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Response of the Mediterranean Sea thermohaline circulation to observed changes in the winter wind stress field in the period 1980-1993

Sarah Samuel, and Keith Haines

Department of Meteorology, University of Edinburgh, Edinburgh, Scotland, United Kingdom

Simon Josey

James Rennell Division for Ocean Circulation, Southampton Oceanography Centre, Southampton, England, United Kingdom

Paul G. Myers

Department of Meteorology, University of Edinburgh, Edinburgh, Scotland, United Kingdom

Abstract. This paper seeks to model changes in deep water production in the eastern Mediterranean induced by changes in winter wind stress. An analysis of individual monthly wind stress fields over the Mediterranean for 1980-1993 from the SOC flux data set shows that an intensification of the winter mean (mainly January) wind stress over the Aegean Sea and Levantine basin occurred in the latter half of this period. A weakening of the Mistral occurred at the same time. Two monthly wind stress climatologies were created using the 1980-1987 and 1988-1993 periods, and these were used to force an ocean general circulation model of the Mediterranean, with climatological surface T,S relaxation. The Levantine intermediate water (LIW) dispersal path in the Ionian is altered in the 1988-1993 experiment with no pathway to the Adriatic and, consequently, greatly reduced exchange at Otranto and a collapse in Adriatic deep water formation. In contrast, there is an increased exchange of LIW at the Cretan arc straits and enhanced Aegean deep water production in the 1988-1993 experiment. Much more Aegean water exits into the Levantine and Ionian basins as is shown by an east-west cross section south of Crete, along a similar path to the Meteor cruise in 1995. Changes in air-sea fluxes are diagnosed from the model showing a small increase in wintertime cooling over the Aegean and reduced cooling over the Adriatic after 1987. While the changes in air-sea fluxes are probably underrepresented by this simulation, the large changes induced by the wind forcing suggest this could be a mechanism in the altered thermohaline state of the eastern Mediterranean since 1987.

1. Introduction

In the last few years it has been recognized that the circulation and properties of waters in the Mediterranean Sea have been changing. The most notable change occurred in the eastern Mediterranean between 1987 and 1995, during which around 1/3 of the deep eastern basin was filled with new waters which formed in the Aegean Sea, replacing older waters which formed in the Adriatic basin [*Roether et al.* , 1996; *Klein et al.* , 1999]. The most conspicuous feature of these new Aegean waters is their higher salinity, which leads to the question of how the eastern basin freshwater budget may have changed. Air-sea flux changes over the eastern basin, although not necessarily confined to the Aegean, are an obvious candidate. Data on evaporation and precipitation are not, however, of a high quality, and small changes might go undetected for long periods. Another contribution which is known to have altered is the Nile River discharge, which has decreased considerably since the construction of the Aswan Dam. *Bethoux et al.* [1990], *Leaman and Schott* [1991] and *Rohling and Bryden* [1992] have all detected a long-term increase in the salinity of the waters of the western Mediterranean which might be attributed to the reduction of river discharge from both the Nile and Eurasian rivers draining through the Black Sea. However, the new deep water formation in the Aegean has occurred suddenly, and we

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need a mechanism to explain the rapidity of the observed change.

It is now becoming accepted that the increased salinity below 1500 m in the eastern basin detected in 1995 must represent the rapid conversion of a large amount of Levantine intermediate water (LIW) to form the new Aegean water [*Klein et al.* , 1999]. Such a conversion would involve a large salt flux to the deep but not necessarily a simultaneous change in the air-sea fluxes. This suggestion predicts that the new deep waters should be accompanied by a reduction in the amount of LIW in the eastern basin and evidence for this is now being found [*Roether et al.* , 1998]. The problem remains: What kind of changes in forcing conditions would have been necessary to initiate the formation of the new Aegean deep waters between 1987 and 1995?

There is some meteorological evidence that the winters of the early 1990's, were anomalously cold in the eastern Mediterranean [*Sur et al.* , 1993]. There have been measurements of decreased precipitation and increasing salinity over the Aegean [*Theocharis et al.* , 1996]. It is even possible that some of these changes might be associated with large-scale changes in North Atlantic weather patterns, as manifest in the North Atlantic Oscillation [*Hurrell*, 1995]. So far, there have been no systematic investigations of changes in meteorological conditions over the entire Mediterranean for the period in question. In this paper we will investigate the changes in wind stress over the Mediterranean from a new data set compiled from COADS data. We will then consider, using a numerical model, how these observed changes may alter the Mediterranean thermohaline circulation.

There are three ways in which altered meteorological conditions might affect the deep water properties of the eastern Mediterranean:

1. A change in evaporation and precipitation (including river runoff) would change the eastern basin freshwater balance but would almost certainly need to be sustained for several years to produce a large effect. Salt budgets [*Klein et al.* , 1999] show that while changes in excess evaporation may be important, they cannot by themselves produce the significant salt changes that have been seen in the Eastern Mediterranean deep waters.

2. A change in heat fluxes around the northern edge of the basins in winter could alter the main convection site to the Aegean. However, the enhanced heat fluxes would need to be repeated over a number of winters to produce the large-scale and lasting changes observed. Here the circulation changes might be expected to occur after water formation, particularly at the Cretan arc straits, as the new deep waters disperse.

3. Finally, a change in wind conditions would alter the upper and intermediate water circulation paths. These circulation changes occur before deep convection each winter, and would then feedback on deep water production by altering the hydrography at the water formation sites, possibly enhancing convection in the Aegean. Increased salinity at middepth would act to promote deep convection by destabilizing the water column.

An advantage of studying the changes in water formation with a modeling approach is that it is possible to consider each of the above effects in isolation to some degree. This paper focuses on changes in wind stress as a potential contributor to changes in eastern Mediterranean deep water formation. This is the simplest effect to investigate because wind changes are easy to observe, and recent work by *Myers et al.* [1998] has already demonstrated the sensitivity of the Mediterranean thermohaline circulation to wind stress fields. Other work is underway to consider changes in the surface heat flux field (P. Wu and K. Haines, Towards an understanding of deep water renewal in the eastern Mediterranean, submitted to *Journal of Physical Oceanography*, 1999).

In section 2 we show the monthly wind stress fields over the entire Mediterranean basin from 1980-1993 on the basis of an analysis of observational data compiled at Southampton Oceanography Centre [*Josey et al.* , 1999]. These indicate significant changes in average winter winds over the Mediterranean from the 1980-1987 period to the 1988-1993 period. In section 3 we describe a Mediterranean ocean circulation model, and section 4 shows results from water formation experiments performed using the wind stress fields from the earlier (1980-1987) and later (1988-1993) periods. In section 5 we look at the changes in surface heat and freshwater fluxes which have accompanied these experiments. In section 6, results are shown from a full interannual experiment running from 1980 to 1993, and conclusions and further discussion are presented in section 7.

2. Wind Stress Fields, 1980-1993

The wind stress fields presented here have been subsetted from the individual monthly data set, covering the period 1980-1993, used to produce the Southampton Oceanography Centre (SOC) air-sea heat and momentum flux climatology [*Josey et al.* , 1999]. They have been generated from ship meteorological reports contained in the Comprehensive Ocean-Atmosphere Dataset 1a (COADS1a) [*Woodruff et al.* , 1993] using the bulk formula

$$(\tau_x, \tau_y) = \rho_a C_D |\mathbf{w}| (w_x, w_y), \tag{1}$$

where τ, w are the wind stress and speed respectively and subscripts x and y denote zonal and meridional components. Here $\rho_a = 1.2$ kg m^{-3} is the surface air density, and C_D is a stability dependent drag coefficient determined using the neutral relationship suggested by *Smith* [1980] and the stability dependent profiles of *Edson et al.* [1991]. The subsetted fields consist of the northward and eastward component of wind stress on a 1° latitude by 1° longitude grid from −10°E to 40°E

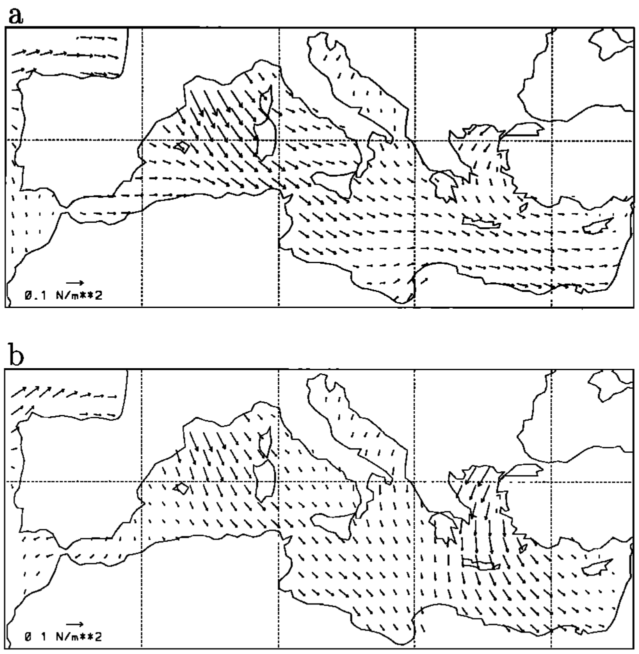

Figure 1. SOC wind stresses, averaged over (a) 1980-1987 and (b) 1988-1993.

