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Modeling the paleocirculation of the Mediterranean: The last glacial maximum and the Holocene with emphasis on the formation of sapropel S_1

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Abstract. An ocean general circulation model is used to simulate the thermohaline circulation in the Mediterranean sea during the last glacial maximum and the Holocene, when the sapropel S_1 was deposited. The model is forced by prescribed surface temperatures and salinities, where present-day values lead to very realistic surface buoyancy fluxes. Different paleoreconstructions for the surface salinity and temperature distributions during these periods are tested. In both periods, under all reconstructions, antiestuarine flow is maintained at Gibraltar and Sicily. The Holocene circulation has fresh intermediate water produced in the Adriatic and an upward salt flux from the old waters below help maintain its outflow at Sicily. The depth of ventilation around the basin is broadly consistent with the shallowest sapropel layers observed. Shoaling of the eastern pycnocline occurs in all experiments in both periods, possibly indicating enhanced productivity, although the reasons for this are different in each case.

1. Introduction

At the southern boundary of Europe the Mediterranean region is today generally warm, with a semi arid climate. The present-day Mediterranean (Figure 1) circulation is lagoonal or antiestuarine (both as a whole and for each of its main subbasins), with surface inflow of relatively fresh, light waters. Strong evaporation allows the Mediterranean to act as a concentration basin, converting the surface waters into a salty intermediate water mass (Levantine Intermediate Water (LIW)). This water exits the basin as a deeper outflow of dense water, which may play a significant role in the thermohaline circulation of the North Atlantic [e.g., *Reid*, 1979]. The LIW also reaches several sites in the northern parts of the basin (Adriatic, Gulf of Lions, and the Aegean) where it can precondition the water column for the formation of dense deep water masses (Eastern Mediterranean Deep Water (EMDW) and Western Mediterranean Deep Water (WMDW)). However, paleo-oceanographic records suggest that the Mediterranean circulation has not always been as it is at present.

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1.1. The Last Glacial Maximum

The most recent Pleistocene glaciation reached its maximum at ~20,000 years B.P. During this last glacial maximum (LGM) extensive ice sheets, up to several kilometers thick, covered the northern parts of North America and Eurasia [*Crowley*, 1988]. Mediterranean temperatures were seriously depressed, by 5°-10°C in winter and 1°-3°C in summer [*Prentice et al.*, 1992], with regions of permafrost extending into southern France [*Goudie*, 1992]. Around the northern shores of the Mediterranean and in the Levant the presence of steppe-like and salt-tolerant plants suggest a low level of precipitation prevailed [*Guoit*, 1987; *Goudie*, 1992; *Cheddadi and Rossignol-Strick*, 1995a]. Along the southern boundary of the Mediterranean there is more uncertainty regarding the water cycle. Fossil dunes [*Goudie*, 1992], groundwater [*Sonntag et al.*, 1980], and pollen samples [*Rossignol-Strick and Duzer*, 1980] suggest an expansion of aridity in the southern Saharan region; however, radiocarbon dating of high lake levels may suggest wetter conditions along the southern Mediterranean shore [*Street and Grove*, 1979; *Street-Perrott and Roberts*, 1983], especially in the west [*Goudie*, 1992]. There may also have been a shift in the seasonality of the precipitation to a more extreme dry summer and cold wet winter regime [*Prentice et al.*, 1992]. Changes in the monsoon over East Africa led to a significant decrease in precipitation there [*Goudie*,

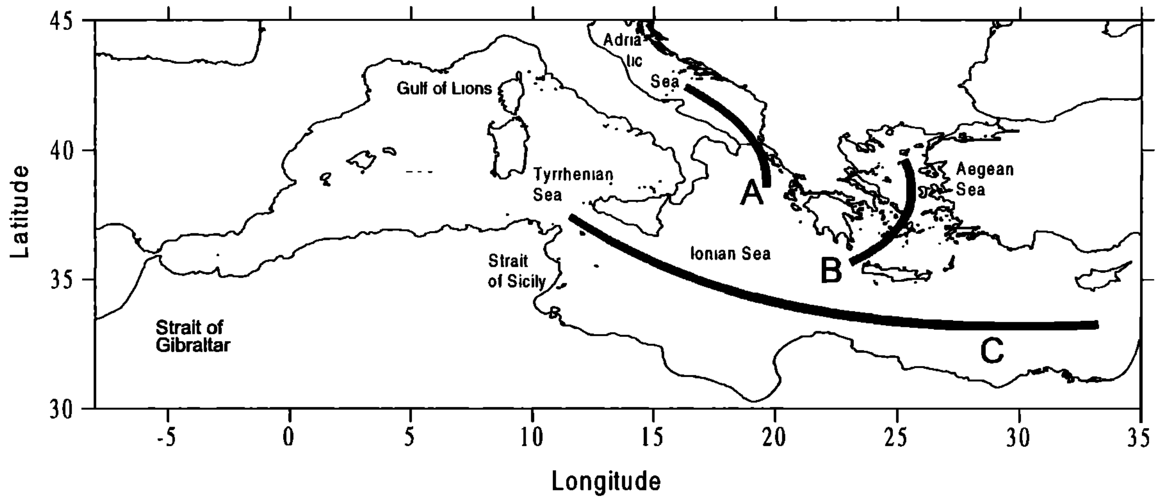

Figure 1. A map detailing some of the main Mediterranean locations mentioned in this paper. The sections shown are those for the schematics in Figure 14.

1992], turning the Nile River into a low discharge, seasonal river [*Adamson et al.*, 1980].

With large amounts of fresh water locked away in the continental ice sheets, global sea level was ~120 m below that of today [*Fairbanks*, 1989]. With decreased air temperatures, Mediterranean sea surface temperatures were also lower by 6°-10°C in winter, with generally smaller decreases in summer [*Bigg*, 1994]. In general, the temperature decreases are thought to have been larger in the western basin than in the east and Levantine [*Thiede*, 1978]. Changes in salinity also occurred, although these are harder to quantify. It is generally believed that the basin salinity increased, although the size of the increase is questioned. Some researchers have postulated a large increase of several practical salinity units (psu) [(e.g., *Bethoux*, 1984; *Thunell and Williams*, 1989; *Rohling and Bryden*, 1994], while others have suggested only a small increase [*Bigg*, 1994, 1995].

One previous attempt to model the circulation of the Mediterranean at the LGM using an ocean general circulation model (OGCM) was by *Bigg* [1994]. He used a fairly coarse resolution (for the Mediterranean) version of the Bryan-Cox [*Cox*, 1984] OGCM, with a spatial resolution of 0.5° and 18 vertical levels. The model was forced by the output from the atmospheric GCM of *Kutzbach and Guetter* [1986], spatially and temporally interpolated onto the ocean grid, using an interactive heat and salt flux scheme. *Bigg* [1994] found that his modeled sea surface temperatures were similar to the reconstructions, but that his modelled sea surface salinity (SSS) fields were at most only 0.1-0.2 higher than today. He also found a complete loss of stability in the western basin in winter, with WMDW overflowing the Strait of Sicily to fill the deep eastern basin. However, the model had a poor initial Mediterranean climatology which was significantly too fresh and warm. The present-day climate simulation from this model also completely lacked deep water formation in both the western basin and especially the eastern basin, where no Adriatic deep water was formed.

1.2. The Holocene and Sapropel S_1

Within the sedimentary record obtained by deep-sea coring, numerous layers of black, often laminated, sediments rich in organic matter (sapropels) have been found between the normal pelagic sediments [*Kullenberg*, 1958]. Sapropels contain abundant and well-preserved calcareous microfossils of planktonic origin but are mostly devoid of benthic fossils [*Castradori*, 1993; *Rohling*, 1994]. They have been found throughout the eastern Mediterranean, with upper depth limits (for the overlying water column) quoted at 700 m [*Thunell et al.*, 1984], to 300 m [*Rohling and Gieskes*, 1989], to as shallow as 150 m in the North Aegean Sea [*Perissoratis and Piper*, 1992]. Their presence suggests anoxic conditions below the specified depths. Sapropels have also been found in the Tyrrhenian Sea [*Thunell et al.*, 1984; *Castradori*, 1993] and possibly other parts of the western Mediterranean [*Rohling*, 1994]. Sapropels have been formed throughout the past 7 Myr [*Rohling and Hilgen*, 1991; *Nijenhuis et al.*, 1996], with at least 11 formed during the last 450 kyr [*Cheddadi and Rossignol-Strick*, 1995b]. Although the majority were formed during climatic warm periods, several were formed under glacial conditions, and astronomical forcing is thought to control the timing of Sapropel formation, associated with minima of the precession cycle which occur about every 21,000 years [*Rohling and Hilgen*, 1991]. These astronomical conditions create changes in the monsoon, and thus precipitation and runoff patterns in the Mediterranean watershed.

The most recent sapropel S_1 was formed during the

Holocene, a period of warm and wet climate in the Mediterranean region [*Mangini and Schlosser*, 1986; *Rohling and Hilgen*, 1991]. It was deposited between ~9000 and 6000 years B.P. [*Jorissen et al.*, 1993; *Troelstra et al.*, 1991; *Perissoratis and Piper*, 1992; *Fontugne et al.*, 1992], on the basis of accelerator mass spectrometry (AMS) ^{14}C dating. Several studies suggest that the processes culminating in the formation of S_1 started much earlier [e.g., *Howell and Thunell*, 1992; *Rohling et al.*, 1993], with the most recent minimum in the precession cycle occurring at 11.5 kyr B.P. [*Berger and Loutre*, 1994].

Studies of $\delta^{18}O$ records [e.g., *Vergnaud-Grazzini et al.*, 1977; *Cita et al.*, 1977; *Mangini and Schlosser*, 1986] suggested that eastern Mediterranean surface salinities were very low at the time of the formation of S_1, with a decrease of up to 4.0 from present. A number of causes could underlie these low surface salinities: enhanced Nile discharge [*Rossignol-Strick et al.*, 1982; *Rossignol-Strick*, 1985], outflow from a recently connected Black Sea [*Olausson*, 1961; *Lane-Serff et al.*, 1997] and/or increased precipitation over the Mediterranean and its borderlands [*Rossignol-Strick*, 1987; mar*Rohling and Hilgen*, 1991]. Any of these would lead to a decrease in the excess evaporation over precipitation and runoff in the eastern Mediterranean. Very low surface salinities would act to decrease the densities of surface waters and hence reduce the potential for convective overturning and deep water formation. With no abyssal ventilation the oxygen content of the deep waters would gradually decrease as it was utilized for oxic degradation of organic matter. Consequent development of deep anoxia favors preservation of organic matter in the sediments and is consistent with the observed absence of benthic fossils from sapropelic sediments.

Two possible thermohaline circulation patterns have been suggested as likely to be associated with fresh (and buoyant) surface waters and anoxic deep waters. It is possible that the upper level circulation was antiestuarine, as it is today, but with limited LIW and Adriatic water formation ventilating only to intermediate depths [*Rohling and Gieskes*, 1989; *Rohling*, 1994]. This would be associated with decreased transport through the Straits of Sicily and Gibraltar.

