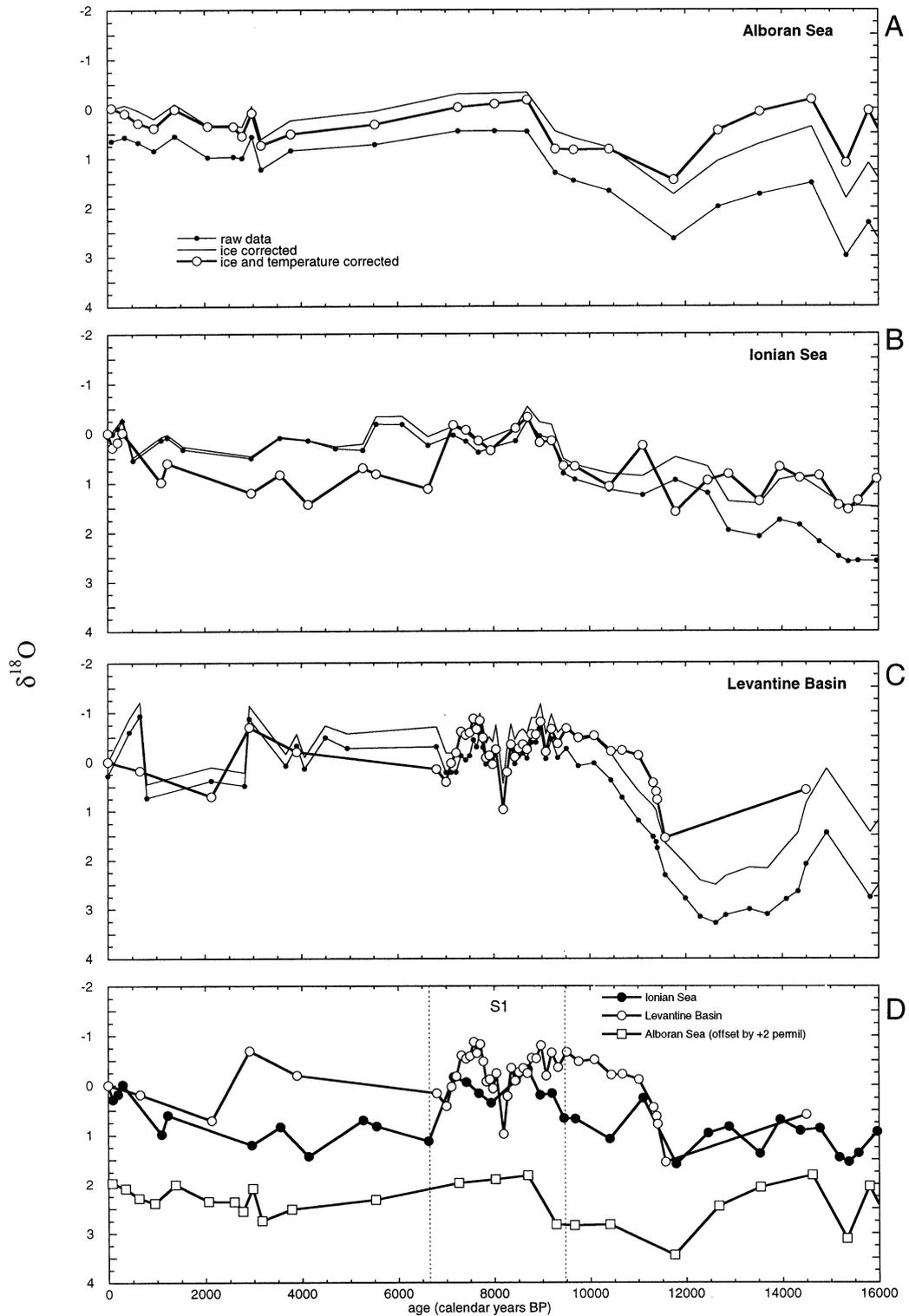