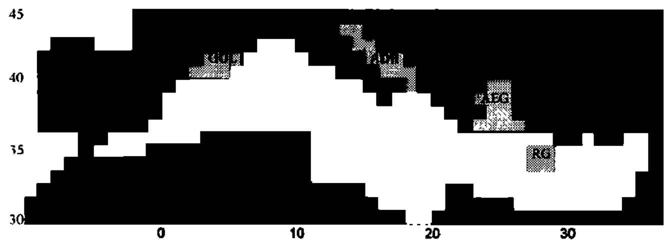

Figure 2. Plot of the regions where the wind stress amplitude is examined: GOL, Gulf Of Lions, ADR, Adriatic Sea, AEG, Aegean Sea, and RG, Rhodes Gyre.

and from 30°N to 50°N for each month from January 1980 to December 1993 inclusive. An extension of the data set to 1994-1995 was generated at a late stage in our analysis using early release COADS data. We have made use of the additional data in the observational analysis discussed below but not in the model experiments.

Analysis of the individual monthly wind stress fields shows a clear intensification of the winter mean wind stress over the Aegean Sea and Levantine Basin between the early to mid-1980s and the late 1980s to early 1990s. This change is evident in plots of the winter mean (DJF) wind stress for the two periods 1980-1987 and 1988-1993, shown in Figures 1a and 1b. These show a strengthening of the northerly wind stress values over the Aegean from 0.07 to 0.10 N m^{-2} between the two periods. In contrast, there is little change in the wind forcing of the Adriatic Sea and a slight weakening of the Mistral forcing of the Gulf of Lions. We note that the 1980-1987 winter mean field over the Eastern Mediterranean is similar to those found in the climatological analyses of *May* [1982] and *Hellerman and Rosenstein* [1983], which are based on data covering the periods 1950-1970 and 1870-1976, respectively. This suggests that it is the 1988-1993 period which is climatologically atypical. Closer examination reveals that January is the most persistently anomalous month although contributions from February 1992 and February 1993 are also significant.

The choice of winter 1987-1988 as the transition between the two periods considered is somewhat arbitrary as the time series discussed below show that the change in wind forcing took place over several winters. However, the choice has also been motivated by the hydrographic observations reported by *Roether et al.* [1996] which suggest that winter 1987-1988 is the earliest date at which the intrusion of bottom waters from the Aegean Sea, observed in the January 1995 cruise, may have started.

We have investigated the interannual variability in wind stress in four regions in which convection is known to occur: the Gulf of Lions, Adriatic Sea, Aegean Sea, and Rhodes Gyre. The climatological mean annual cycle in wind stress and direction for each of the four regions (indicated in Figure 2) is shown in Figure 3. The strongest forcing is observed over the Gulf of Lions, where the wind stress is persistently directed toward the southeast with a magnitude of the order of 0.09 N m^{-2} in winter, dropping to 0.03 N m^{-2} in summer. In contrast the wind forcing of the Adriatic is relatively weak (note the vectors in Figure 3 for this region have been scaled up by a factor of 2) and shows a seasonal variability in direction, shifting toward the northwest in midautumn. The peak wind stress in the Adriatic, 0.025 N m^{-2}, occurs in December and the mean averaged over

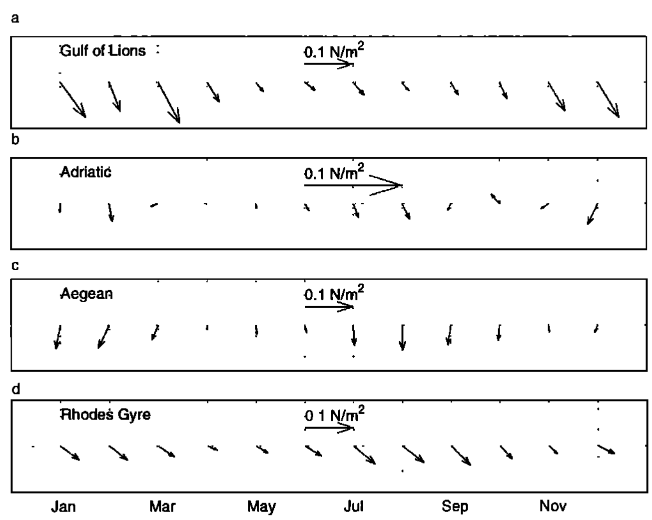

Figure 3. The climatological mean annual cycle in wind stress and direction for each of the four regions: (a) Gulf of Lions, (b) Adriatic, (c) Aegean, and (d) Rhodes Gyre.