Another suggestion is that the thermohaline circulation of the Mediterranean could have reversed to estuarine, with water inflowing at depth through Gibraltar and Sicily, upwelling, and then exiting as a surface outflow [e.g., *Thunell and Williams*, 1989]. Without local ventilation and oxygen supply to the abyss the deep basins of the Mediterranean would act as nutrient traps [*Sarmiento et al.*, 1988]. The geologic evidence, however, does not support a reversal in circulation through the Strait of Gibraltar, but rather a weakened antiestuarine circulation [*Zahn et al.*, 1987; *Vergnaud-Grazzini et al.*, 1989].

Other work suggests that the productivity of the Mediterranean had substantially increased during sapropel formation [e.g., *de Lange and ten Haven*, 1983; *Calvert*, 1983]. Drawing a parallel to conditions in the present-day Black Sea, *Pederson and Calvert* [1990] suggest that it is high productivity, not bottom water anoxia, that controls the increase of organic carbon in sediments. The enhanced productivity may have been related to enhanced riverine nutrient fluxes [*Rohling and Hilgen*, 1991] compounded with basin-wide shoaling of the pycnocline into the euphotic zone [*Rohling*, 1994]. The consequent enhanced rain of organic debris from the euphotic zone may not have been fully decomposed below, thus forming organic-rich sediments.

The general consensus now seems to be that sapropel formation is related to a combination of both anoxic bottom waters and enhanced primary productivity [*Rohling and Gieskes*, 1989; *de Lange and ten Haven*, 1983; *Howell and Thunell*, 1992]. For example, *Rohling* [1994] suggests that although pycnocline shoaling can lead to enhanced primary production, it cannot trigger sapropel deposition on its own.

1.3. Overview

Despite the effort that has gone into developing possible explanations for the formation of sapropels in the Mediterranean, no one (to our knowledge) has attempted to simulate the oceanic circulation during the period of sapropel deposition using an oceanic general circulation model (OGCM). We shall use an OGCM to examine the circulation of the Mediterranean during Holocene for the first time, and we shall also reexamine the circulation during the LGM. These two periods are of interest because they may be considered as two opposite extremes of Mediterranean climate, with the cold and arid LGM and the warmer and wet Holocene.

The model and some numerical choices are discussed in section 2. Results of a present-day control run are described in section 3. Paleo forcing fields and boundary conditions for the Holocene experiments are given in section 4, while the Holocene results are shown in section 5. LGM forcing fields and boundary conditions are described in section 6, with the results using these fields provided in section 7. Further discussion of the results and a summary are given in section 8.

2. Numerical Model Formulation

The model used is the Modular Ocean Model - Array (MOMA), a Bryan-Cox-Semtner-type OGCM using the *Killworth et al.*[1991] free surface scheme and revised horizontal and vertical advection schemes in the baroclinic momentum equation. The tracer advection schemes are modified to include the *Gent and McWilliams* [1990] eddy parameterization and a flux-limiting scheme based on the work of *Thuburn* [1996] and Stratford (A note on the implementation of flux-limited advection for tracers in an ocean circulation model, submitted to *Journal of Oceanic and Atmospheric Technology*, 1998). The basic model is described in greater detail by *Webb* [1993] and was used in the Mediterranean by *Myers et al.*[1998] to examine the sensitivity of the basin circulation to the wind stress forcing.

The model setup is based on the work of *Haines and Wu* [1995] and *Wu and Haines* [1996, 1998], who were able to reproduce an accurate representation of the present-day circulation. They demonstrated the importance of higher resolution, especially with regards to water mass dispersal, and we follow their lead with a horizontal resolution of 0.25° × 0.25° and 19 vertical levels (mainly concentrated in the upper water column to resolve the thermocline). The horizontal biharmonic viscosity coefficient is $A_h = 1.5 \times 10^{10}\ m^4\ s^{-1}$. The vertical momentum diffusion is $A_v = 1.5 \times 10^{-4}\ m^2\ s^{-1}$. The *Gent and McWilliams* [1990] thickness diffusion parameter is 20.0 $m^2\ s^{-1}$, and the maximum (reciprocal) slope of isopycnals is 100.0. Convective adjustment is performed using the complete convection scheme of *Rahmstorf* [1993]. To handle the exchanges with the Atlantic, a small box is added outside of Gibraltar, where the temperature and salinity at all depths are relaxed on a 1 day timescale to the climatological values.

Our topography and coastlines are developed from the present-day Mediterranean Models Evaluation Experiment (MEDMEX) data set [*Beckers et al.*, 1996] by *Wu and Haines* [1996], with a narrow Strait of Gibraltar of width 25 km (one grid point). Present-day wind stress data are obtained from the European Centre for Medium Range Weather Forcasting (ECMWF) and are based on a 7 year climatology (1986-1992) reanalyzed monthly (see *Myers et al.*[1998] for more details).

Haney [1971] relaxation conditions are applied at the sea surface for temperature and salinity. This type of boundary condition, while having some limitations, is widely used in ocean modeling because of the poor quality of air-sea fluxes. The required surface tracer fields, given monthly, are linearly interpolated by the model to give a value for each time step, and these repeat every year. The temperature relaxation timescale is 2 hours (acting on a top layer of 10 m thickness). The relaxation timescale is 5 days for salinity, except in the Levantine, where it smoothly decreases to 2 hours east of 23°E [*Wu and Haines*, 1996]. This change in the Levantine makes little difference for the paleoscenarios (verified with an additional experiment), although it is important to produce a realistic present-day simulation. For a CONTROL simulation we relax toward present-day surface T and S data based on the Mediterranean Oceanic Data Base (MODB) — MED5 [*Brasseur*, 1995]. This has an annual average surface salinity in the western Mediterranean of 37.6, and an annual average surface salinity of 38.7 in the eastern basin.

3. Present-Day Control Experiment

The run, CONTROL, has modern day bathymetry, surface salinity and temperature fields and provides a baseline against which to compare our paleoceanographic experiments. CONTROL was integrated for 100 years (the overturning timescale of the Mediterranean) to allow the system to reach a statistical steady state. Annual average fields over the last year of integration are shown. The main direct cell of the thermohaline circulation consists of Modified Atlantic Water (MAW) flowing east through the basin near the surface (getting saltier through excess evaporation) to the Levantine, where in winter it is converted into LIW which returns to the west at intermediate depths.

Separate deep thermohaline cells exist in the eastern and western basins, where LIW is provided to the Adriatic and Gulf of Lions, respectively, to precondition the water column for deep water formation. In the eastern basin, Figure 2a shows a salinity section (path inset) with a thick layer of salty LIW flowing into the Adriatic and fresh EMDW flowing out through the Strait of Otranto and sinking to the bottom to fill the deep Ionian and Levantine with a very uniform water mass. In the west, LIW flows north along the west coasts of Sardinia and Corsica to the Gulf of Lions, where it is converted into WMDW, which fills the deep western basin (Figure 2b).

A quantitative measure of the strength of the circulation is obtained from the mass, heat, and salt transport through the straits (Table 1). The transports for Sicily and Otranto are very much in line with observations (discussed in *Wu and Haines* [1996]) while our value for Gibraltar agrees with some estimates [*Lacombe and Richez*, 1982; *Bethoux*, 1984] but is significantly higher than some others [*Bryden and Kinder*, 1991]. The large transport is in part because the model strait is wider and, being rectangular, has a greater cross-sectional area than in reality.

Surface fluxes of heat and freshwater in the model can be recovered from the *Haney* [1971] boundary conditions and used to calculate basin averages. These fluxes (Table 2) show the reason for the strong anti-estuarine circulation present in the Mediterranean, a strong surface buoyancy loss in both main basins. The evaporation minus precipitation (E-P) figures (noting that runoff (R) is implicitly included) agree very well with a number of observational estimates from different sources [*Bryden and Stommel*, 1982; *Gilman and*

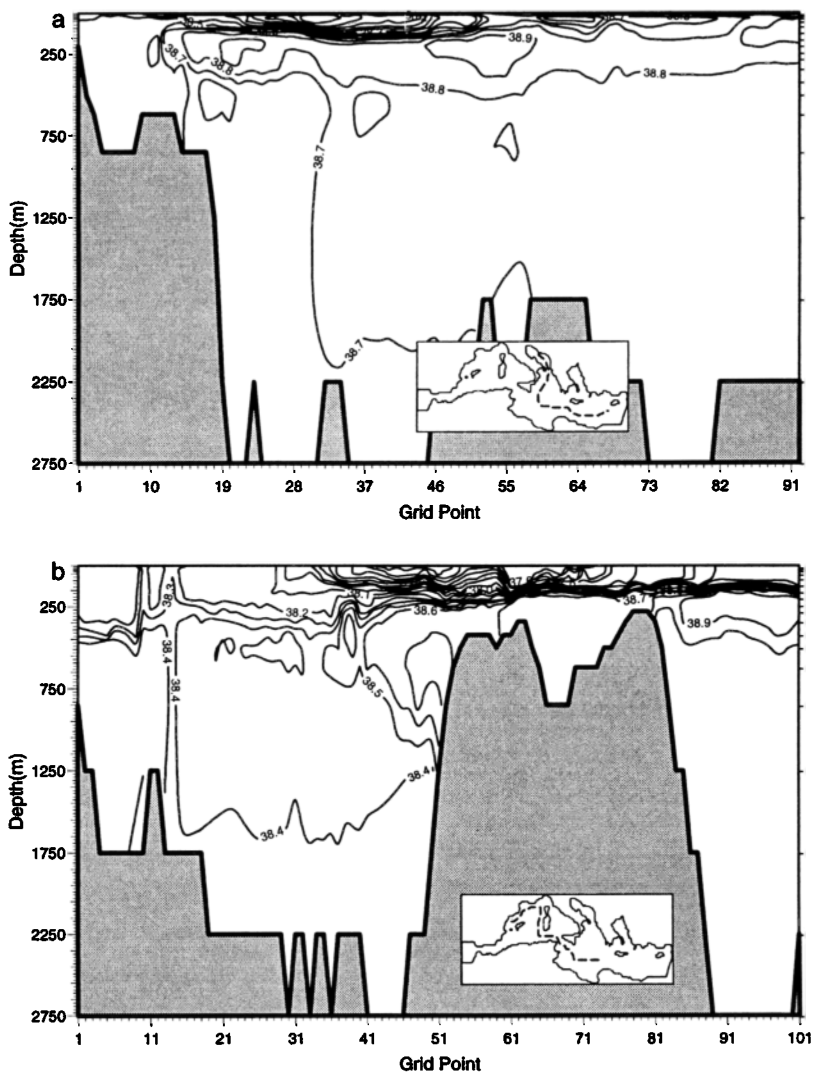

Figure 2. Salinity cross sections for the CONTROL experiment along the paths shown in the inset: (a) annual mean along a transect through the eastern Mediterranean and Adriatic and (b) snapshot along a transect through the Strait of Sicily and the western Mediterranean during late winter (March). The contour interval is 0.1 in both.