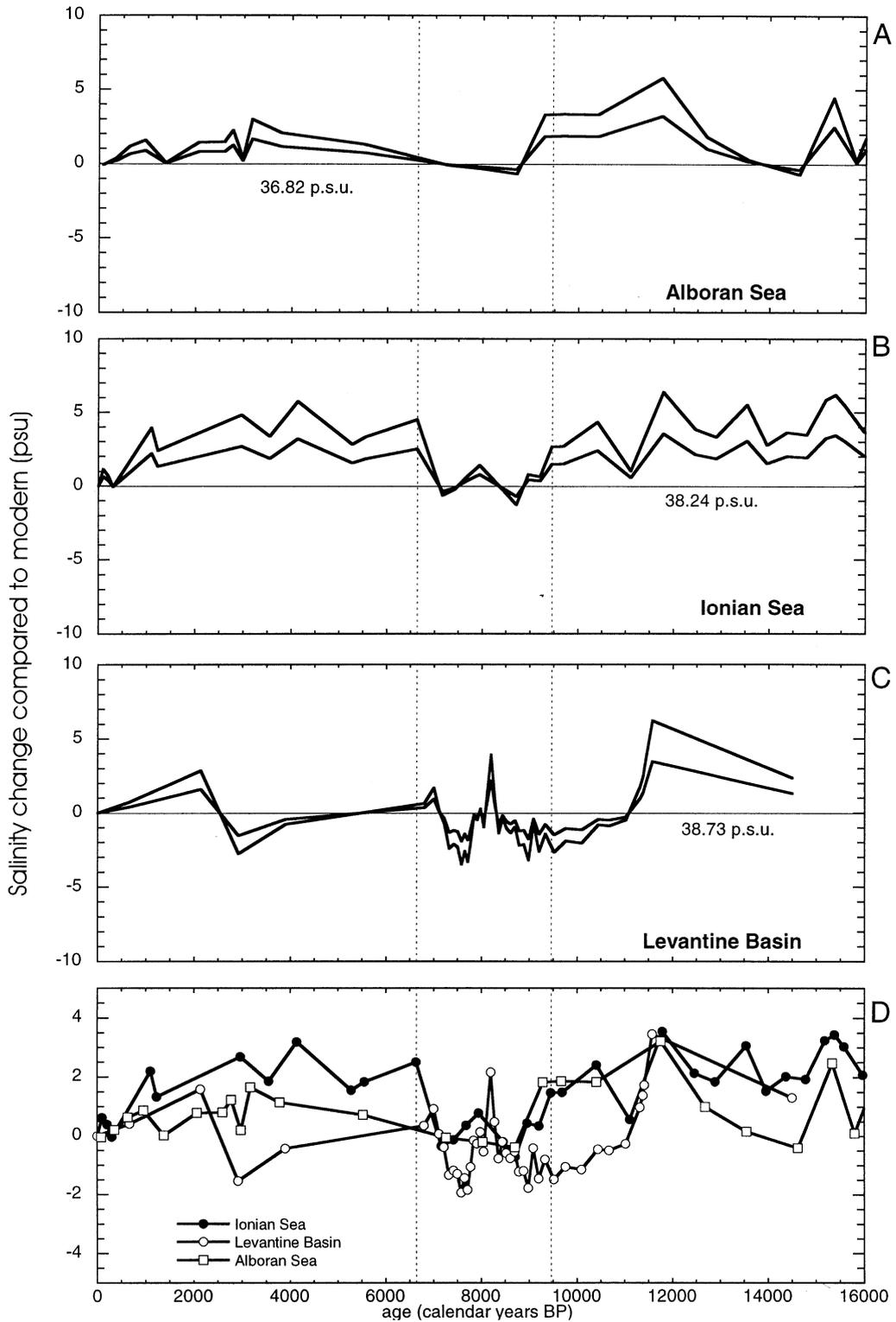