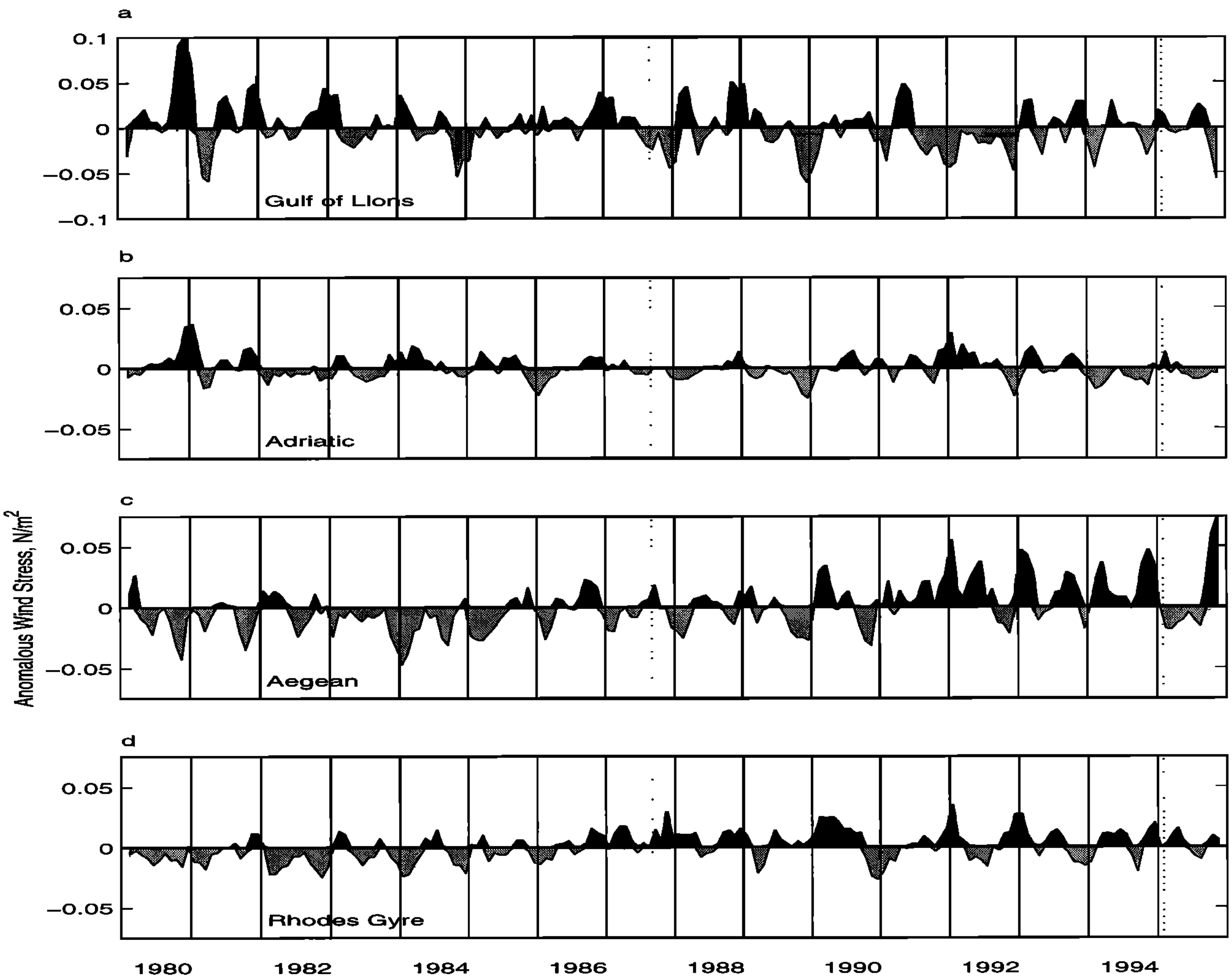

Figure 4. Time series of the monthly mean anomalous wind stress amplitude, smoothed using a three month running average, for each of the four regions: (a) Gulf of Lions, (b) Adriatic, (c) Aegean, and (d) Rhodes Gyre.

the whole year is 0.012 N m^{-2}, which compares with 0.055 N m^{-2} for the Gulf of Lions and 0.033 N m^{-2} for the Aegean Sea. In the Aegean the wind stress is generally directed toward the south and there is a noticeable semiannual cycle with the strongest stresses, of the order of 0.05 N m^{-2}, occurring in January and February and July - September. Finally, note that the forcing over the Rhodes gyre is similar in magnitude to that for the Aegean but directed toward the southeast.

Time series of the monthly mean anomalous wind stress amplitude were calculated by taking the difference between the actual wind stress and the long-term mean for each month. The time series of this, smoothed using a 3 month running average, is shown in Figure 4 (with the relevant regions indicated in Figure 2). The time series for the Aegean shows positive monthly wind stress anomalies of up to 0.05 N m^{-2} toward the end of the period considered. As one might expect, a similar trend is seen over the adjacent Rhodes Gyre region although the anomalies are somewhat smaller in magnitude. There is some indication that the Gulf of Lions has a tendency toward lower peak wind stresses in later years, while no clear trend is found for the Adriatic.

These monthly wind stress data were linearly interpolated onto a 1/4° grid to force a numerical model of the Mediterranean Sea. When the data are used for forcing, a linear interpolation between each month is also used. We did not attempt to increase the strength of the peak winds to provide the correct monthly averages, as done by *Killworth* [1996].

3. Model and Experimental Details

An ocean general circulation model (OGCM) was used to investigate the possible response of the Mediterranean Sea to changes in the wind stress patterns. The model used was the Modular Ocean Model Array (MOMA) primitive equation code. The code was developed by D. Webb and D. Stevens, and includes a free surface scheme [*Killworth et al.* , 1991]. The Mediterranean version of the code is described by *Myers et al.* [1998], and it includes additional improvements including a *Gent and McWilliams* [1990] eddy parameterization and a flux limiting advection code [Stratford, 1999]. The resolution is 1/4° with 19 vertical levels and it covers the whole Mediterranean basin including a small Atlantic box to allow free exchange at Gibraltar. This model has been used to successfully model the thermohaline circulation of the Mediterranean using climatological winds [*Myers et al.* , 1998] and also to examine water budgets in the Levantine under flux forcing (P.G. Myers and K. Haines, Seasonal and interannual variability in a model of the Mediterranean under flux forcing, submitted to *Journal of Physical Oceanography*, 1998).

In the runs to be discussed here, the surface fluxes of heat and freshwater are not imposed, but instead relaxation of surface temperature T and salinity S to an observed climatology is used. Although heat and freshwater fluxes are not specified or controlled, the implied fluxes are derived a posteriori and are shown to be quite realistic on large scales. The surface T and S we use for relaxation vary monthly and are taken from the Mediterranean Oceanographic Data Base, Version 5 (MODB5) [*Brasseur et al.* , 1996]. The salinity relaxation is much weaker than the temperature relaxation, enabling the subsurface salinity profiles, and hence the model's own circulation, to determine the amounts and density of the deep waters formed in the model each winter. This is the freedom which we use here to determine sensitivity in dense water formation to interannually varying winds. The main set of experiments performed with the model is described below.

3.1. Spin-Up Phase

The model was initialized with MODB5 data at all model levels and then spun up for 20 years using a monthly varying wind climatology based on the 8 year average SOC winds from 1980-1987 inclusive. After 20 years the circulation and air-sea fluxes show that the model has settled into a fairly consistent annual cycle with little drift in water properties formed each year.

3.2. Period Experiments

Experiment 1A continues the spin-up run for a further 20 years using the 1980-1987 wind climatology to make 40 years of total integration. Experiment 1B takes the 20 year spin-up fields and continues the integration for 20 more years using a 6 year average monthly climatology made up of winds from 1988-1993 inclusive. The final years of these two experiments are compared to determine differences in intermediate water dispersion and deep water formation.

The aim of these experiments is to assess the impact of wind changes on the thermohaline circulation of the Mediterranean. A sensitivity of the Mediterranean OGCM to winds has already been shown with different sources of climatological winds by *Myers et al.* [1998]. The experiments here are focused on whether the observed changes in the eastern Mediterranean water formation, which occurred between 1987 and 1995 [*Roether et al.* , 1996], could have been partly initiated by wind changes.

4. The Dispersal Pathways of Intermediate and Deep Waters

Water mass distributions depend critically on the dispersal pathways of the water masses after formation. Here we investigate the dispersal pathways of eastern Mediterranean water masses using salinity on isopycnal surfaces as a tracer. These diagnostics show clear differences in circulation under the two different wind stress forcings.

4.1. Intermediate Water Dispersal

Figure 5 shows salinity on the 29.00 isopycnal surface, the characteristic density of LIW formed in the model. Annual average values are shown from the final year of the integrations. From the Rhodes Gyre the LIW spreads westward into the Ionian Sea. Under the 1980-1987 forcing (Figure 5a), there are two separate pathways along which the LIW disperses in the Ionian. One branch takes the LIW into the southern Ionian, toward the Strait of Sicily and into the western basin. A second branch takes a northwestward direction along the west coast of Greece, through the Strait of Otranto and into the Adriatic Sea. These pathways are in general agreement with those described by *Malanotte-Rizzoli et al.* [1997], although our path toward Sicily is more southerly and not as direct. However, with 1988-1993 wind forcing (Figure 5b), there is only one LIW dispersal pathway apparent, directly through the central and southern Ionian Sea toward Sicily. As a result, there is reduced salt transport into the Adriatic Sea. The salinity of the LIW at Otranto decreases from ~ 38.85 under the 1980-1987 forcing to $\sim 38.70-38.75$ under the 1988-1993 forcing, close to the ~ 38.70 found by *Klein et al.* [1999] for this period. There is also a reduction in the average salinity across the eastern Mediterranean on the 29.00 isopycnal surface of between 0.01 and 0.02 practical salinity units (psu) between the earlier and later period experiments with a compensating small temperature reduction of around 0.05°C. The salinity change between runs is much larger than the salinity variance on the LIW surface within each experiment: of the order of 0.004 psu. This is consistent with the work of *Roether et al.* [1998], who suggest a significant dilution causing a cooling and freshening of LIW in the later period.