Garrett, 1994]. The model heat fluxes lie within the uncertainty found in observations made from hydrological measurements at Gibraltar [*Macdonald et al.*, 1994] and from bulk formulas [*Bethoux*, 1979; *Gilman and Garrett*, 1994]. For more details of the present-day simulation, see *Wu and Haines* [1998] or *Myers et al.*[1998]. The excellent agreement of the flux data adds confidence that the technique of relaxing to estimates of paleo surface properties can provide accurate flux and circulation estimates.

4. Holocene Paleoforcing and Boundary Conditions

One of the most basic changes that has occurred over the past 20,000 years is the difference in sea level associated with deglaciation. During the period of S_1 formation the sea level on average stood ~20-25 m lower than today [*Fairbanks*, 1989]. To apply the sea level reductions, the depth at each grid point is reduced to the nearest level that this sea level drop would entail.

It is generally agreed that the near-surface salinities in the Mediterranean were lower during the Holocene than at present [*Thunell and Williams*, 1989; *Kallel et al.*, 1997; *Rohling and De Rijk*, 1998]. However, there are some differences in magnitudes and spatial structure between reconstructions. For our main Holocene experiment (HOL1) we use the salinity reconstruction of *Kallel et al.*[1997] with its intermediate change in the mean zonal salinity gradient (which suggests the west-east salinity gradient was approximately zero at the time of S_1). We examine the sensitivity of our re-

Table 1. Mass, Heat, and Salt Transport Through Each of the Straits of Gibraltar, Sicily and Otranto (Mouth of the Adriatic Sea) in Each Experiment

Strait	Experiment	Mass, Sv	Heat, 10^{12} W	Freshwater, Sv
Gibraltar	CONTROL	1.49	13.97	0.059
	HOL1	0.73	7.56	0.016
	HOL2	0.94	8.33	0.019
	HOL3	0.73	6.71	0.015
	HOL4	1.49	16.34	0.020
	LGM1	0.61	5.19	0.030
	LGM2	0.73	10.97	0.033
	LGM3	0.93	9.57	0.023
Sicily	CONTROL	0.98	7.05	0.031
	HOL1	0.32	3.26	0.002
	HOL2	0.57	3.02	0.036
	HOL3	0.20	1.37	-0.002
	HOL4	0.29	2.12	0.001
	LGM1	0.49	2.47	0.023
	LGM2	0.54	1.81	0.025
	LGM3	0.53	1.74	0.009
Otranto	CONTROL	0.43	3.80	-0.010
	HOL1	0.27	2.52	-0.003
	HOL2	0.29	2.69	-0.008
	HOL3	0.19	2.24	-0.002
	HOL4	0.17	0.90	-<0.001
	LGM1	0.05	0.26	-0.001
	LGM2	0.06	0.32	-0.001
	LGM3	0.25	-0.36	-0.003

In all cases, the values given are 10 year averages. The sign convention for the heat and freshwater transports is that a positive value indicates transport into the basin, and a negative value is a transport out (Mediterranean, eastern Mediterranean, and Adriatic).

sults by using two additional reconstructions with positive and negative zonal salinity gradients. *Rohling and De Rijk* [1998] suggest a smaller change from present-day conditions, retaining a significant eastward salinity increase, while *Thunell and Williams'* [1989] reconstruction suggests a reversed salinity gradient (with a salinity increase to the west). The surface fields are obtained as follows.

Vergnaud-Grazzini et al.[1989] find that in the Gulf of Cadiz, there was an oxygen isotope depletion of the order of 0.5 ppt during S_1. Using a modern day salinity-oxygen isotope relationship, valid for the eastern Atlantic near Gibraltar, $\delta_w = 0.41S$ [*Kallel et al.*, 1997] suggests a freshening of 1.2. We apply this modification to the MED5 data in the Atlantic box for all months. *Kallel et al.*[1997] suggest that the sea surface salinity was almost uniform over the entire Mediterranean during S_1 with an average salinity of ~36.0 [see *Kallel et al.*, 1997, Figure 6b]. To preserve some spatial structure, we do not use a homogeneous field but instead fit an exponential curve for a zonal salinity decrease to the basin such that the majority of the decrease is concentrated in the east and Levantine, giving us a ΔS for each grid box based upon its longitude. This ΔS is then applied to the present day monthly MED5 salinity data to produce the paleo surface salinity restoring fields. This gives a basin averaged surface salinity of 36.0, with a west-east gradient of 0.1 (i.e., very small, consistent with *Kallel et al.*[1997]). Despite this small gradient, significant spatial structure still exists, especially in winter [Figure 3a].

Experiment HOL2 is based on the reconstruction of *Rohling and De Rijk* [1998], who suggest from foraminiferal oxygen isotope records that the $\delta^{18}O(\delta_w)$ concentration in the western Mediterranean was depleted by ~0.5 ppt, not taking into account the effect of temperature changes. They also suggest a depletion of 0.9 ppt in the eastern Mediterranean and 1.2 ppt in the Levantine. Using a newer S:δ_w ratio of 0.5, as suggested by *Rohling and De Rijk* [1998], this leads to a decrease in salinity in the Levantine of 2.4. Using these salinity decreases as typical of the western and eastern Mediterranean, we again fit an exponential curve to produce a smooth variation in ΔS across the Mediterranean, which is again applied to the monthly MODB5 data. This gives a west to east annually averaged salinity gradient of 0.7, similar to that suggested by *Rohling and De Rijk* [1998] (Figure 3b shows the conditions in January).

Our final reconstruction, HOL3, is based on δ_w measurements of *Thunell and Williams* [1989], who used two foraminifera species and suggested that the present west-east salinity gradient had switched sign at ~8000 BP. At this time they estimated that the mean salinity was 36.5 in the western basin and 35.9 in the eastern basin. On the basis of this salinity structure they postulated a reversed thermohaline circulation. Again assuming an exponential zonal salinity profile, we produce a surface salinity field with a basin averaged value of 36.6 in the western basin and 36.0 in the eastern basin (Figure 3c for January). The gradient of −0.6 is the same as that estimated by *Thunell and Williams* [1989].

Kallel et al.[1997] suggest that in the Holocene a maximum temperature decrease from present of ~2.5°C cooling occurred in the northern Tyrrhenian and also the Gulf of Lions (M. Paterne, personal communication, 1998) with smaller changes in the Alboran and the rest of the western basin. They found few changes in the eastern basin. We therefore use the present-day surface temperature field in most of our model runs but later examine the sensitivity of our results to the above changes in paleotemperatures in run HOL4. Using the *Kallel et al.*[1997] temperature reconstruction gives colder winter temperatures in the northern parts of the western basin (Figure 4).

We use the present-day wind stress in most of our

Table 2. Surface Heat, Freshwater Fluxes and the Associated Buoyancy Loss (in Brackets) for Our Experiments, Averaged Over 15 Years

Experiment	Heat Flux W m^{-2} (10^{-9} m^2 s^{-3})				E-P-R cm yr^{-1} (10^{-9} m^2 s^{-3})			
	MED	WMED	EMED	ADR	MED	WMED	EMED	ADR
CONTROL	-6.1	-7.3	-5.1	-40.8	76	94	71	-20
	(2.9)	(3.5)	(2.4)	(19.6)	(6.7)	(8.2)	(6.2)	(-1.8)
HOL1	-4.0	-5.5	-3.4	-36.0	8	23	0	-12
	(1.9)	(2.6)	(1.6)	(17.3)	(0.7)	(2.0)	(0.0)	(-1.0)
HOL2	-3.6	-2.8	-4.1	-30.3	20	27	16	-23
	(1.7)	(1.3)	(2.0)	(14.5)	(1.8)	(2.4)	(1.4)	(-2.0)
HOL3	-4.3	-9.3	-1.9	-26.2	5	27	-7	8
	(2.1)	(4.5)	(0.9)	(12.6)	(0.4)	(2.4)	(-0.6)	(0.7)
HOL4	-10.0	-27.8	-0.9	-10.8	16	44	1	-7
	(4.8)	(13.3)	(0.4)	(5.2)	(1.4)	(3.9)	(0.01)	(-0.6)
LGM1	-0.2	-1.6	0.6	-3.3	73	55	80	-14
	(0.1)	(0.7)	(-0.4)	(1.6)	(6.5)	(4.9)	(7.1)	(-1.8)
LGM2	-2.6	-6.0	0.0	-2.0	69	42	81	-26
	(1.2)	(2.9)	(0.0)	(0.9)	(6.1)	(3.7)	(7.2)	(-2.3)
LGM3	-4.1	-7.7	-1.2	12.0	64	29	79	-116
	(2.0)	(3.7)	(0.5)	(-5.8)	(5.7)	(2.6)	(7.0)	(-10.3)

The regional abbreviations are MED, basin average; WMED, the western Mediterranean; EMED, the eastern Mediterranean (including the Adriatic and Aegean); ADR, the Adriatic, and E-P-R, evaporation-precipitation-runoff.

paleoexperiments but also test the sensitivity to winds in experiments HOL4. We obtained paleowinds from a 6 kyr B.P. simulation using the atmospheric GCM of *Dong and Valdes* [1995]. The monthly averaged wind stresses are linearly interpolated from 2.8125° × 2.8125° onto the much higher resolution model grid. The Holocene wind stresses are similar to the present day, but some differences exist, including a more zonal flow over the northwestern basin in winter and a westward shift of the cyclonic circulation in the eastern basin. Overall, differences in magnitude are not large.

5. Holocene Circulation and Sapropel S_1

In all Holocene experiments the sea level has been reduced by 20 m, and the surface salinity restoring fields are based on the three different reconstructions in Figure 3 [*Kallel et al.*, 1997; *Rohling and De Rijk*, 1998; *Thunell and Williams*, 1989], each with its own west-east salinity gradient (Table 3). The additional experiment, HOL4, includes changes in the paleotemperatures and paleowinds with the *Kallel et al.*[1997] surface salinities. In all cases we will only examine the large-scale features of the circulation, as we feel it is impossible to simulate detailed paleostructure for the Mediterranean because of the uncertainties, particularly in the forcing. A separate experiment (not shown) with no changes other than the small change in sea level produced little change in the model's circulation. Present-day initial conditions are used, and the deep waters are therefore saltier and denser than the Holocene surface forcing might produce. However, this is realistic because old deep waters remaining after the deglaciation were probably present during the early Holocene [*Mangini and Schlosser*, 1986; *Rohling*, 1994].