Fig. 5. (A–C) Residual variations of $\delta^{18}\text{O}$ (plotted stepwise as original data, after temperature, and after ice volume correction) for the three records; (D) is a summary plot for all three records.



the Levantine Basin record, modern values have been established only during the last few hundred years following a substantial decrease from higher than modern salinities during the last 500 years before Present.

According to our reconstruction, the salinity gradient between the two Eastern Mediterranean locations has been higher in the past compared with the modern situation. It also may have been reversed (lower salinity in the Levantine Basin) over extended periods of time: two periods of prominent and long-lasting reversal in the salinity gradient are seen between 12,000 and 10,000 years, and in the time after S1 deposition until 3000 years.

To summarize the results: temperatures and salinities in the three cores show marked temperature increases and salinity decreases in the time from the late glacial to the onset of sapropel S1 deposition in the Eastern Mediterranean Sea. The two records from the eastern basin are similar in temperature evolution, but differ significantly in the timing of salinity changes. The Alboran Sea record, which is a western basin control, traces the temperature signal found in the east, but the range of temperature variation is reduced. The same is seen in salinity reconstructions for the time before and during S1 deposition.

4. Discussion

We set out to reconstruct the temperature and salinity histories at three locations in the Mediterranean Sea. With these data we wished to examine how gradients in the physical properties may have changed since the last glacial and, in particular, how these changes may be related to sapropel S1 formation in the Eastern Mediterranean Sea. The data gathered suggest that the narrow ranges in temperature and density that govern the modern configuration of surface, intermediate and deep waters in the Mediterranean Sea

have been much wider and that hydrographic conditions may have been significantly altered.

Because the temperature is the most direct variable of our reconstructions (with all due reservations discussed above), we begin our discussion with an evaluation of SST history. Our data suggest that glacial–interglacial temperature change, which is close to 5°C in the Alboran Sea and up to 8°C in the Ionian Sea, was higher than previously thought. In these two locations, minimum temperatures in the LGM were as low as 14°C. Site 967, which is located in an area bordering both the Rhodes Gyre where Levantine intermediate water forms and the low salinity area along the Israeli coast expressing Nile discharge, has experienced temperatures as cold as 12°C, compared with a modern alkenone surface sediment value of 17°C and a winter and spring average SST (from climatology) of 18.7°C.

How do these data compare with previous reconstructions? Thiede (1978) examined SST variations based on faunal assemblages in surface sediments and sediments from the last glacial maximum (LGM). His estimate is that winter SST during the LGM ranged from 7 to 21°C, whereas summer SST were 13 to 25°C. In general, faunal reconstructions showed that SST in the Western and Eastern Mediterranean were on average 4 and 6°C colder than the present SST, respectively (Thunell, 1979; Thiede, 1978). CLIMAP (1981) found an increase in SST of 4–6°C from LGM. The paleoceanographic reconstructions (based on faunal assemblages and stable oxygen isotopes) are consistent with vegetation patterns in the Eastern Mediterranean catchment area as recorded in pollen spectra (Cheddadi et al., 1991). The pollen suggest that the climate in the Ionian Sea area during the last glacial was arid and very cold during winter, and colder than present during summer (consistent with 4–8°C lower than modern). The alkenone data also show a larger temperature change in the Ionian Sea compared

Fig. 6. (A–) Salinity envelopes reconstructed for the three records plotted versus time. The lines denote salinities calculated from residual $\delta^{18}\text{O}$ at rates of 0.25 (upper line) and 0.45‰ $\delta^{18}\text{O}/\text{p.s.u.}$, respectively; (D) is a summary plot for all three records using the 0.45‰ $\delta^{18}\text{O}/\text{p.s.u.}$ ratio.

with the other basins, as previously described (Thiede, 1978; Thunell, 1979; Cheddadi et al., 1991). All these reconstructions, including ours, differ significantly from model calculations: Bigg (1994, 1995) estimates that SST in the western basin were 3–4°C lower than modern, whereas the model suggests no significant change in SST for most of the Eastern Mediterranean Sea. It seems implausible that four independent proxy indicators of temperature (i.e. faunal, floral, isotopic, and alkenone unsaturation) should all give comparable results, but still be in error.

Significant temperature variations at the sea surface are important in four respects. Firstly, they determine the evaporation and thus control the salinity and $\delta^{18}\text{O}$ signature of the surface water (Bigg, 1995). Secondly, they must be taken into account when correcting the $\delta^{18}\text{O}$ of foraminiferal calcite for temperature-related fractionation of seawater $\delta^{18}\text{O}$ (Epstein et al., 1953). Thirdly, they contribute to changes in surface water density. Finally, SST reconstructed from isochronous sediments are indicators of spatial gradients between the core locations. We return to our hypotheses formulated in the Introduction and first examine Hypothesis 3: *Do temperature gradients indicate that upwelling or a related circulation reversal was a factor in sapropel deposition?*