The LIW pathways in the western basin are similar between the two periods, but in the later period, there appears to be an increase in the salinity in the Tyrrhenian Sea and a reduced penetration of salinity into the Balearic Sea, west of Sardinia, on the 29.00 isopycnal. The increased salinity in the Tyrrhenian of around 0.04 psu is probably the result of accumulation of LIW, because we found no evidence of a consistent change in

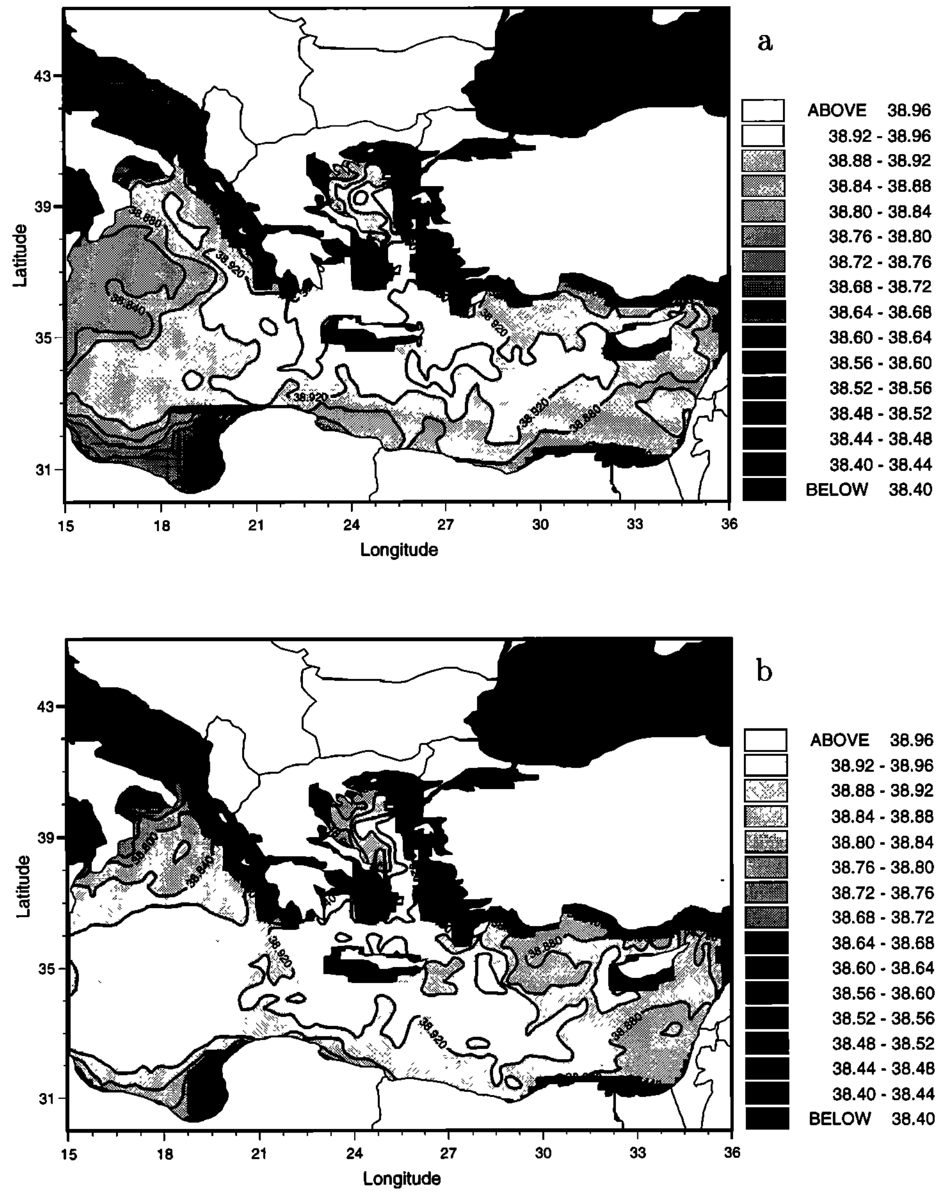

Figure 5. Annually averaged salinity along the 29.00 isopycnal for the (a) 1980-1987 period and (b) 1988-1993 period. The contour interval is 0.04 practical salinity units (psu).

freshwater exchange between the eastern and western basins at the Strait of Sicily from the final years of the two experiments.

4.2. Deep Water Formation Sites

The main impact of the altered dispersal of the LIW is on the deep water formation in the Adriatic. The water formed here is less dense during the later period run, as shown in Figure 6. The density of water at the core of the deep convection site is only 29.16 in February during the later period (Figure 6b), compared to 29.23 in the earlier period (Figure 6a). This difference is entirely due to a salinity reduction in the Adriatic of 0.1 psu from the earlier to the later period, the temperatures in the Adriatic remaining unchanged. The salinity of water in the northern Ionian is reduced by a similar amount in the later period (Figure 5b). The deep outflow from the Adriatic is therefore also much fresher (by $0.05 - 0.1$), consistent with the reduced salt transport to the deep convection site.

The changes in LIW path and the resulting change in Adriatic water formation are also reflected in the strait fluxes, particularly at Otranto. The mean volume transport at Otranto in the 1980-1987 experiment is 0.51 Sv, compared to 0.34 Sv in the 1988-1993 experiment (*Roether et al.* [1994] estimate 0.3 ± 0.1 Sv). The freshwater flux out of the Adriatic in 1980-1987 is 8×10^{-4} Sv, compared to only 2.5×10^{-4} Sv in the 1988-1993 experiment. The Adriatic outflow also does not spread nearly as far south in the Ionian basin, as illustrated

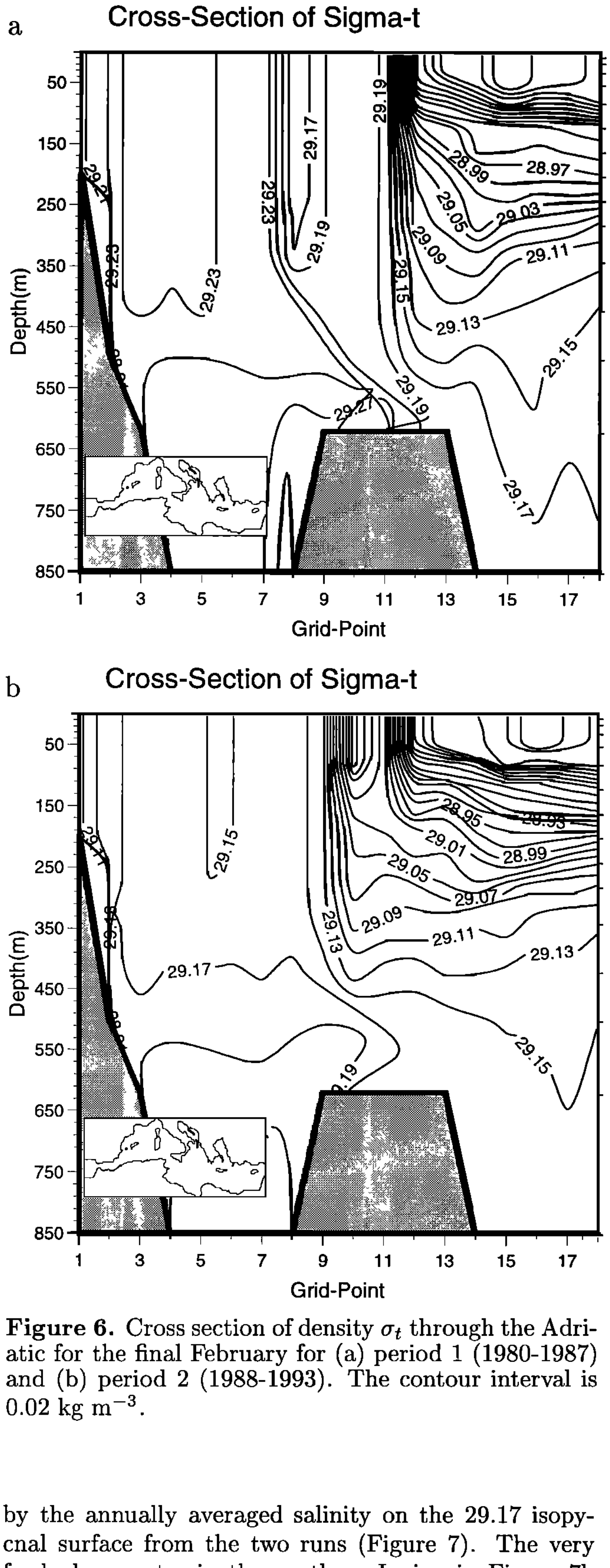

Figure 6. Cross section of density σ_t through the Adriatic for the final February for (a) period 1 (1980-1987) and (b) period 2 (1988-1993). The contour interval is 0.02 kg m^{-3}.

by the annually averaged salinity on the 29.17 isopycnal surface from the two runs (Figure 7). The very fresh deep water in the northern Ionian in Figure 7b reflects the reduced LIW transport and resultant freshening of waters emerging from the Adriatic. Observations by *Klein et al.* [1999] also suggest a reduction in the amount of salt (by about 0.1 psu, similar to the model difference mentioned above) being advected into the Adriatic in the late 1980s and early 1990s, which is consistent with these model results.