Each experiment was integrated for 40 years until an equilibrium circulation was reached. One experiment, HOL2, was integrated for an extra 40 years to verify that this circulation was robust, and no trends were present. Results are shown annually averaged over the last year of integration. Despite the strong differences from the present-day surface salinity fields a fairly similar to present thermohaline circulation is set up within the eastern Mediterranean under all three reconstructions. Salinity and temperature cross sections are shown only from HOL1, in Figure 5, and any differences for the HOL2 and HOL3 experiments are noted. LIW production is shifted to the Aegean north of Crete, where there is high wintertime surface salinity (see Figure 3a). The depth of convection has decreased, with the base of the LIW only reaching 75-125 m. Little water leaves the Aegean except for a weak flow along the Greek coast toward the Adriatic. Wintertime convection in the Adriatic is limited to 300 – 400 m, producing a cold and relatively fresh outflow. This water mass ventilates the whole eastern basin (except for the Aegean, which it does not penetrate) between 200 and 450 m (slightly shallower in HOL3 with the *Thunell and Williams* [1989] reconstruction), clearly seen as a core of low temperature water (Figure 5b). Other than the changes in properties (and depth), this water mass behaves similarly to the present day EMDW, flowing

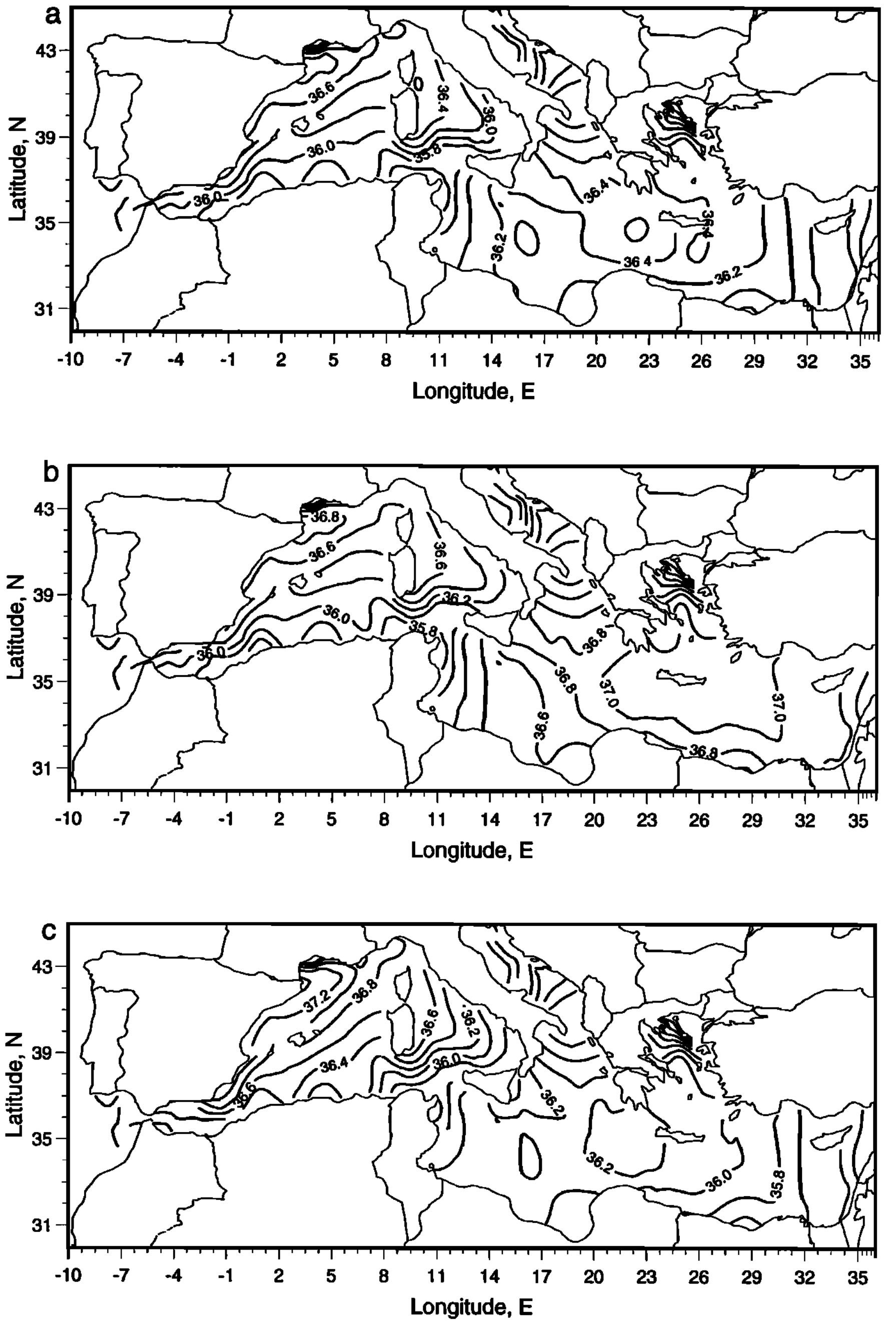

Figure 3. Surface salinity restoring field for the Holocene in January for the reconstructions of (a) HOL1 [*Kallel et al.*, 1987] (b) HOL2 [*Rohling and De Rijk*, 1998] and (c) HOL3 [*Thunell and Williams*, 1989]. The contour interval is 0.2.

cyclonically in a western boundary current around the Ionian and then into the Levantine. Although existence of a similar water mass has in the past been suggested, under the name Eastern Mediterranean deep Intermediate Water (EMdIW) [*Rohling*, 1994], we prefer Adriatic Intermediate Water (AIW) to signify that it is a true intermediate water mass, found in the upper part of the water column and that it originates in the Adriatic. The depths of ventilation, to ~125 m in the Aegean Sea and ~200-450 m in the Adriatic Sea and open eastern Mediterranean, are consistent with observations of the upper depth limits for S_1 formation, which reach up to 150 m or so in the Aegean [*Perissoratis and Piper*, 1992] and up to 300 m in the open eastern Mediterranean [*Rohling and Gieskes*, 1989].

Salinity increases with depth at nearly all locations within the basin (Figure 5a). A strong halocline is formed at ~400-450 m, separating stagnant old waters

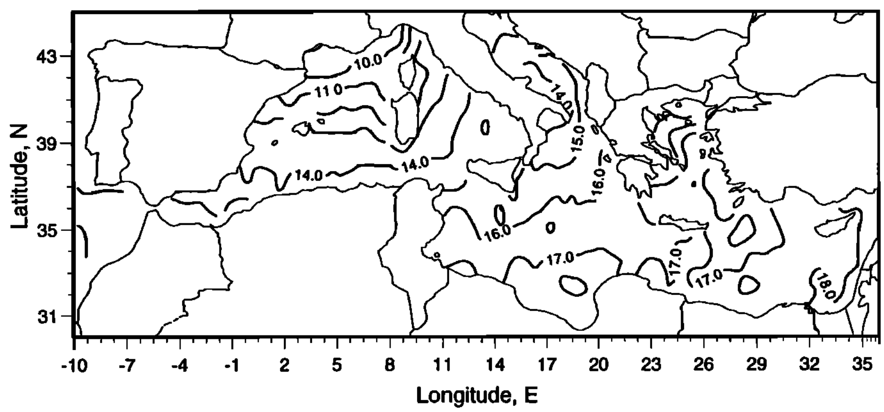

Figure 4. Temperature restoring field for the Holocene for January for HOL4, the reconstruction of *Kallel et al.*[1997]. The contour interval is 1.0°C.

from the well-ventilated upper ocean circulation. Although upward diffusive salt transport acts to significantly increase the salinity of the AIW from its fresh origins in the Adriatic, it is the regular resupply of this fresh water that buffers the upper water column and maintains the strong halocline.

Decreased amounts of MAW penetrate into the eastern Mediterranean, associated with reduced transport at Sicily Straits relative to CONTROL (Table 1). The transport decreases as the west-east salinity gradient decreases. Surprisingly, eastward surface flow persists even when the salinity gradient has reversed in HOL3. To balance the surface inflow to the eastern basin, there is a return flow to the west at depth (Figure 6). The water flowing to the west at Sicily is almost entirely composed of AIW (in contrast to LIW at present), and thus it is colder and fresher.

Despite the differences in the west-east salinity gradient between experiments the antiestuarine nature of the eastern Mediterranean persists throughout. No circulation reversal is seen at Gibraltar either, even though the Mediterranean ceases to act as a net concentration basin (Table 2). This is for two reasons: (1) with the very small net surface evaporative component to the buoyancy flux the surface forcing becomes thermally controlled (net cooling in the Adriatic and other localized regions is sufficient to drive water formation and a weakened version of the present thermohaline circulation) and (2) The effect of the old deep water is to act as a salt source, with strong upward diffusive transfer of salt throughout the water column (Figure 5a). The contribution of this transfer to the buoyancy budget of the AIW can be estimated from the eastern basin averaged vertical flux at 400 m. When converted to units equivalent to the surface E-P, the net buoyancy loss to the AIW is equivalent to an evaporation of 20.6 cm yr^{-1}. This transfer rate suggests a timescale for the dispersal of the old deep water of ~2000-3000 years, meaning this process should continue throughout the duration of the sapropel.

Table 3. A Summary of the Holocene Experiments Performed in This Paper

Experiment	ΔS Gradient
CONTROL	1.1
HOL1	0.1
HOL2	0.7
HOL3	-0.6
HOL4	0.1

For each experiment, the average west-east salinity gradient is listed.

There is a very deep halocline (~600 m) in the western Mediterranean (not shown) separating stagnant deep waters from those above. With little provision of salt from the east (Table 1), deep water formation in the Gulf of Lions does not occur. Instead, temperature-driven intermediate convection occurs in the Provencal Gyre and Balearic Basin, producing a cold intermediate water mass (Figure 7), ventilating to ~500 m in HOL1 and HOL2 and to 800-1000 m in HOL3. The deeper ventilation in this last experiment is related to the weaker vertical salinity gradients, thus allowing winter cooling to drive more penetrative convection. Still more ventilation occurs in HOL4, with its lower surface temperatures, reaching down to ~1200-1300 m. The changed winds play little role because there is little change to the cyclonic circulation in the Gulf of Lions. Transport at Gibraltar weakens (Table 1) with the decrease in buoyancy forcing except in HOL4 where the increased temperature component of the buoyancy forcing leads to larger freshwater import and a transport similar to the present day.