The temperature gradients in the coeval sapropel interval of S1 at the two locations in the Eastern Mediterranean Sea (Ionian and Levantine Basins) do not substantiate the hypothesis of circulation reversal and upwelling as a cause for sapropel deposition. A reversed gradient would be expected if shoaling of the thermocline and ascent of sub-thermocline waters in the eastern part of the basin were the reason for increased organic carbon burial in sapropel S1. However, we see no reversal in the SST gradient; the data thus argue against reversal. Today, the E–W temperature gradient between the two locations is 2–3°C during the winter months, when the phytoplankton bloom occurs in the Eastern Mediterranean Sea. The modern gradient is indeed reflected in core-top alkenone temperatures across the Eastern Mediterranean Sea (Fig. 3). It has persisted over the entire period of S1 deposition (Fig. 4C) at a

lower level of 1°C ($17.7 \pm 1.5^\circ\text{C}$ in the Ionian Sea core, $18.7 \pm 1.2^\circ\text{C}$ in the Levantine Basin core).

While the gradient is not changed significantly during S1, a reversed temperature gradient is indicated before 11,000 years: during and before the Bölling/Alleröd warm phases, the alkenone temperature reconstructions show significantly higher values in the Ionian Sea than in the Levantine Basin, and higher values are also evident in the period after S1 until 3000 calendar years ago. The dramatic cooling associated with the Younger Dryas chronozone (at 11,000 calendar years in our SST record) is more pronounced in the Ionian Sea core (from >20 to 14°C) than in the Levantine Basin location (from 18 to 14°C). During this short climatic deterioration, the temperatures decrease again to glacial values and coincide with a return to cold and dry conditions in the Northern Mediterranean catchment area (Cheddadi et al., 1991).

Whereas both Eastern Mediterranean basins may have been colder during the Younger Dryas, the dry conditions inferred from the Ionian Basin salinity record (Fig. 6B) apparently have not influenced salinity in the Levantine Basin. In our reconstruction, the maximum amplitude of salinity change over the last 16,000 years is between 10.5 and 6 p.s.u. (depending on the salinity to $\delta^{18}\text{O}$ ratio used to calculate the paleosalinity) in the Levantine Basin, between 8.7 and 4.7 p.s.u. in the Ionian Sea, and between 5.6 and 3.1 p.s.u. in the Alboran Sea. Thunell and Williams (1989) used isotope measurements on planktonic foraminifers of three time slices (18,000 yr, 8000, and modern) to reconstruct surface water salinities in the Mediterranean basins with an assumed ratio of 0.41‰ $\delta^{18}\text{O}$ /p.s.u. According to their results, salinity changes in the western basin are 2.9 p.s.u. between LGM and 8000 years. Salinities in the Eastern Mediterranean exceeded modern values by almost 3 p.s.u. during the glacial and have a glacial–interglacial amplitude of 5.6 between LGM and the climatic optimum at 8000 yr B.P. A similar salinity increase was found by Béthoux (1984) based on oceanographic balances. In contrast, Bigg (1994) postulates that salinities were almost unchanged. Our data agree best with those of Thunell and Williams (1989), because our

approach is similar, but employs an independent estimate of SST from the alkenone unsaturation ratios.

Our Hypothesis 2 stated that *the salinity (and density) change is larger in the Eastern Mediterranean than in the other basins and that an E–W gradient in surface water salinity is indicative of a point source of freshwater.*

Our data so far may be interpreted as an indication for a dominantly eastern source of freshwater prior to and during sapropel S1 time, as proposed by Rossignol-Strick (1985). The range of salinity variations is greater in the Levantine Basin record than in the Ionian Sea and the Alboran Sea prior to and within the sapropel S1 period. Continued decrease in salinity in the Levantine Basin is indicated after 12,000 calendar years. In contrast, the Ionian Sea record sees a return to higher salinity accompanied by colder sea surface temperatures in the time around 11,000 years.

The longer duration of low-salinity conditions and the higher range of values argue for an eastern source of low-salinity water rather than a source in the northern watershed. This finding is consistent with previous results by Fontugne et al. (1994). On the other hand, the amplitude of salinity change within the intra-sapropel salinity spike at around 8000 years is more pronounced in the Levantine record. The eastern basin may thus have experienced a temporary decline in freshwater input, which is only weakly seen in the central basin and not at all in the western basin. Pollen records from the Eastern Mediterranean region (Rossignol-Strick, 1993) give evidence of dryer and colder conditions lasting a few hundred years prior to 7400 years that punctuate the climatic optimum indicated by the sapropel.