No evidence was found to suggest any change in the properties (Figure 8) or density of waters within the Aegean Sea between the two runs, unlike in the observations where there is a large increase in salinity (0.2) in the Aegean [*Klein et al.* , 1999]. However, Figure 7 shows that water from the Aegean is much more widespread in the later period. Salinities greater than 38.775 psu fill the area south of Crete and the northern Levantine. In the Ionian the 38.725 psu contour covers more of the region southwest of Crete, while a sharp front with the older Adriatic waters runs across the Ionian around latitude 35.5°N. The salinity is reduced in the northern Ionian compared to the earlier run because the Adriatic outflow is fresher. The greater extent of the Aegean waters may suggest an increase in outflow from the Aegean. However, the Adriatic outflow is also less dense in the later period, so it is possible that on the 29.17 isopycnal, the Aegean water is simply retaining its characteristic salinity over a wider area because it is not mixing as strongly with water of a similar density from a different source.

To test these suggestions, we can look at the transports through the western and eastern straits of the Aegean, to the west and east of Crete. In 1980-1987 the eastern straits have 1.73 Sv of inflow to the Aegean (mainly near the surface) and 0.71Sv of outflow from the Aegean (mainly at depth). In contrast, in 1988-1993 there is 1.82 Sv of inflow but 1.01 Sv of outflow, a considerable increase in the mainly deep outflow. At the western straits in 1980-1987, there is 0.32 Sv of inflow and 1.34 Sv of outflow, while in 1988-1993 there is 0.63 Sv of inflow and 1.44 Sv of outflow. So while the net water flux, east to west following the general cyclonic circulation, through the Aegean, has decreased in the later period, the net exchange of the Aegean with the Levantine and Ionian has considerably increased in the later period. We can also look at the net freshwater transport from the Aegean through both straits. In the 1980-1987 period, there was a freshwater outflow of 4×10^{-4} Sv, which increased to 5.3×10^{-4} Sv in the 1988-1993 period. The fact that the Aegean is a net source of freshwater is due to the strong freshening in the north due to the Bosphorous input from the Black Sea. There is a net heat transport into the Aegean which remains virtually unchanged between the 2 experiments. Thus there is a reduction in the net transport of buoyancy into the Aegean from the earlier to the later period, despite the increase in strait transports. We will look at the corresponding surface fluxes in section 5.

Roether et al. [1996] detected a shift in the formation site of eastern deep waters between 1987 and 1995 from salinity cross sections just south of Crete, across the Ionian and Levantine basins (see their Figure 1.4). We can compare our model results with these observations by plotting salinity along a similar cross section (Figure

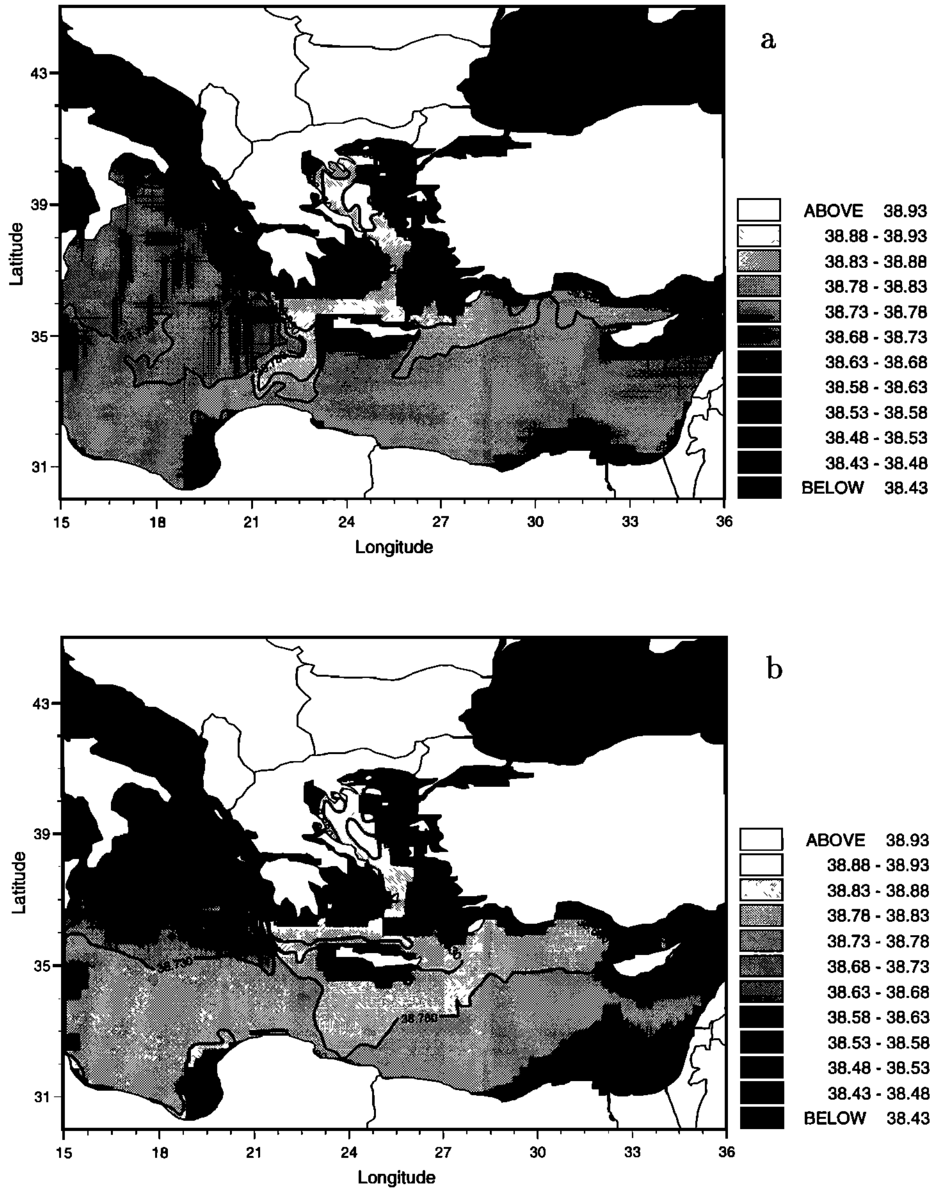

Figure 7. Annually averaged salinity along the 29.17 isopycnal for the (a) 1980-1987 period and (b) 1988-1993 period. The contour interval is 0.05 psu.

9). The position of the cross section is shown in the inset. The plot in Figure 9a is for March at the end of the 1980-1987 experiment, and, similarly, the plot in Figure 9b is for the 1988-1993 experiment. The difference between the two salinity distributions in the vicinity of Crete is striking. The 1988-1993 period shows a significant salinity maximum at depths of 750–1350 m to the east of Crete and a smaller salinity maximum to the west of Crete. Perhaps even more important, the intermediate and deep water in both the Ionian and Levantine is saltier down to 2000 m compared to 1980-1987. In 1980-1987, to the west of Crete, there is a region of saline water at shallower depths of 300–800 m, showing that there was Aegean outflow in the earlier period, but it remained at shallower depths. Of course, it must be pointed out that the changes shown here are much smaller than those found by *Roether et al.* [1996]. We will discuss some reasons for this in section 7.