At present the Mediterranean pycnocline lies below the base of the euphotic layer [*Wüst*, 1961]. This can be

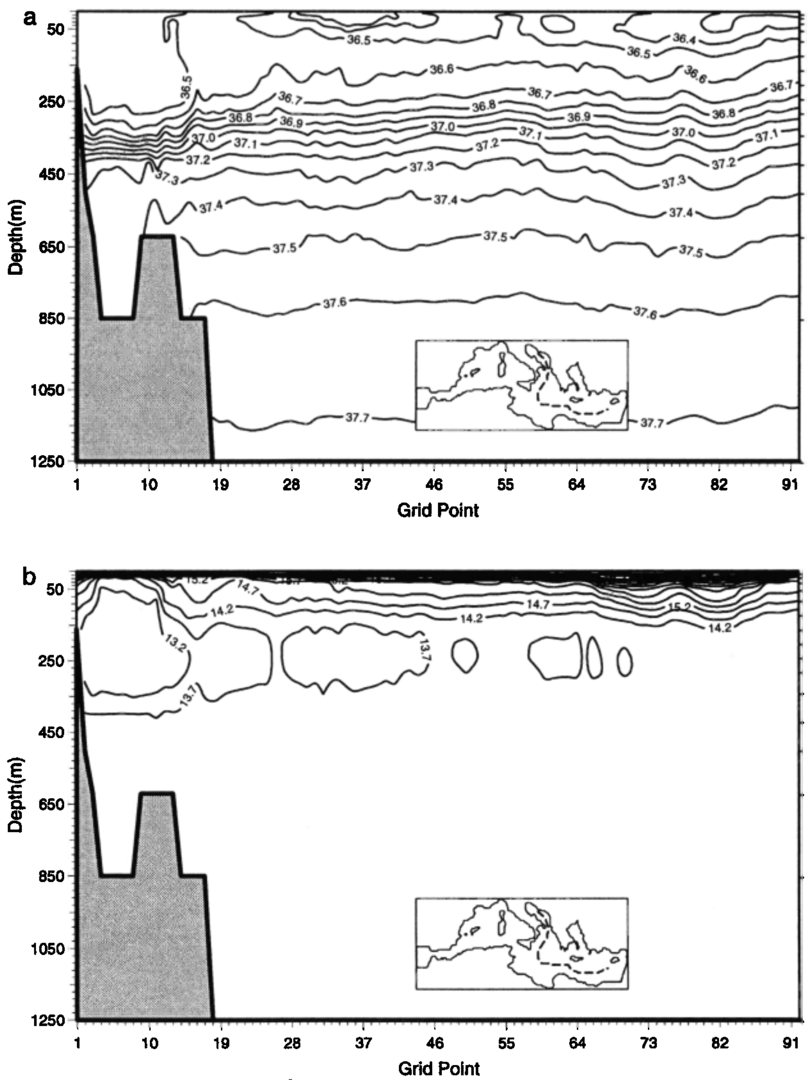

Figure 5. Annually averaged (a) salinity and (b) temperature along a transect (path shown in inset) through the eastern Mediterranean and Adriatic for experiment HOL1 [*Kallel et al.*, 1997]. The contour interval is 0.1 in Figure 5a and 0.5°C in Figure 5b.

examined in the model by looking at the Brunt-Vaisala frequency N given by

$$N^2 = -g\rho_0^{-1}\frac{d\rho}{dz} \tag{1}$$

where g is the acceleration due to gravity, ρ is the density, and ρ_o is a reference density. The Brunt-Vaisala frequency is a measure of the stability of the water column, which reaches a maximum at the pycnocline. In CONTROL, the pycnocline is indeed below the base of the euphotic layer (Figure 8a), at ~120-140 m in the Ionian and close to 200 m in the Levantine. The deep pycnocline is also found in the western basin. In experiments HOL1, HOL2 and HOL3 the pycnocline has shoaled to 50-80 m in the Levantine and even shallower in the northern Ionian, to within the euphotic layer (Figure 8b). Such shallow pycnoclines are in agreement with reconstructions based on micropaleontological evidence [*Rohling and Gieskes*, 1989, *Castradori*, 1993] and may support the theories of high primary production during sapropel formation.

6. LGM Paleoforcing and Boundary Conditions

It is estimated that sea level was 120 m lower than today at the LGM [*Fairbanks*, 1989]. To apply the sea level reductions, the depth at each grid point is again reduced to the nearest level that this sea level drop would entail.

Since semienclosed basins such as the Mediterranean are usually more sensitive (than the open ocean) to

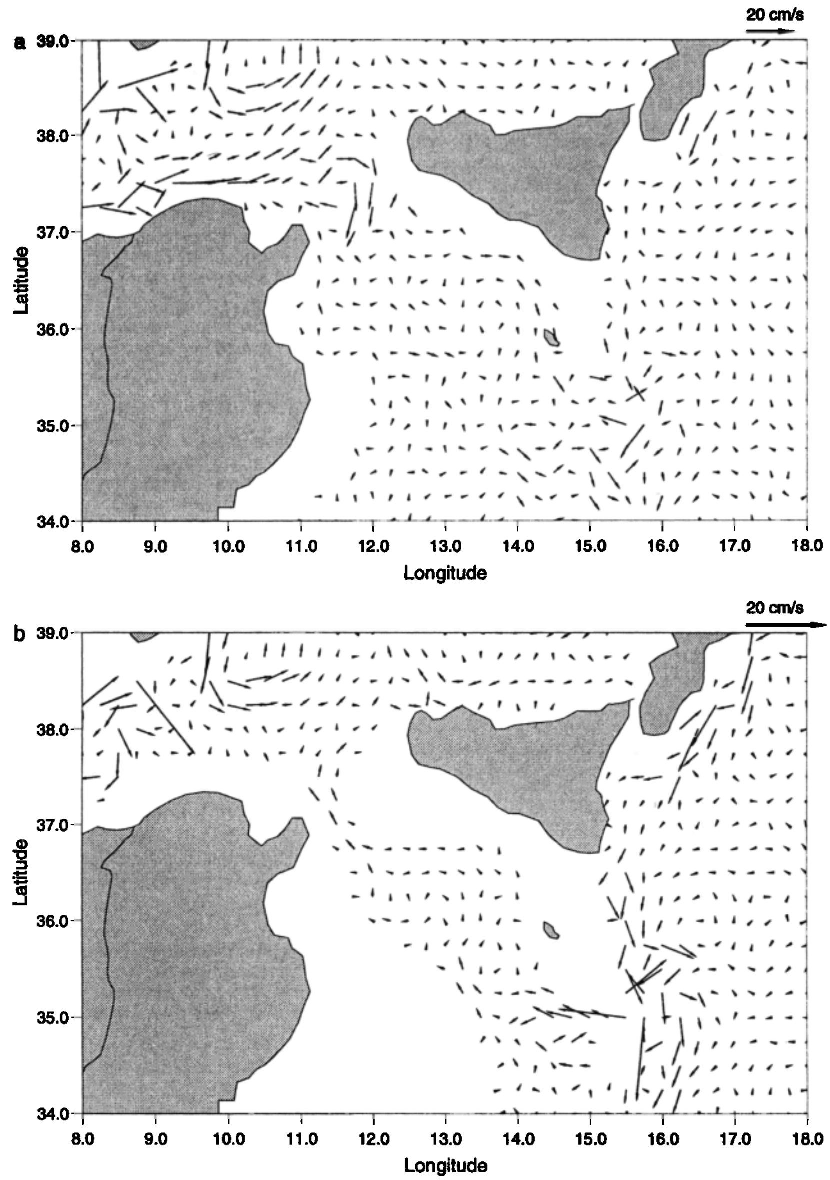

Figure 6. Annually averaged velocities through the Strait of Sicily at depths of (a) 40 m and (b) 220 m for HOL1 [*Kallel et al.*, 1997] showing the antiestuarine flow through the strait. Note that the scale arrow is different in each level.

changes in aridity and river runoff, *Thunell and Williams* [1989] suggested a salinity increase of 1.2 in the western basin and up to 2.7 in the eastern basin and Levantine. Using a simple heat budget model, *Bethoux* [1984] also found significantly higher salinities, especially in the eastern basin.

Recently, *Rohling and De Rijk* [1998] updated *Thunell and Williams'* [1989] work by incorporating new data and an extensive statistical assessment to distinguish significant trends. Their analysis helps to resolve some of the controversy over the magnitude of the salinity increase in the Mediterranean at the time of the LGM, as summarized by *Bigg* [1995]. Their reconstruction suggests relatively smaller salinity changes in the western basin, consistent with decreased evaporation related to the drop in temperature [*Bigg*, 1995], higher lake levels [*Street and Grove*, 1979], and altered precipitation patterns [*Prentice et al.*, 1992]. Meanwhile, significantly higher salinities in the Levantine can be explained with little change in evaporation (associated with only small decreases in temperature [*Thiede*, 1978]) but decreased precipitation in the region [*Goudie*, 1992] (although some of the lake level data are inconsistent with this [*Street-Perrott and Roberts*, 1983]) and

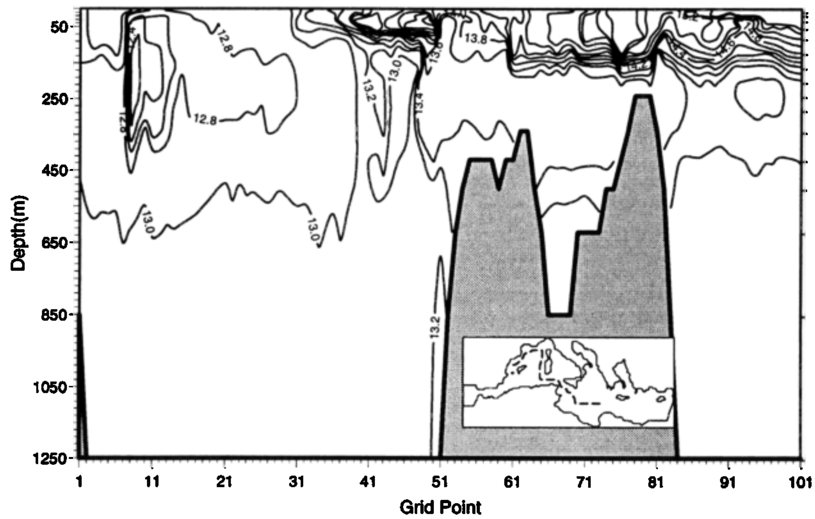

Figure 7. Late wintertime (March) temperature snapshot along a transect (path shown in inset) through the Strait of Sicily, western Mediterranean, and Gulf of Lions for experiment HOL2. The contour interval is 0.2°C.

significant decreases in runoff. Significantly decreased runoff, particularly from the Nile [*Adamson et al.*, 1980], would act to increase the net freshwater loss. Assuming a reduction in discharge by a factor of 4 from the pre-Aswan Dam discharge of 91 km^3 yr^{-1} [*Adamson et al.*, 1980], and applying this over the Levantine, produces an equivalent surface loss of 10 cm yr^{-1}. Such a flux, applied over a period of 1000 years (a period far shorter than the full extent of the dry glacial conditions in East Africa), would on its own lead to an increase of salinity in the Levantine by 1-2.

Using $\delta^{18}O$ in planktonic foraminifera, *Duplessy* [1993] showed that the salinity of the glacial eastern Atlantic was enriched by ~1.0 from present, which is applied to our Atlantic box. *Rohling and De Rijk* [1998] find an oxygen isotope enrichment in the Levantine of the order of 1.0 ppt, taking into account the correction for the glacial ice-volume effect and also for changes in temperature. Using a salinity oxygen isotope ratio of order 0.3 (from *Rohling and Bigg* [1998] based on data from *Pierre et al.*[1986] and *Pierre* [1998]) appropriate to the present-day Mediterranean, we find an increase of salinity in the Levantine of the order of 3.5. *Kallel et al.*[1997] suggest another salinity-oxygen isotope relationship valid for regions of the Mediterranean where evaporation is strong, such as in the Levantine. Their coefficient of 0.2 would lead to an even larger salinity value for the Levantine.