What conclusions do the records of temperature and salinity change permit with regard to water mass characteristics of the surface layer? The anoxia model of sapropel S1 formation requires a decrease in oxygen recharge into deep basins and thus decreases of deep and intermediate water formation rates. Both deep and intermediate water masses are crucial for the supply of oxygen to the deeper Mediterranean Sea. Lower rates or absence of deep water formation and of intermediate water

in particular would diminish the supply of oxygenated deep water, resulting in ‘stagnation’ and anoxia in the deep Eastern Mediterranean basins. In the modern Mediterranean Sea, minute (in comparison to our reconstructions of past T and S changes) seasonal differences in σ_t result in the formation of the Eastern Mediterranean deep water (EMDW; typical T and S : 13.6°C and 38.7 p.s.u.; σ_t of 29.13), as well as the Levantine intermediate water (LIW; typical T and S : 15.5°C and 38.97–39.04 p.s.u.; σ_t of 29.05; Malanotte-Rizzoli et al., 1996). Deep water formation today is keyed to winter convection in the Adriatic and the Aegean Seas when fresh but cool water sinks and spreads out throughout the Ionian Sea and spills over the Mediterranean Ridge into the Levantine Basin (POEM, 1992). The deep water formation in the Southern Adriatic process entrains up to 80% of Levantine intermediate water (Wu and Haines, 1996). This intermediate water mass forms in the Rhodes Gyre, spreads out in a depth interval above 500 m and is cooled again in the Adriatic Sea (Malanotte-Rizzoli et al., 1996). It thus supplies a source of preconditioned dense water for cooling and sinking farther to the west. In some instances, the Rhodes Gyre region develops a vertically mixed convective chimney during winter cooling and produces Levantine deep water (typical T and S : 14°C and 38.8 p.s.u.; σ_t of 29.15), which admixes with EMDW.

The density contrasts between surface waters and intermediate and deep waters today are small and seasonal cooling is sufficient to create convective overturn. From a present-day perspective, increased density contrasts between the different water layers are required to offset the seasonal temperature lowering and convection resulting in LIW formation. A decrease in LIW formation and supply of preconditioned water to the Southern Adriatic would then result in a lack of deep convection and decrease in EMDW formation rates.

Fig. 7 depicts plots of values for σ_t which have been calculated from paired T and S estimates of our data sets; we again chose the lower boundary of salinity deviations from modern for the computation. Because spatial density differences are important in the formation of deep and intermedi-

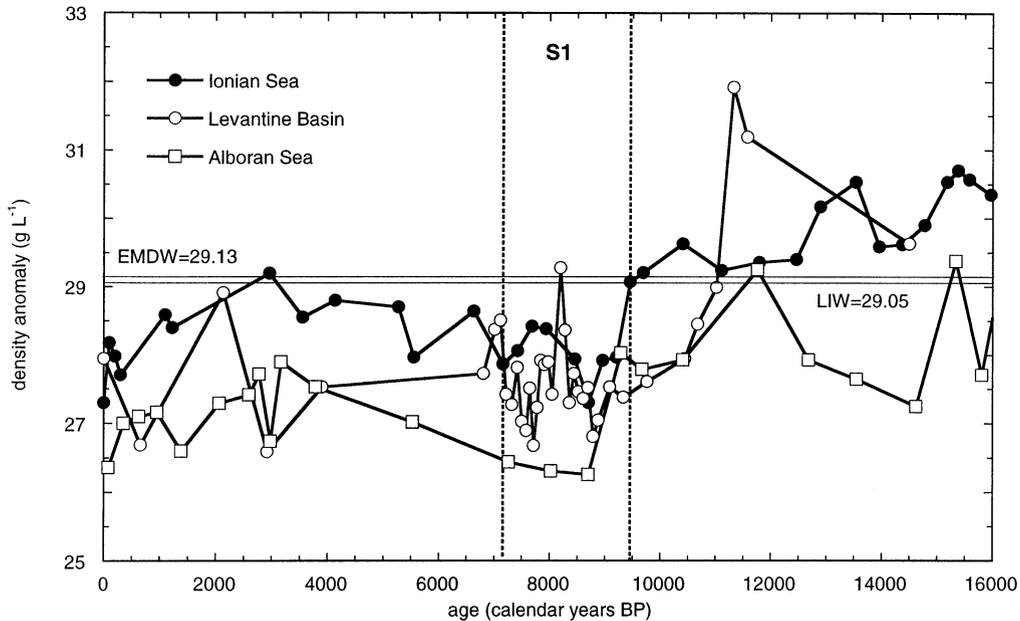


Fig. 7. Density anomalies (σ_t) for surface waters in the Ionian Sea, Levantine Basin and the Alboran Sea calculated from alkenone SST and salinity estimates plotted versus time. Present-day values of typical σ_t values for intermediate (LIW), deep (EMDW; WMDW) are given as straight lines for reference.