To study the overall freshwater budget of the eastern Mediterranean, we can compare the total salt content of the two runs. We find that there is a negligible difference in the total eastern freshwater budget (consistent with no change in Sicily transports), but there is a redistribution of salt from the upper 500 m (surface and intermediate waters) to the deep waters (500 m+) in the 1988-1993 experiment. The amount of redistribution at the end of the 20 year run is consistent with a mean reduction in salinity of 0.009 psu for waters between 140 and 500 m and an increase of 0.005 psu below 500 m (where one standard deviation from the mean is

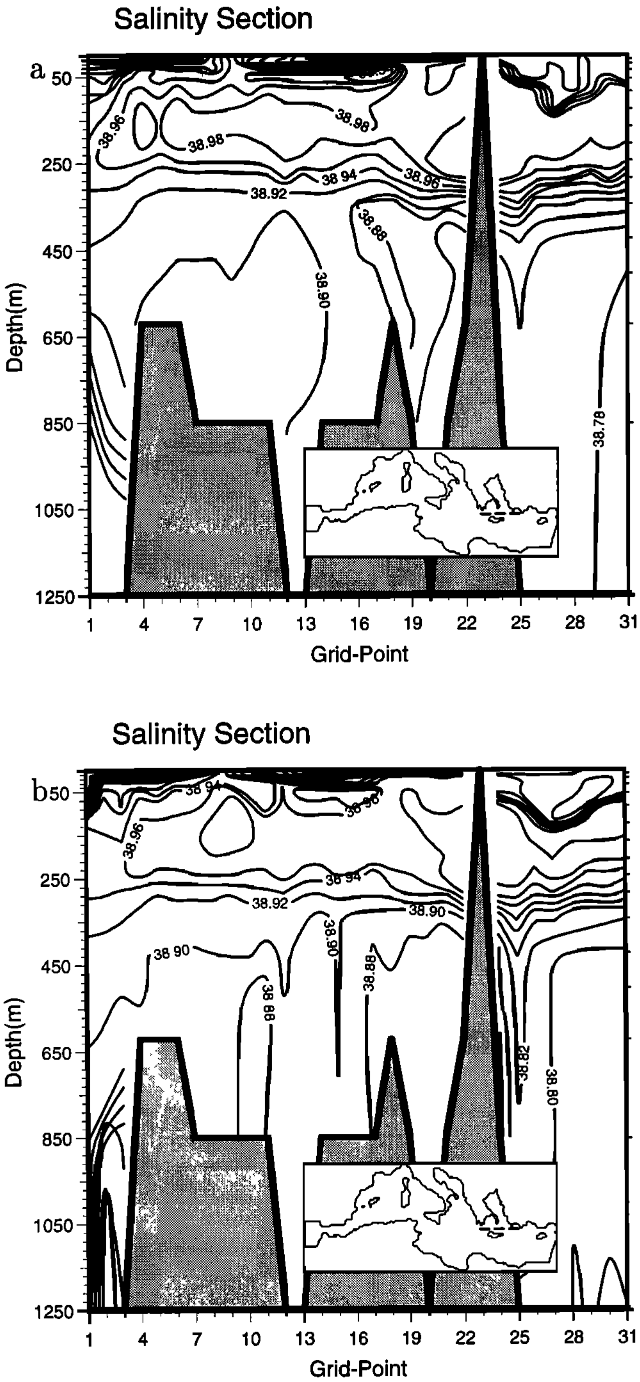

Figure 8. Annually averaged salinity along an east-west transect (path shown in inset) within the lower Aegean for the (a) 1980-1987 period and (b) 1988-1993 period. The contour interval is 0.02 psu.

0.005 and 0.003 psu, respectively), between the 1980-1987 and 1988-1993 experiments. These changes are in the same direction but much smaller than the observations of *Klein et al.* [1999] who find an average salinity change of 0.14 psu below 1200 m. This redistribution process takes place in the Aegean, as will be shown in section 5.

5. Surface Heat and Freshwater Fluxes

The surface heat and freshwater fluxes, which were necessary for the maintenance of the surface water properties, can be calculated a posteriori. Although these experiments are designed to show changes in thermohaline flows particularly induced by wind stress changes, we do not have direct control of fluxes, because of the relaxation forcing, so their impact on the final results must be carefully assessed. Table 1 shows these fluxes, averaged over the last 10 years, for the two experiments, for each of five regions. The basin average values for both the heat flux and evaporation - precipitation are generally consistent with observed climatological values of these properties, as reported in previous work [*Myers et al.* , 1998].

There are only minor changes in the surface heat loss, except in the Adriatic Sea, where a decrease in surface cooling occurred between 1980-1987 and 1988-1993, suggesting that less deep water is being formed and that there is a decrease in the density of the deep water. With less inflow of LIW (Figure 5), there is a decrease in surface fresh water gain over the Adriatic in the later period, consistent with the behavior of the restoring boundary conditions when there is less salt in the surface layer.

The Aegean Sea shows virtually no change in annual surface heat loss but an increase in surface fresh water gain of around 30% in the 1988-1993 period. This is consistent with the increased transport of freshwater out of the Aegean through the strait, described in section 4. However, this result is puzzling because it does not explain the increased Aegean deep water formation which has clearly taken place in the later period. To understand what is happening we need to look at Table 2 , which shows the breakdown of fluxes over the Aegean into winter (December, January and February (DJF)), and the other 9 months of the year, for the two experiments (all fluxes are averaged over the last 10 years). We find an increase in heat loss from the Aegean in winter equivalent to about 11.4 W m^{-2} in the 1988-1993 period. It is this heat loss which is responsible for the conversion of more LIW into Aegean deep waters which then flow out into the Ionian and Levantine between 500 and 2000 m depth. The extra heat loss in winter is, however, compensated by extra heat gain during the rest of the year, so that the signal of the increase in Aegean deep water formation does not appear in the annual averages. The winter E-P seasonal differences are much smaller and do not contribute significantly to the new water formation in the model.

6. Interannual Experiment

A final experiment was carried out using the full interannual winds for the 14 years from 1980-1993. In this experiment the model fields from the end of the spin-up period were used as initial conditions, and the