Using these two salinity increases (1.0 and 3.5) as end-members for the western and eastern ends of the Mediterranean, we fit an exponential curve to these values. A large coefficient in the exponential reduces the salinity increase over most of the basin, with the exception of the general glacial enrichment, except in the far east, as suggested by the observations. Using a similar procedure to that used for the Holocene, a surface ΔS is calculated and added to the monthly MED5 fields (Figure 9). This gives a west to east annually averaged salinity increase of 1.6, slightly larger than the present gradient, consistent with *Vergnaud-Grazzini et al.*[1988], which we use in experiment LGM1. The zonal gradient is steepest between the Ionian and the Levantine, where the bulk of the salinity increase is located. We also apply this paleosalinity modification to our initial conditions (MED5) at all depths. If we did not apply this initial salinity change, the present-day initial conditions would rapidly convect under the saltier paleoconditions, leading to large unstable transient features.

Bigg [1995] suggests that in the LGM the Mediterranean temperature was roughly 5°C colder than today in winter, with larger decreases in the west than the east. In determining the oxygen isotope depletions for changes in the salinity field, *Rohling and De Rijk* [1998] assumed coarse temperature corrections: -6 ± 2°C for the western basin, -5 ± 2°C for the eastern basin and -3 ± 2°C for the Levantine Sea, which includes a temperature difference of ~4°C between the northern and southern boundaries of the basin in the LGM. *Rohling et al.*[1998] find temperatures as much as 8°C cooler than today in the north west Mediterranean. *Paterne et al.*[1998] suggest seasonal temperature decreases of 3°-12°C in the western basin and Adriatic (largest decreases in summer) and 1°-5° C (with less seasonality) in the eastern basin and Levantine.

We elected to introduce the most extreme temperature estimates to assess the maximum impact on the thermohaline circulation in experiment LGM2. We

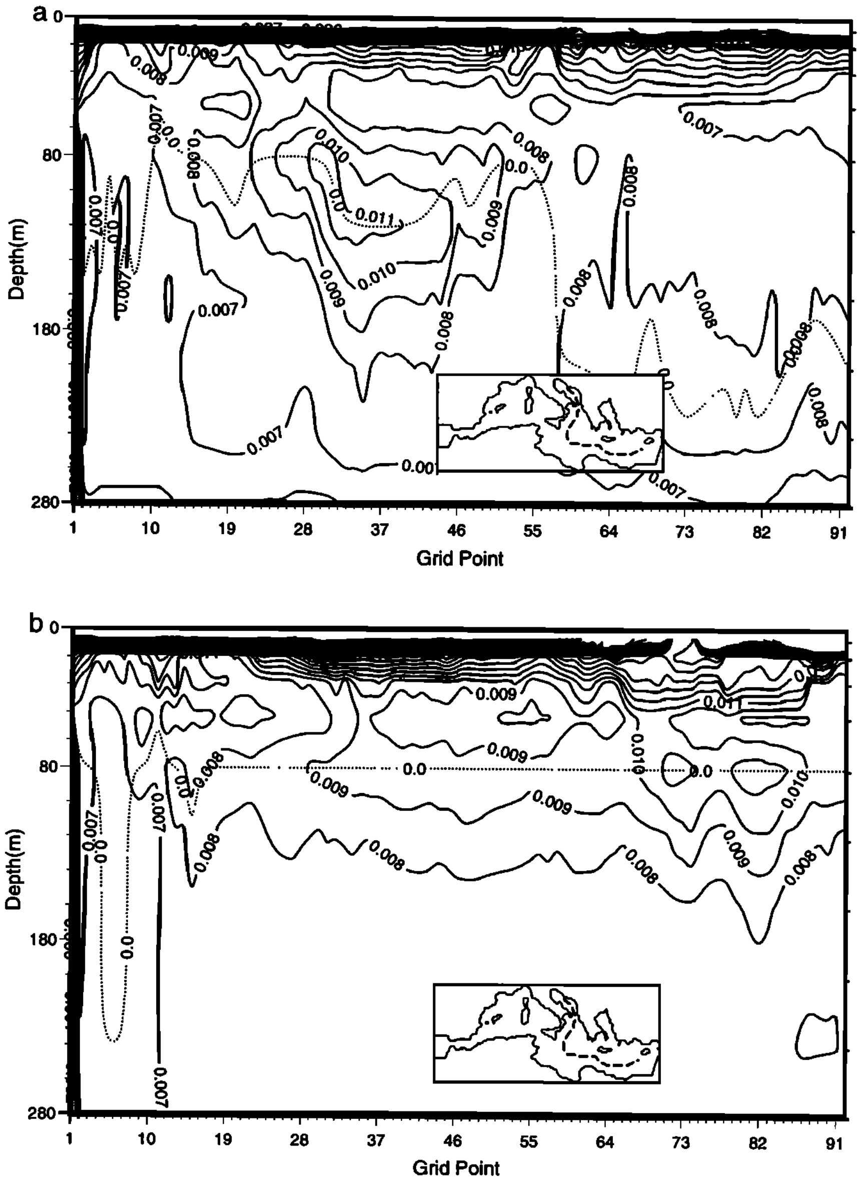

Figure 8. Plot of the Brunt-Vaisala frequency N for an annually averaged transect (path shown in inset) through the eastern Mediterranean and Adriatic showing the depth of the pycnocline for experiments: (a) CONTROL and (b) HOL1. The contour interval is 0.001 s^{-1}. The dashed line indicates the approximate position of the model pycnocline (to the nearest model level depth).

therefore use a temperature modification of −8°C in the north western basin, −7°C in the central basin (e.g., Adriatic), and −1°C in the southern extremity of the eastern basin (Levantine). Intervening points are determined by a similar exponential curve to the one used for salinity. A north-south gradient of 4°C is applied linearly, with the southern margins of the basin being 4°C warmer than the north. The same temperature modification is applied in all months for simplicity since it is the wintertime fields (Figure 10) that will determine any changes to the thermohaline circulation. The modifications are also applied to the model initial conditions for the same reasons as listed for the salinity above.

We test the sensitivity of our LGM results to paleowinds in LGM3. The 21 kyr B.P. paleowinds are obtained from simulations using the atmospheric GCM of *Dong and Valdes* [1995]. The monthly averaged wind stresses are linearly interpolated from 2.8125° × 2.8125° onto the much higher resolution model grid. These wind stresses show an increase in magnitude of the winds in winter, especially in the northern part of the basin. The mistral is significantly enhanced, and the stresses are more zonal (because of the coarse resolution atmospheric model data). The location of the main cyclone is shifted to northern Italy from over the Aegean, leading to a vigorous northwestward flow over the Aegean

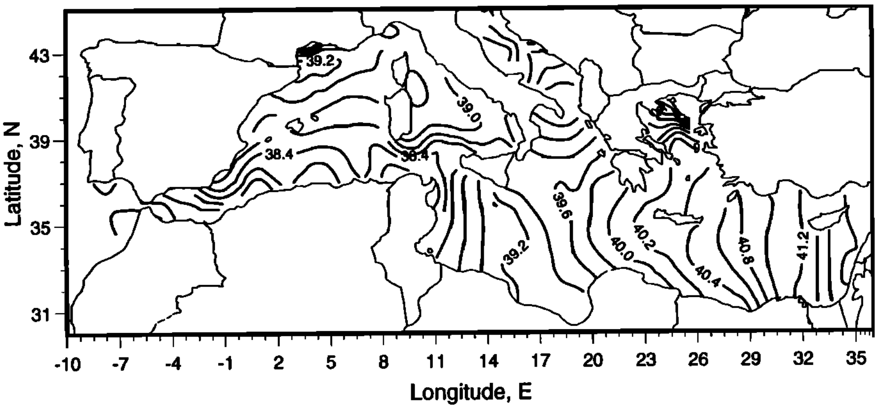

Figure 9. Surface salinity restoring field for the last glacial maximum (LGM) for January for all experiments. The contour interval is 0.2.

and Adriatic (compared to southwestward flow today). The flow in the Ionian is more zonal and is shifted to the north. The 21 kyr B.P. wind stresses are generally similar in direction and magnitude to the present in the western basin in summer but weaker over the eastern basin, except over the Adriatic where they are significantly strengthened.

7. The Last Glacial Maximum

In LGM1 both the bathymetry and surface salinity restoring fields are changed to LGM values, while sea surface temperatures (SSTs) are present-day. This increases significantly the salinity in the Levantine, with a smaller salinity increase elsewhere. This experiment was integrated for 40 years, allowing the basin's thermohaline circulation to be fully set up. An additional 100 year experiment (not shown) revealed that the circulation pattern set up after 40 years was stable and representative of the final "seasonal steady state". Results are summarized as the average over the final year of integration.

In the eastern basin, deep, rather than intermediate, convection now occurs in the Rhodes gyre region (Figure 11). The higher LGM salinities allow the wintertime cooling to increase the Levantine surface water density sufficiently for it to form bottom water. This 15°-15.6° C water mass (as compared to ~13.2°C for present-day deep waters formed in more northerly, colder regions) with high salinity, which we will call Levantine Deep Water (LDW), fills the deeper parts of the eastern basin. Throughout the basin it upwells to intermediate depths. The formation of deep water in the Levantine is not without precedent as *Sur et al.*[1993] found deep penetrative convection in the Rhodes gyre region during the anomalously cold winter of 1992.

The circulation in the Adriatic is completely reversed from present (Figure 11). The upper layer is fresh (compared to the surrounding eastern Mediterranean) and overlies salty and dense LDW that flows up over the Otranto sill to replace the exiting relatively fresh surface water. This stabilizes the stratification, allowing only weak, shallow, temperature-driven winter convection. Thus the Adriatic switches to an estuarine circulation. Deep-sea cores suggest that the deep LGM Adriatic likely was characterized by cold and stagnant con-

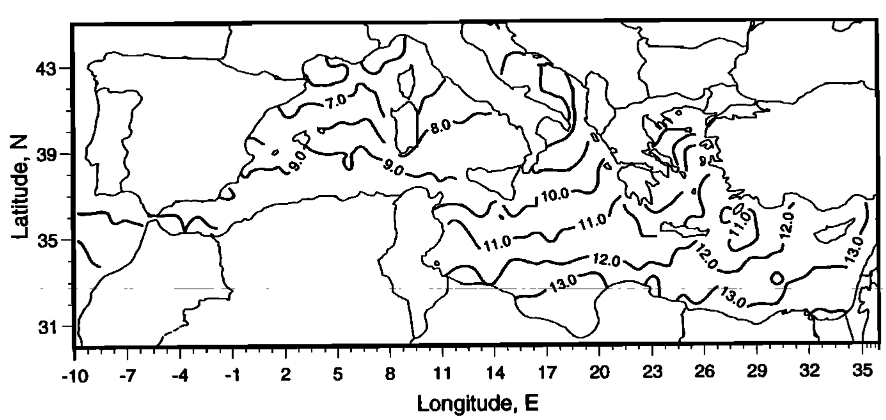

Figure 10. Surface temperature restoring field for LGM2 for January. The contour interval is 1.0°C.