ate waters, the absolute values, which may be in considerable error, are of lesser importance in our reconstructions, which also do not consider the glacial–interglacial salinity change [ΔS_0 in Eq. (3)]. Values of σ_t typical for modern LIW and EMDW are plotted as straight lines for reference. The density of surface water in the Eastern Mediterranean was higher than modern during most of the time before 12,000 years in the two eastern basins; the Levantine Basin in particular has had very dense surface water. This configuration suggests that at the end of the last glacial, the circulation system was similar to the present one. With the onset of warmer and wetter climate during the Bölling and Alleröd chronozones, the surface water density in the Eastern Mediterranean Sea converged: the surface water in the Levantine Basin (and in the Alboran Sea) became progressively lighter, whereas it remained at close to present-day density in the Ionian Basin. The pronounced cooling of surface waters in the Ionian Sea region during the Younger Dryas maintained a high density there, so that from 11,000 years B.P., surface waters in both the Levantine Basin

and in the Alboran Sea were lighter than the coeval water in the Ionian Sea.

During the late glacial, density of surface water in the Levantine Basin appears to have been much denser than both modern intermediate and deep water. Even though the nature of the deep water of that time is unknown, the sedimentary facies and low organic carbon accumulation implies that oxygenation continued and that the formation of intermediate water may have been either unchanged or even invigorated. We thus speculate that EMDW (or better, its equivalent at that time) and LIW at the end of the glacial may have been denser than modern at 32 σ_t as a minimum, because by definition EMDW should be the densest water around and LIW is denser than the surface water mass.

Within the period from 11,000 years B.P. to the beginning of sapropel deposition, the density of the Levantine surface water had decreased significantly to approximately 27 σ_t . In concert with intermediate and deep waters of higher densities that formed at approximately 11,500 years to 11,000 years, this implies that density stratification

in the Eastern Mediterranean Sea was very stable. The density contrast may have lasted for a considerable period of time until eddy diffusion erased the density gradient: Rohling (1994) estimates that the density contrasts may have persisted over 1500 years. This estimate of hydrographic stability is considerably longer than the time required to deplete the dissolved oxygen content of the deep waters at increasing productivity rates in the surface (estimated at <700 years; Rohling, 1994; Ströhle and Krom, 1997).

Why did no sapropel form after oxygen depletion in the Levantine Basin (expected at 11,500–700 years = 10,800 years)? The reduced density at the location of Site 967 is not conclusive evidence for a complete absence of intermediate water formation. Our core location is east of the formation area in the Rhodes Gyre and under the influence of the Nile plume (see Fig. 2). Here, we may safely assume that sinking was much weaker. At the same time, the data are consistent with a possible shift of intermediate and deep water formation to the Ionian Sea, which in that critical time interval had the densest surface waters and was bracketed by areas of low density surface waters to the east and to the west. It appears that only after surface water density in the Ionian Sea decreased after 10,000 calendar years B.P., sapropel formation began almost instantaneously. The decrease in the Alboran Sea appears to slightly lag the decrease in the Ionian Sea. This pattern may be an indication for a hydrographic control by communication with the Atlantic Ocean and not by the adjacent marginal sea.

We now examine evidence for our Hypothesis 1, which stated that an *initial requirement for sapropel S1 formation is a density decrease in the sea surface; sapropel S1 is triggered by an input of freshwater which lowered the surface water density in the areas of intermediate and deep water formation to an extent which precluded sinking of oxygenated surface waters.*

Conditions in the sea surface at the two sites in the Eastern Mediterranean Sea, where S1 is present in the sediment record, do indeed conform with the hypothesis. However, the density decreased in the Levantine Basin alone, predating sapropel formation by 1500 years, and the local decrease

appears to be insufficient to upset the advection of oxygenated intermediate and deep water from the Ionian Sea. The increasing temperatures paired with decreasing salinities decreased the density of surface water to a considerable extent (by more than two units of σ_t) and the low density layer may have been as thick as 30 m (Tang and Stott, 1993). Intermediate and deep waters emplaced in the period of high sea surface densities before 11,500 years B.P. in the Levantine Basin, and during the Younger Dryas in the Ionian Basin, may have still occupied the deep and intermediate depth ranges so that the renewal rate of oxygenated water by convection and advection was low during S1 time.