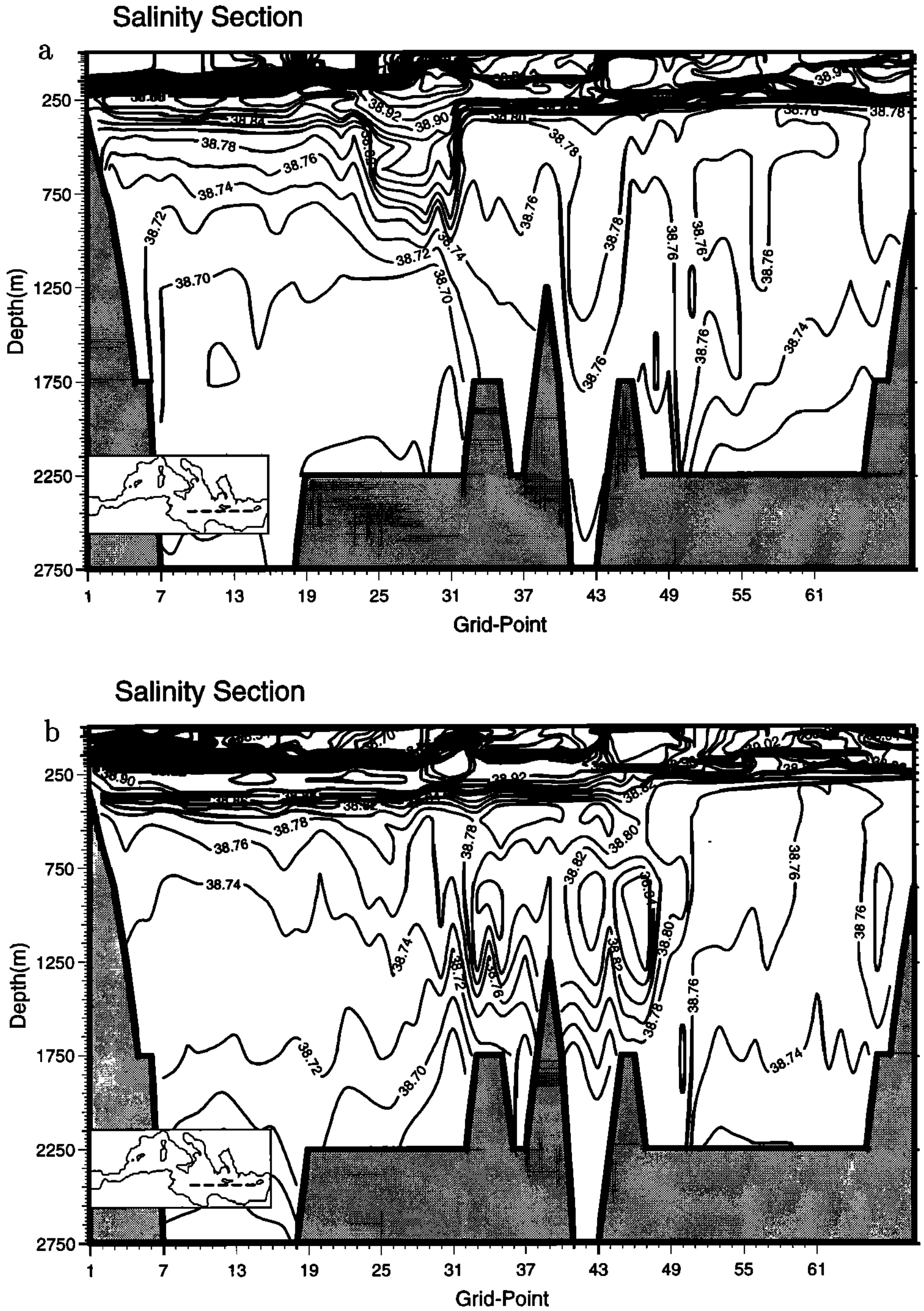

Figure 9. Cross section of salinity across the Levantine and Ionian for the final March for (a) period 1 (1980-1987) and (b) period 2 (1988-1993). The contour interval is 0.02 psu.

model was run for 14 years with full interannual winds (linearly interpolated monthly) from 1980-1993 inclusive. It is less easy to draw strong conclusions about the change between 1987 and 1988 from this experiment because there is considerable interannual variability, particularly in the intermediate water pathways. However, several of the integral measures of water production, including strait and surface fluxes, are consistent with the period experiments.

In this experiment, after 1987, salinity on the 29.00 isopycnal is strongly reduced across the whole of the eastern basin (Figure 10), consistent with the period experiments and the calculations of *Roether et al.* [1998]. The LIW dispersal pathways are less clear from isopycnal diagnostics (only 2 years shown, Figure 10), with the southern Ionian pathway showing up distinctly in only a few years. The northern pathway to the Adriatic exists most years, although not always as a coherent current.

Table 1. Surface Heat and Freshwater Fluxes for Our Experiments, Averaged Over the Last 10 Years of Each Run.

Experiment	Heat Flux W m^{-2}					E-P cm yr^{-1}				
	MED	WMED	EMED	ADR	AEG	MED	WMED	EMED	ADR	AEG
1980-1987	-6.5	-9.7	-2.1	-36.8	-9.1	78	96	69	-21	-14
1988-1993	-6.3	-10.2	-1.9	-30.9	-9.0	76	100	64	-5	-18

Regional abbreviations: MED, basin average; WMED, the western Mediterranean; EMED, the eastern Mediterranean (including the Adriatic and Aegean); ADR, the Adriatic; and AEG, the Aegean. E-P, evaporation - precipitation.

However, salt transport into the Adriatic does decrease from the early years to the later years (by about a factor of 3, from 0.011 to 0.004 Sv), including some years where the net freshwater transport changes sign. This is associated with the formation of a strong front outside of the Adriatic, related to the change in wind derection in the late 1980s and early 1990s. This leads to a decrease in the salinity in the Adriatic of 0.07 – 0.08 psu after 1987 (close to that shown in Figure 6) (Table 3).

As a result of this, there is a clear decrease in excess precipitation over the Adriatic basin between 1987 and 1988 (Table 4) after 1987. This is a key signal of reduced Adriatic deep convection. This leads to a decrease in mean outflow at Otranto (Table 3) with the outflow also decreasing in salinity, although there is significant interannual variability in outflow volume because the strait is not hydraulically controlled and is strongly influenced by winds.

Over the Aegean, there is consistently a greater heat loss in winter (DJF) in the years 1988-1993 than in 1980-1987, as can also be seen in Table 4. The average value for 1988-1993 is approximately -99 W m^{-2} while for 1980-1987 it is only approximately -90 W m^{-2}. As with the period experiments this is the signature of increased convection and deep water production in winter. We might expect to see changes in the Aegean outflow associated with increased convective winter cooling after 1987, but as with the Otranto strait transports, the winds produce strong variability, which masks any consistent transport signal at the Cretan arc straits. Plots of salinity on the 29.17 isopycnal (not shown) are also less informative than for the period experiments because of the increased interannual variability which masks the mean water pathways.

Table 2. Surface Heat Flux for the Final 10 Years of Our Experiments, over the Aegean, for Winter and for the Rest of the Year.

Experiment	Heat Flux, W m^{-2}	
	DJF	Rest of Year
1980-1987	-92.0	18.4
1988-1993	-103.4	22.4

DJF, December, January, and February.

7. Discussion and Conclusions

Roether et al. [1996] found evidence for a significant change in the deep water properties in the eastern Mediterranean that occurred between cruises in 1987 and 1995. This was caused by a shift of deep water formation to the Aegean, whose waters have filled about 1/3 of the deep eastern basin. An open question is the source of the additional salinity to the Aegean to allow for the formation of these new dense saltier deep waters. Suggestions for the extra salt source include external changes through the freshwater budget and/or internal changes through the redistribution of salt. *Roether et al.* [1996] suggest that to get the observed Aegean salt increase, an increase in E-P of 20 cm yr^{-1} maintained over the period 1987-1995 would be needed over the entire eastern basin, and the meteorological evidence does not support such a large change.

However, the meteorological data does show that a significant change in the wind stresses in winter occurred over the Mediterranean, between pre-1987 and post-1988. Results in section 2, particularly, show there was a significant strengthening of the winter winds over the Aegean Sea after 1987. Changes in wind stress have been previously shown to have significant effects on intermediate water pathways [*Myers et al.* , 1998]. These pathways are crucial for transporting LIW to the deep convection sites around the Mediterranean, so we have examined the impact of the observed wind changes on a model of the thermohaline circulation.

Two monthly wind climatologies were created, one with the winds from 1980 to 1987 and the second from 1988 to 1993, from the observed SOC wind data set. These two different wind stress fields were used to force an ocean general circulation model of the Mediterranean, each for a period of 20 years. In the 1980-1987 experiment, LIW is dispersed along two paths in the Ionian, toward the Adriatic and westward toward Sicily. Significant amounts of salt reach the Adriatic, encouraging strong deep convection in winter and the formation of dense waters. Little water formation occurs in the Aegean with only small amounts of Cretan Interme-