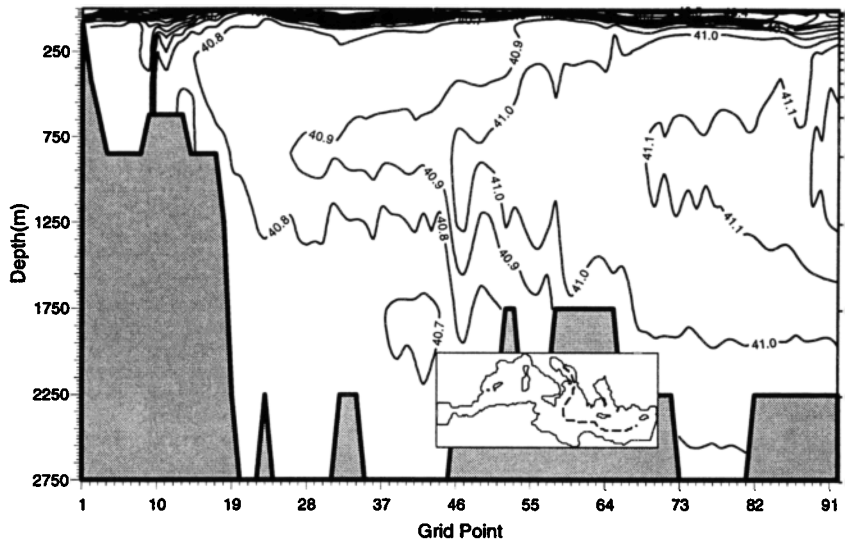

Figure 11. Annually averaged salinity along a transect through the eastern Mediterranean and the Adriatic (path shown in the inset) for LGM1 (LGM2 is similar). The contour interval is 0.1.

ditions, because of intense runoff flowing directly into the southern part of the basin [*Asioli*, 1996].

Unlike the Adriatic subbasin, for the eastern Mediterranean as a whole, an antiestuarine circulation prevails, but the transports through the straits of Sicily and Gibraltar are less than the present day (Table 1). After passing over the sill at Sicily the LDW sinks to fill the deeper parts of the Tyrrhenian and part of the Algerian basin. WMDW continues to form in the Gulf of Lions (Figure 12a), filling the deepest parts of the basin with cold water and preventing, for the most part, the passage of LDW to the far west. These findings are consistent with a well-ventilated western basin and a shallow thermocline, as inferred by *Vergnaud-Grazzini et al.*[1986] on the basis of enhanced productivity. With the passage of eastern water toward Gibraltar blocked by the WMDW the outflowing Gibraltar water is mainly upwelled WMDW (mixed with some LDW) and is less saline.

The basin average E-P is similar to present day (Table 1), although net evaporation has decreased in the western basin, mainly in the Tyrrhenian. In the eastern basin, E-P decreases in the Ionian but markedly increases in the Levantine, associated with the signifi-

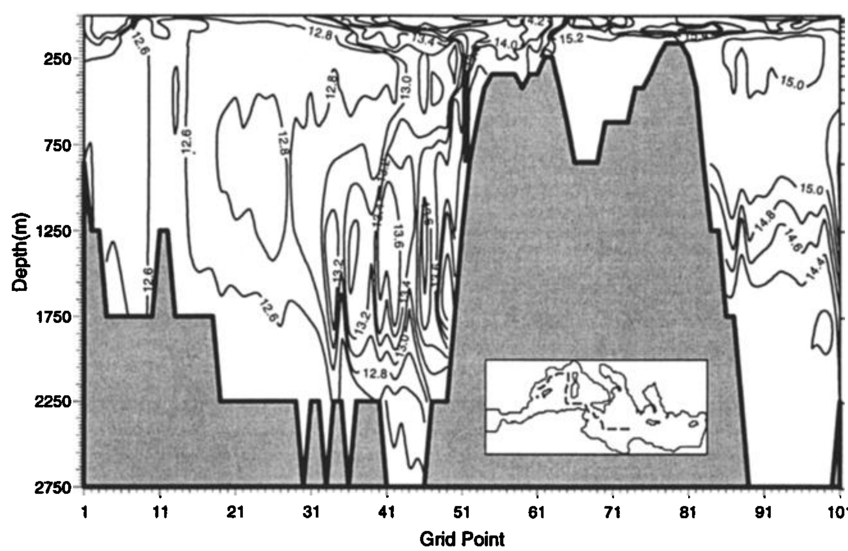

Figure 12. Late wintertime (March) temperature along a transect through the Tyrrhenian and the Gulf of Lions (path shown in the inset) for LGM1. The eastern basin is to the right. The contour interval is 0.2°C.

cantly higher surface salinities in this region. It is in this region that the MAW acquires the high salinities responsible for triggering deep convection in the Rhodes Gyre in winter.

The transports of mass, heat, and salt through the various key straits are given in Table 1. Transport through Gibraltar has decreased by a factor of 2 (from 1.49 to 0.61 Sv). This is what one would expect from the reduced cross-sectional area of the strait, associated with the sea level drop, assuming the flow is hydraulically controlled [*Rohling and Bryden*, 1994]. Decreased Mediterranean outflow during the LGM was postulated by *Vergnaud-Grazzini et al.*[1989]. The reduced transport at Sicily (from 0.94 to 0.49 Sv) is also as expected for a strait deeper and wider than Gibraltar. Flow through the Strait of Otranto has nearly collapsed (0.05 Sv), because of the lack of deep water formation in the Adriatic.

During the LGM, *Rohling and Giekes* [1989] argued that abundances of *Neogloboquadrina* imply that the permanent pycnocline had shoaled to a depth within the euphotic layer. In our LGM experiment the pycnocline has indeed shoaled (Figure 13) to a depth of 60-80 m in the Ionian, decreasing to ~50 m south of Crete and then increasing to the east into the Levantine. The pycnocline has also shoaled in the Alboran Sea from ~120-140 m in CONTROL to 80 m in the LGM experiment, as was first suggested by *Pujol and Vergnaud-Grazzini* [1989].

Rohling [1991] suggested that shoaling of the pycnocline in the LGM was due to glacial sea level lowering. Using a simple two- layer model of the eastern Mediterranean, he showed that the reduction in transport at Sicily associated with the shallower sill (and assuming present-day buoyancy forcing, although other possible buoyancy forcings were examined) led to a shoaling of the pycnocline to a depth of ~80 m, within the euphotic zone. To test if the shoaling found in our LGM experiment is due to the sea level lowering or to the change in salinity boundary conditions, we performed an additional experiment. Using the changed bathymetry (120 m sea level reduction) but present-day surface forcing, we find that the pycnocline in the eastern Mediterranean shoals to a depth of ~60 m, well inside the euphotic zone.

In a second experiment, LGM2, we examine the robustness of the above circulation to the large temperature changes associated with the LGM. In particular, we wish to determine if the formation of LDW and the estuarine circulation in the Adriatic are stable under the strong LGM thermal forcing which was concentrated in the northern regions of the basin, including the Adriatic. The 120 m sea level reduction and the modified restoring salinity fields are as in LGM1.

Deep convection still occurs in the Rhodes gyre, but the LDW formed is now colder (with a core temperature of 10.5°C) and denser. A greater fraction of the deep eastern basin is filled with this newly ventilated water. Despite the strong drop in surface temperature in the Adriatic (Figure 10), deep convection does not occur because of the very high salinity water flowing in at depth. Thus the Adriatic retains the estuarine circulation found in experiment LGM1 despite winter temperatures as low as 8°C.

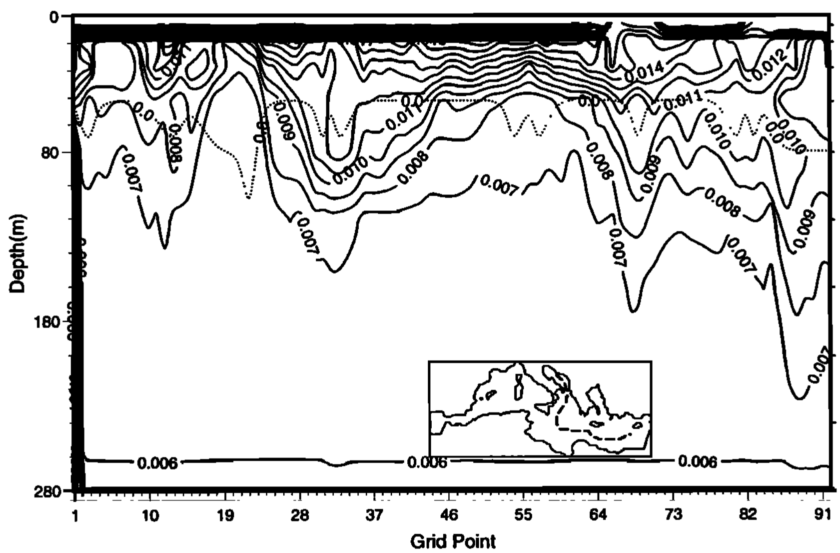

Figure 13. Plot of the Brunt-Vaisala frequency N for an annually averaged transect through the eastern Mediterranean and Adriatic (path shown in the inset) showing the depth of the pycnocline for LGM1 (LGM2 is similar). The contour interval is 0.001 s^{-1}. The dashed line indicates the approximate position of the model pycnocline (to the nearest model level depth).

There is little change at the Strait of Sicily from LGM1 (Table 1). After overflowing the sill the eastern water sinks into the Tyrrhenian and Algerian basins but not to as great a depth as in LGM1. The reason for this is the vigorous deep convection occurring in the Gulf of Lions, associated with the very cold winter surface temperatures in that region (Figure 10). With the WMDW being nearly 7°C cooler than the overflowing LDW at Sicily the eastern water mass is no longer able to sink to the bottom in the western basin, instead occupying more intermediate levels (~750-1500 m) in the Tyrrhenian and neighboring regions. The extra buoyancy forcing associated with the cold western temperatures leads to a greater outflow at Gibraltar (Table 1) and a greater heat import from the Atlantic. With the colder surface temperatures the surface heat loss over the basin has increased but there is little change in the net evaporation (Table 2). The pycnocline remains shoaled in many areas because this is related to the sea level changes more than to the differences in buoyancy forcing.

The paleosalinity reconstructions suggest that the salinity increase driving the thermohaline circulation at the LGM is located in the Levantine. This high salinity dense water, which acts as the driving engine for the thermohaline circulation, is very far away from the cold convective region in the Gulf of Lions. Would a salinity increase distributed more evenly across the basin alter the results? To examine this, we performed an additional experiment where a linear, rather than exponential, west to east change in salinity distribution was used. The results of this experiment show no significant change to the LGM2 results. In the western basin, the increased salinity leads to even stronger deep water formation, and most of the west remains ventilated by cold WMDW. However, since the eastern basin continues to lose buoyancy at the surface the flow at Sicily remains antiestuarine.

In LGM3, paleowind stresses were used for a 60 year integration along with the same bathymetry, surface salinity, and temperature fields as LGM2. The large-scale thermohaline circulation changes little from LGM2. The eastern basin is still dominated by the formation of LDW; the main formation region remains the Rhodes Gyre, but the more zonal surface flow under the 21 kyr B.P. wind stresses leads to more salt transport into the eastern Levantine and additional deep water formation of warm and very salty water (up to 0.3 higher than LDW) off Cyprus. The Adriatic remains an estuarine basin with no water formation.