The time lag of density decrease in the Ionian Sea remains unexplained, where the density at the onset of S1 deposition is not significantly different from before, compared with the much earlier decrease in the Levantine Basin. Available AMS ^{14}C ages for the onset of S1 deposition, when plotted against longitude in the Eastern Mediterranean Sea (Fig. 8), give no evidence of an east–west progression of anoxic conditions:

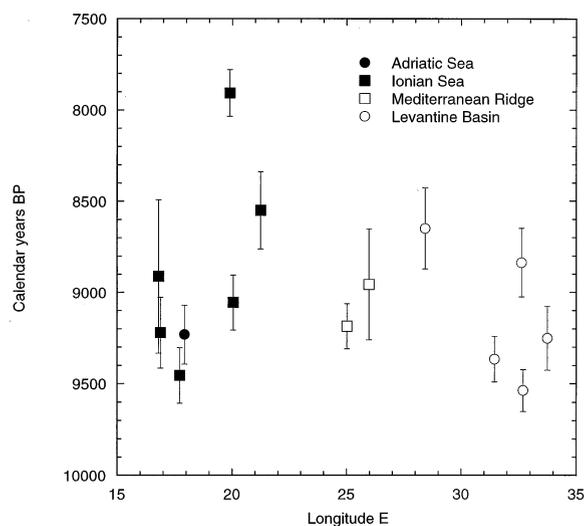


Fig. 8. Compilation of AMS ^{14}C ages for the bases of sapropel S1 plotted versus longitude in the Eastern Mediterranean Sea. The data have been converted to calendar years with the CALIB 3.0 program of Stuiver and Reimer (1993), assuming a reservoir age of 400 years where not already corrected in the source data. Data are from Fontugne et al. (1994), Troelstra et al. (1991), Thomson et al. (1995) and this work.

they appeared synchronously at all locations, but have a possible relation to water depth (Ströhle and Krom, 1997). We speculate that either a large pool of nutrients had to become available in the sea surface, or that patterns of organic matter flux increased to respond to the formation of hydrographic fronts. Both changes may have pushed the changed hydrographic environment into a mode of sapropel deposition. Amongst the possible sources of nutrients are (isotopically light) metabolites accumulated in a dysoxic to anoxic deep or intermediate water body that became available in the euphotic zone at the beginning of sapropel deposition (deep chlorophyll maximum; Rohling and Gieskes, 1989).

5. Conclusions and outlook

To improve our knowledge on the mechanisms leading to sapropel deposition in the Mediterranean Sea, it is necessary to establish proxy indicators that allow us to examine the physical situation in the sea surface layer. This layer is the source of intermediate and deep water masses today and the place where biological processes of organic carbon production interface with physical and chemical processes. Time series of surface water T and S proxy indicators since the last glacial were examined to test hypotheses on water column stability, temperature structure at the sea surface, and freshwater sources.

We interpret our results as indications that a current reversal is unlikely, because SST gradients are not reversed during S1 formation. Water column stability in the Eastern Mediterranean was enhanced prior to sapropel deposition and both intermediate and deep water formation rates were reduced. The homogeneity of salinity in the surface of the Eastern Mediterranean Sea during S1 time is consistent with multiple sources of freshwater. The increased runoff was not solely caught by the Nile watershed, and there is no indication that it entered the Mediterranean from the Adriatic Sea. Possibly it was channeled from a combination of sources.

To improve future work requires more robust knowledge about several variables. Firstly, the

$U_{37}^{K'}$ into sea surface temperature conversion needs to be improved for the Mediterranean Sea to improve hindcasts of the temperature evolution. A major potential source of error remains the assumption of a conservative $S:\delta^{18}O$ relationship at changed climatic conditions. Finally, the chemical and biological regimes associated with changed hydrographic conditions need to be reconstructed with appropriate proxies to find the triggering mechanism which set the Eastern Mediterranean into sapropel mode long after hydrographic change had occurred.

Acknowledgements

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