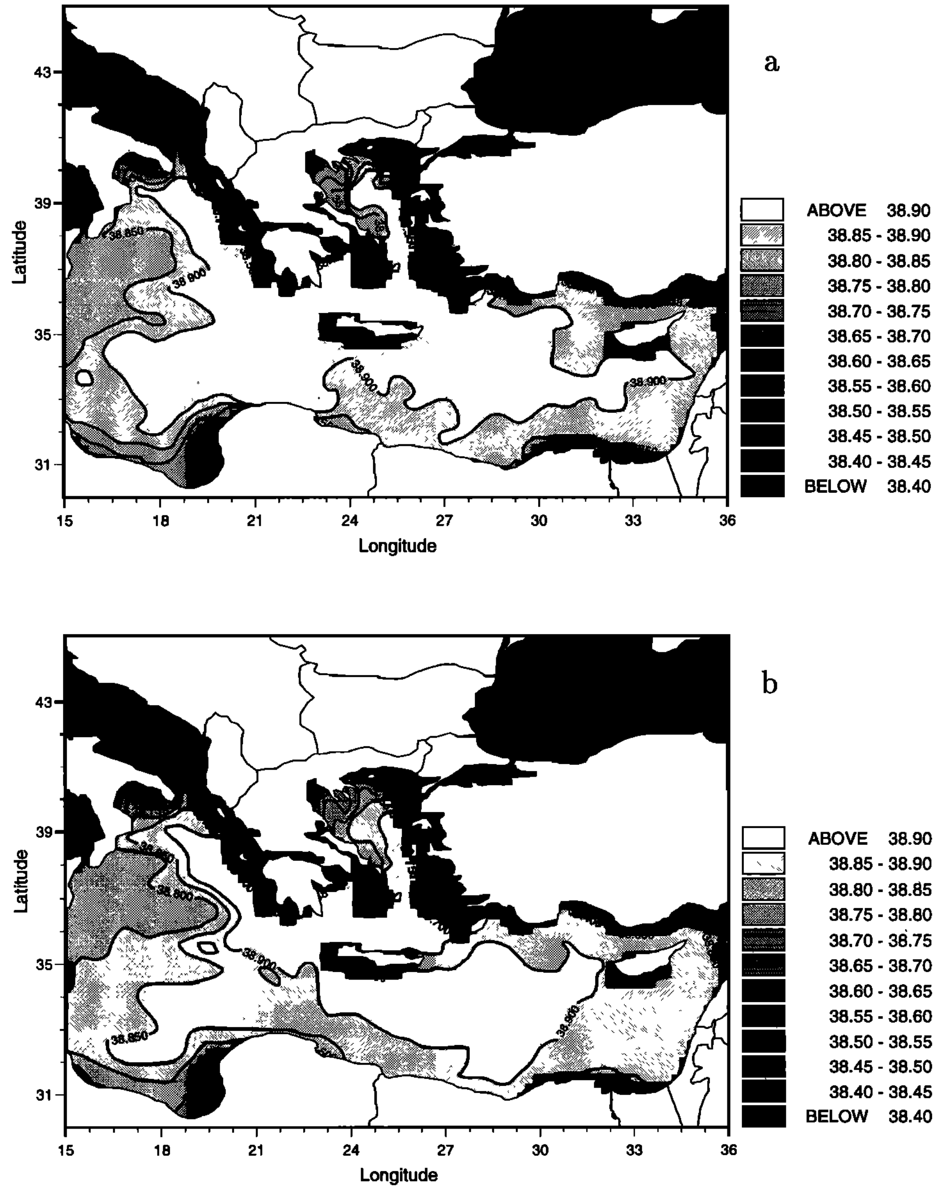

Figure 10. Annually averaged salinity along the 29.00 isopycnal for a) year 3 of the integration (July 1982 — June 1983) and b) year 11 of the integration (July 1990 — June 1991). The contour interval is 0.05 psu.

diate Water (CIW) emerging from the western straits at shallow depths.

In the 1988-1993 period run a major change in LIW path occurs. The westward LIW path across the Ionian remains, leading to few changes at Sicily and in the overall freshwater budget of the eastern basin. However, the LIW pathway to the Adriatic is greatly reduced, and, consequently, less salt is transported into the northern Ionian. As a result, Adriatic deep water production is also greatly curtailed, and the densest water emerging from the Otranto straits is much fresher and less dense.

In contrast, there is an increase in the export of dense Aegean waters in the 1988-1993 period. The signature of more Aegean water production is indicated by increased cooling over the Aegean in the winter (DJF) months of $\sim$ 11.4 W m^{-2} and increased transport at the Cretan arc straits. A salinity section to the south of Crete (Figure 9) also shows the export of considerably more Aegean water which sinks to greater depths in the Ionian and Levantine basins. A comparison of the eastern basin salt content shows a mean reduction in salinity of 0.009 psu for waters between 140 and 500 m and an increase of 0.005 psu below 500 m, between the 1980-1987 and 1988-1993 experiments. A number of features in the observed changes during the post-1987 period are not seen in the model. The transfer of salt to deeper levels is less than in observations, and the significant increase in salinity in the Aegean is not produced. Hence the salinity of the outflowing waters is not sufficiently increased, and thus they occupy a middepth

Table 3. Annual Mass Transport Through the Strait of Otranto in Sverdrups, Salinity of the Outflowing Water, and the Average Salinity Within the Adriatic During the Interannual Experiment.

Period	Otranto Mass Transport	Salinity of Otranto Outflow	Adriatic Salinity
July 1980 - June 1981	0.40	38.63	38.56
July 1981 - June 1982	0.50	38.61	38.56
July 1982 - June 1983	0.44	38.59	38.55
July 1983 - June 1984	0.52	38.61	38.55
July 1984 - June 1985	0.48	38.59	38.55
July 1985 - June 1986	0.45	38.63	38.55
July 1986 - June 1987	0.42	38.63	38.55
July 1987 - June 1988	0.40	38.59	38.53
July 1988 - June 1989	0.32	38.59	38.53
July 1989 - June 1990	0.39	38.55	38.52
July 1990 - June 1991	0.42	38.53	38.51
July 1991 - June 1992	0.38	38.54	38.49
July 1992 - June 1993	0.29	38.52	38.48

Note the downward trend in all quantities over time.

position in the water column rather than sinking to the bottom.

An interannual experiment using monthly winds from all the 14 years from 1980 to 1993 shows that broadly similar changes in water production occurred around 1987-1988. Air-sea fluxes over the Adriatic and Aegean show sudden changes at this time (Table 4) although the diagnostics of water distributions are more ambiguous because of large interannual variations in water distributions caused by the varying winds.

This experiment has focused on changes in deep water production over the Mediterranean initiated by observed changes in winter wind stress. The mechanism being emphasized is the altered pathways of intermediate water which in turn alter the hydrography at the sites of winter time cooling and hence alter the deep water formation sites and water properties formed. The observed changes in winds seen after 1987 do occur in early winter (mainly January), which is consistent with modifying hydrography prior to water formation. However, we do not control surface fluxes in the model, and small but significant changes in surface fluxes do accompany the changes in deep water production in the model. There is no reason why the mechanism of altered hydrography should not alter the deep water formation sites even if fixed surface fluxes were imposed. However, such an experiment, with altered wind but the same fluxes, would be even more idealized than that performed. The altered winds seen in Figure 1 would undoubtedly lead to altered air-sea fluxes over the Mediterranean, probably of a considerably greater magnitude than those seen in our model.

The change in eastern basin deep water production site from the Adriatic to the Aegean after 1987 is clearly initiated in our model by the mechanism of modified hydrography, which we describe and is fully consistent with observations from the Mediterranean Sea. Clearly, important changes in air-sea fluxes are also part of the mechanism which would need to be treated to fully reproduce the results observed in the 1995 Meteor cruise

Table 4. Adriatic Annual Freshwater Flux and Aegean Winter Heat Flux During the Interannual Experiment.

Period	Adriatic E-P, cm yr^{-1}	Aegean DJF Heat Flux, W m^{-2}
July 1980 - June 1981	-25	-100
July 1981 - June 1982	-30	-95
July 1982 - June 1983	-26	-86
July 1983 - June 1984	-26	-92
July 1984 - June 1985	-26	-87
July 1985 - June 1986	-30	-88
July 1986 - June 1987	-29	-84
July 1987 - June 1988	-16	-90
July 1988 - June 1989	-18	-95
July 1989 - June 1990	-17	-103
July 1990 - June 1991	-17	-98
July 1991 - June 1992	-17	-103
July 1992 - June 1993	-18	-106

Note the significant changes that occur between 1987 and 1988.

[*Klein et al.* , 1999]. However, our results do show that the observed changes in winter winds from 1980 to 1993 produced changes in both intermediate water pathways and deep water production sites which are consistent with observations, and we therefore propose this mechanism as a component of the changes which lead to the altered thermohaline state of the eastern Mediterranean after 1987.

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K. Haines, P. G. Myers, and S. Samuel, Department of Meteorology, University of Edinburgh, JCMB, Kings Buildings, Mayfield Road, Edinburgh, EH9 3JZ, Scotland, U.K. (email: kh@met.ed.ac.uk; paulm@met.ed.ac.uk; ssamuel@plume.mit.edu)

S. Josey, James Rennell Division for Ocean Circulation, Southampton Oceanography Centre, Empress Dock, Southampton, SO14 3ZH, England, U.K. (email: Simon.A.Josey@soc.soton.ac.uk)

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