In the western basin the more zonal winds lead to a weakening of the cyclonic circulation in the Gulf of Lions. Without the important preconditioning, deep convection breaks down in the Gulf of Lions, with only cold and fresh intermediate water forming in winter. The deep western basin is filled with LDW that has overflowed the sill at the strait of Sicily. However, ***Rohling et al.***[1998] studied a core from the Gulf of Lions and concluded that deep (convective) mixing was a persistent feature in this area throughout the past 60 kyr. This suggests (1) that the paleo winds used are not representative of the actual situation, perhaps they are too zonal (because of the the coarse atmospheric GCM small changes in wind direction (such as that associated with the channeling effect of the Rhone Valley) could lead to a reestablishment of a cyclonic circulation and the probable maintanence of western deep convection) or (2) that the model is too sensitive to wind stress changes.

8. Discussion and Summary

Here we used an OGCM to simulate the circulation of the paleo-Mediterranean. In particular, we simulated the conditions in the Mediterranean Sea in the Holocene at the time of the deposition of sapropel S_1 and conditions at the peak of the last glacial maximum. It is important to note that this model, when forced toward present-day surface temperatures and salinities, reproduces very realistic surface fluxes and circulation patterns with realistic production of all major water masses. A summary of some of the key features from our simulations are given in Table 4, and our suggested

Table 4. Summary of Some of the Main Results from the Holocene and Last Glacial Maximum (LGM) Experiments

Circulation Feature	Holocene	LGM
Antiestuarine circulation	yes	yes
Sea level change significant	no	yes
Temperature changes significant	no	yes (in west)
Paleowind significant	no	maybe? (see text)
Rhodes gyre/Aegean convection	Aegean, 125 m	Rhodes gyre, LDW
Adriatic convection	AIW, 200-450 m	collapses
Western convection	~800 m	WMDW
Gibraltar outflow	AIW	WMDW

See text for more details. AIW, Adriatic Intermediate Water; LDW, Levantine Deep Water; and WMDW, Western Mediterranean Deep Water.

eastern Mediterranean paleocirculations for the two periods are given in schematic form (Figure 14).

The fluxes diagnosed from the LGM experiments show large heat losses in the Gulf of Lions and Rhodes gyre, associated with intense convective activity. Very little heat is lost from the glacial Adriatic with no formation of deep water here. There is little change from present in the basin average E-P fields, but there is a shift in the spatial pattern. Excess evaporation is almost halved in the western basin, while a substantial increase occurs in the eastern basin (mainly confined to the southern and eastern Levantine). This changed E-P pattern would seem to be consistent with some of the more recent ideas on the changes to the Mediterranean water balance at the LGM (as discussed in sections 1 and 6.). The E-P field also dominates the buoyancy forcing and drives a strong antiestuarine circulation in both the eastern and western basins.

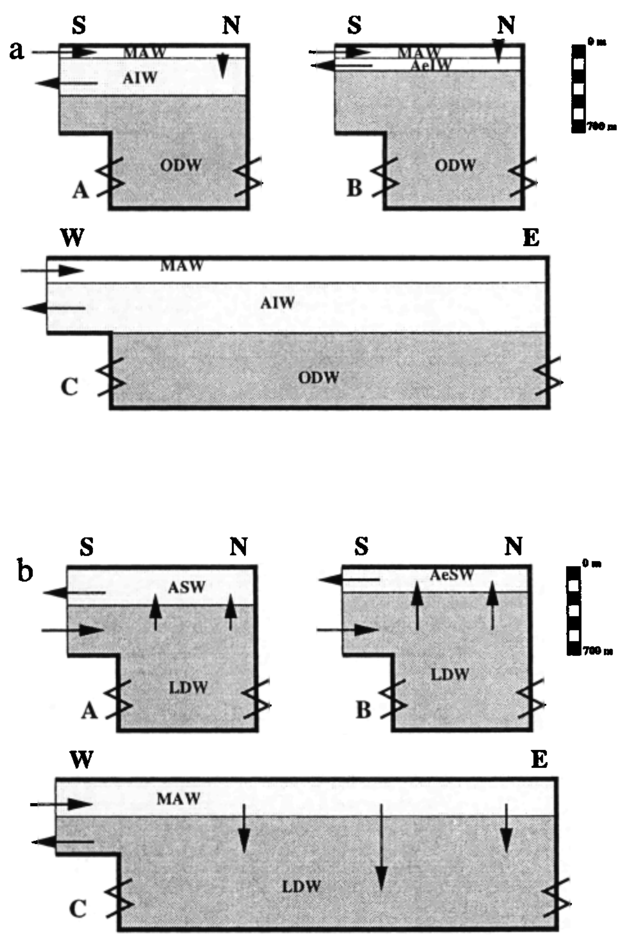

Figure 14. Schematic summarizing the results of our model experiments for the (a) Holocene and (b) LGM, with the locations of the sections shown in Figure 1. The abbreviations for the water masses are AIW, Adriatic Intermediate Water; ODW, Old Deep Water; AeIW, Aegean Intermediate Water; ASW, Adriatic Surface Water; AeSW, Aegean Surface Water; MAW, Modified Atlantic Water; and LDW, Levantine Deep Water.

In contrast, *Bigg* [1994] found in his OGCM a flow reversal at Sicily with WMDW overflowing the sill and filling the deeper parts of the eastern basin. Clearly, the forcing he derived from the *Kutzbach and Guetter* [1986] atmospheric GCM must have provided a net buoyancy gain in the eastern basin at the surface, which contradicts much of the available paleoevidence. All evidence suggests atmospheric temperatures were colder at this time which would decrease buoyancy. Although *Bigg* [1994,1995] does question the increased aridity hypothesis of some authors, he does suggest a similar net E-P to today. With this assumption (which is consistent with our model) the positive E-P leads to a strong surface buoyancy loss in the eastern Mediterranean.

Under all the proposed Holocene surface conditions we studied (*Kallel et al.*[1997], *Rohling and De Rijk* [1998], and *Thunell and Williams* [1989], with zero, positive, and negative zonal surface salinity gradients, respectively) an antiestuarine circulation is maintained. The low surface salinities combined with old saline waters below limit the depth of convection: 200-450 m in the eastern Mediterranean and 125 m in the Aegean, consistent with observations of sapropel upper depth limits [*Rohling and Gieskes*, 1989; *Perissoratis and Piper*, 1992] with unventilated deep waters. The old deep waters provide a significant reservoir of salt that is mixed upward over time, increasing the salinity of the intermediate waters, helping to reduce their buoyancy and maintain the antiestuarine exchange at Sicily, even when the west-east surface salinity gradients are reversed.

Changes to intermediate water formation lead to the near absence of LIW in the Levantine and a consequent shoaling of the pycnocline. A shallow pycnocline here, and its near absence in the Ionian, is consistent with the findings of *Castradori* [1993] and could have lead to the formation of a deep chlorophyll maximum and higher primary productivity [*Rohling and Gieskes*, 1989]. Changes to the nutricline associated with the shoaling of the intermediate/deep waters have been seen during the recent circulation changes in the eastern Mediterranean by *Klein et al.*[1997], and are therefore likely to have occurred during the more significant changes of the early Holocene. Deep water stagnation and increased productivity would both have favored the formation of a sapropel.

Despite the continuing antiestuarine circulation the export of salt to the western basin at Sicily is strongly reduced from the present day. The composition of the outflow has also changed from the salty LIW of the present day to a fresher AIW. This decrease in salt transport leads to a collapse of western convection and the formation of a deep halocline/pycnocline throughout the western basin. With no deep water ventilation occurring in either the Tyrrhenian or the western basin the resulting deep anoxia suggests the possibility of sapropel formation in these regions as well [*Thunell et al.*, 1984; *Castradori*, 1993; *Rohling*, 1994]. *Sarmiento et al.*[1988] assumed no sapropel formation in the western basin and suggested that a reverse circulation was therefore required to maintain a low phosphate transport to the west. *Rohling* [1994] commented that other mechanisms could prevent excessive phosphate concentrations in the surface layers of the western basin: (1) EMDW export (with high phosphate concentrations) was reduced at Sicily, or (2) the nutrients transported to the west remained shielded from the euphotic zone by a deep pycnocline. In the model we find a significantly increased AIW export at Sicily but also a deep shielding pycnocline in the west. Further cores in the west could help to test these hypotheses.

The strait transports under paleo forcing conditions must, of course, balance the surface fluxes but at Gibraltar the additional constraint of hydraulic control has often been used to infer circulation and transport changes, and these can be compared to our OGCM results. A drop in transport at Gibraltar of 50% is found in the Holocene from present, significantly larger than the 9% drop predicted for a 20 m sea level reduction by *Rohling and Bryden* [1994] using the hydraulic control model. However, *Rohling and Bryden* [1994] assumed that the Mediterranean's surface buoyancy forcing (in particular, the excess evaporation) was kept constant, which we do not find. Instead, the fresh water flux (Table 2) is significantly reduced throughout the basin. Using these low excess evaporations in the formula of *Bryden and Kinder* [1991], we calculate very small transports at Gibraltar, far smaller than those found in the model. There are two reasons for this discrepancy. First, the system receives an additional salinity forcing through the gain of salt from the underlying deep water. Second, the role of temperature can no longer be discounted in the buoyancy forcing as *Bryden and Kinder* [1991] for the present day. With very little excess evaporation in the Holocene the surface buoyancy forcing has switched to a thermally dominant regime. Including these terms in the buoyancy calculations gives us a larger transport close to that found in the model.

During the LGM the surface buoyancy forcing is much more similar to the present day (Table 2) with an E-P-dominated thermohaline cell. Changes in Gibraltar transport are consistent with hydraulic control estimates [*Rohling*, 1991] based on the large sea level reduction. The GCM results also produce a basin-wide shoaling of the pycnocline in the eastern Mediterranean, into the euphotic zone, as predicted by *Rohling* [1991]. The pycnocline has also shoaled in the Alboran Sea in the west.

In summary, we find that even though there are significant changes to the basin's thermohaline circulation in both periods, the basic sense of the circulation does not change from its present antiestuarine form, except in the Adriatic during the LGM. This result is consistent with changes in the surface buoyancy forcing controlling the system. For both the Mediterranean as a whole and for the eastern basin, there is a net surface buoyancy loss due to strong evaporation augmented by the surface cooling. Even with the reversed surface salinities (east fresher than west) of the *Thunell and Williams* [1989] reconstruction during the Holocene, the cooling over the eastern basin maintains a very slight net buoyancy loss through the surface. The upward transfer of salt from the stagnant deep water in the Holocene experiments considerably increases this buoyancy loss of the intermediate waters and would continue to do so for 2-3000 years. Clearly, if conditions in the Holocene did introduce a surface buoyancy gain in the eastern basin and if this was large enough to overcome the upward salinity flux from below, then estuarine circulation could have prevailed at Sicily. However, this model has reproduced very realistic fluxes when forced toward present-day surface temperature and salinity, and this strongly supports the view that antiestuarine conditions at Sicily were most likely to have prevailed during the formation of S_1.

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