Toward an Understanding of Deep-Water Renewal in the Eastern Mediterranean

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ABSTRACT

This paper presents a scenario for understanding the unexpectedly large changes of deep waters and thermohaline circulation in the Eastern Mediterranean during the past decade. It is demonstrated as a possible deep-water renewal mechanism in the Eastern Mediterranean in a numerical simulation with a high-resolution model, which has successfully reproduced the observed 1987 and 1995 regimes as shown by the two *Meteor* cruises. A budget study of the model simulation has shown that more than 75% of the salt added to the deep layers of the Eastern Mediterranean could have come from the top 1000 m by a salinity redistribution process triggered by intensive cooling over the Aegean Sea. A water transformation process analysis is carried out in the model simulation to reveal how colder and fresher dense deep water formed in the Aegean is turned into the Eastern Mediterranean Deep Water (EMDW) as observed. Surface heat and freshwater fluxes are diagnosed to show the roles of each component during the transition.

About a 10% to 15% (11% or 6.6 cm yr^{-1} in this experiment) increase of freshwater loss over the Eastern Mediterranean plus some cold winters, which are able to decrease the sea surface temperature in the Aegean by an additional 1°–2°C, would be capable of switching the major EMDW formation site from the Adriatic to the Aegean Sea and altering the EMDW structure from the pre-1987 state to the 1995 regime as revealed by the two *Meteor* cruises, M5 and M31. It does not necessarily require a large increase of $E - P$ in the Aegean itself to produce the salty bottom water observed in the Cretan Sea (the south Aegean). An increased transport of Levantine Intermediate Water (LIW) into the Aegean would increase the salinity of the Aegean bottom water, but the major salinity increase for the new EMDW occurs in the Cretan Sea where the colder but fresher Aegean bottom water meets the LIW and convects to the bottom. Such internal convection causes the temperature to drop but the salinity to rise in the deep layers of the Cretan Sea. When the Cretan deep water flows through Kasos Strait outside of the Cretan Arc, it falls again to the bottom of the Levantine Basin and spreads into the Ionian Sea. During this second course of internal convection the water is diluted further before it sinks to the bottom due to mixing with old EMDW.

1. Introduction

The recent *Meteor* cruise (M31/1, January–February 1995) brought back rather surprising observational results to the Mediterranean science community (Roether et al. 1996). Malanotte-Rizzoli et al. (1996) published an article in *Eos* with the following description: ''The major result that emerged from the preliminary analysis was completely unexpected: the Eastern Mediterranean deep waters were in an entirely new configuration.'' Detailed analysis has now been published by Klein et al. (1999). The core fact is a dramatic increase of salinity in the deep layers of the Eastern Mediterranean, which cannot be easily explained by any reasonable increase of $E - P$ (evaporation minus precipitation) at the surface. It is clear that this increase of bottom salinity, which occurred since the previous *Meteor* cruise M5 in 1987, is due to a switch of source for the Eastern Mediterranean deep water (EMDW) from the Adriatic Sea to the Aegean (Klein et al. 1999). The impacts of this deep water transition on the thermohaline circulation and the Mediterranean ecosystem are already widespread, as shown by the *Meteor* 31/1 cruise data and other more recent measurements. The distributions of all water properties have been affected across the Eastern Mediterranean. There have also been reports of deep water changes in the Tyrrhenian Sea and other impacts on the Western Mediterranean are likely. It is important to understand what caused such unexpectedly large changes and to be able to predict the longer term impacts on the entire thermohaline circulation of the Mediterranean Sea.

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Roether et al. (1996) and Klein et al. (1999) have speculated on two possible processes that may have caused these changes. A large increase in the net surface evaporation would cause a rapid increase of salinity in the Aegean, but this would need to be equivalent to a net evaporation increase of more than 0.2 m yr^{-1} over the entire Eastern Mediterranean for seven consecutive years to form the new deep waters. This emphasizes the magnitude of the salinity increase between 1987 and 1995, detected in the deep waters. An alternative mechanism is that a change of ocean circulation causes a salinity redistribution within the Eastern Mediterranean between the intermediate and the deep waters. In reality, a combination of these is likely. This paper presents a scenario, supported with numerical simulations, to demonstrate how the new waters may have formed due to anomalous cooling in the northern Aegean Sea, with only a moderate increase of $E - P$ over the Eastern Mediterranean.

In a climatological sense, there are four basic water masses in the Mediterranean: the modified Atlantic water (MAW) occupies the surface layers above 200 m, the Levantine Intermediate Water (LIW) lies between 200 and 500 m, with the Eastern Mediterranean Deep Water and Western Mediterranean Deep Water (WMDW) filling the deep basins. The transformation of the warm and fresh Atlantic water entering the Mediterranean at Gibraltar, into the cold and salty Gibraltar outflow (mainly LIW) in the Eastern Mediterranean, has been understood since Wüst (1961), although it is now thought that the Mediterranean outflow actually contains some WMDW (Stommel et al. 1973; Kinder and Parrilla 1987).

The formation and dispersal of WMDW and EMDW can be regarded as secondary components of the Mediterranean thermohaline circulation, both involving the entrainment and cooling of the LIW. It is known that WMDW is formed in the Gulf of Lions by entrainment of LIW into the winter mixed layer, enhanced by the mistral winds causing surface heat loss within the Gulf of Lions gyre (Schott et al. 1994; Wu and Haines 1996). Some controversy surrounded the source of EMDW concerning the relative roles of the Aegean and Adriatic Seas [see Klein et al. (1999) for a more detailed discussion] until a survey of transient tracers by *Meteor* cruise M5/6 during the 1987 POEM (Physical Oceanography of the Eastern Mediterranean) program. The transient tracer data from this cruise clearly showed that the Adriatic has been the primary source of EMDW for a considerable period (Schlitzer et al. 1991; Roether and Schlitzer 1991). There were signs of Aegean water outside the Cretan Arc but it was limited to depths between 500 and 1200 m and the water below was shown to be of Adriatic origin up to 100 years old. It is known that LIW plays an important role encouraging deep convection inside the Adriatic increasing the salinity of the EMDW, as demonstrated in modeling studies such as Wu and Haines (1996). A recent 120-yr simulation by Wu and Haines (1998) has successfully reproduced a pre-1987 circulation climatology of the Mediterranean, consistent with the Mediterranean Oceanic Data Base (MODB) (Brasseur 1995) climatology and measured basin-scale heat and water budgets (e.g., Bethoux and Gentili 1994).

However the post-1987 Eastern Mediterranean thermohaline circulation is characterized by a switch (it is not clear which year it started) of deep water source from the Adriatic to the Aegean, which consequently altered the EMDW from a colder (13.3°C and fresher (38.66 psu) (Schlitzer et al. 1991) to a warmer (13.8°C) and saltier (38.8 psu) water mass (Roether et al. 1996; Klein et al. 1999). The potential density (σ_θ) of the EMDW also increased from below 29.18 to above 29.2, which indicates an increase of surface buoyancy loss either by net heat or freshwater loss, or both, to the atmosphere. Because of the higher salinity, it is natural to speculate on an increased net loss of freshwater, that is, increased evaporation or reduced precipitation, leading to an increase of surface salinity. However, as noted by Roether et al. (1996), the required net increase of $E - P$ is rather too large. The production of saline water in the Aegean from an average increase of $E - P$ of 0.2 m yr^{-1} for 7 yr is equivalent to only 9% of the total salt increase below 1500 m in the Eastern Mediterranean. Theocharis et al. (1996) have reported that from 1986 to 1995 there was an increase of depth-averaged potential density ($\Delta\sigma_\theta \sim 0.2$) in the Cretan Sea (or the south Aegean Sea) induced by an increase of salinity ($\Delta S \sim 0.15$ psu) and *a considerable decrease of temperature* ($\Delta T \sim -0.4$°C). There have been very cold winters over the Eastern Mediterranean during this period, one example being the 1987 extreme cold surge over the Greek peninsula reported by Lagouvardos et al. (1998). During a 10-day period, extremely cold air with very strong winds blew southeastward across Greece and the Aegean Sea. Surface temperatures at Limnos Island in the northern Aegean dropped to below −4°C and in Crete to below 10°C. In a separate period deep convection was reported in the northern Levantine Sea during the winter of 1992 (Sur et al. 1992) while the mean winter air temperature for 1992 over the south Aegean was quoted at 10.5°C (Theocharis et al. 1996). Figure 1 shows a time series of surface heat flux over the Aegean between 1979 and 1994, calculated from the ECMWF reanalysis (Gibson et al. 1997). It clearly indicates repeated cold winters over the Aegean from 1990 to 1994 with the average heat loss for 1990–94 (8.4 W m^{-2}) about three times that of 1979–89 (2.9 W m^{-2}).

Consider a scenario in which several repeated severely cold winters over the Eastern Mediterranean between 1987 and 1995 cause the sea surface temperature in the north and central Aegean to drop by 1°–2°C below the climatological average. The extra cooling would cause denser waters to form. This newly formed dense water would not initially have the high salinity observed in the deep Cretan Sea by Theocharis et al. (1996) or

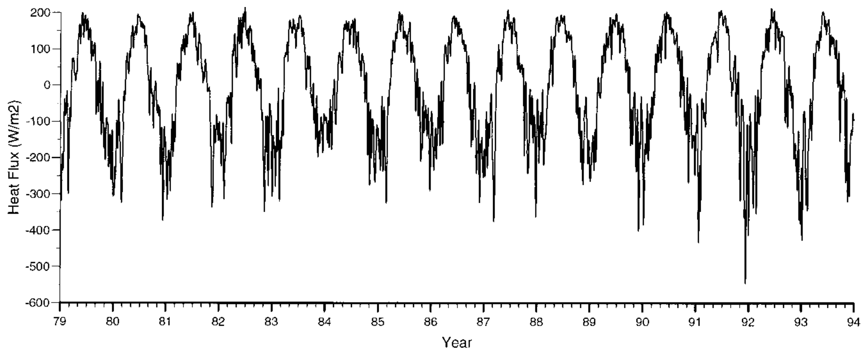

FIG. 1. Oceanic surface heat flux over the Aegean Sea (35.5°–41°N, 23°–27.5°E) during 1979–94 obtained from the ECMWF reanalysis. The flux was calculated daily but the plot was smoothed with a 7-day running average to remove very short time fluctuations.

outside the Cretan Arc in the main Eastern Mediterranean Sea by Roether et al. (1996), but it would be much colder and denser. If it were sufficiently dense, this water would rapidly make its way toward the south Aegean along the bottom. The sill leading to the Cretan Sea is less than 250 m deep, and this north and central Aegean outflow would inevitiably meet the warm and saline LIW, which would be entrained as it sinks to the bottom of the Cretan Sea and flows farther out of the Cretan Arc. Internal convection and mixing would reduce the temperature anomaly but considerably increase the salinity anomaly of the new bottom waters and leave a weak stratification between intermediate and deep waters in the Cretan sea, as reported by Theocharis et al. (1996). Furthermore, the extra bottom outflow from the north Aegean would be compensated by an upper-layer inflow that would be made up of more saline LIW. Such an exchange of water, in and out of the Aegean, would increase the salinity, that would in turn encourage even denser water to form, leading to a positive feedback. In this way there does not need to be a substantial increase of surface $E - P$ to produce high salinity waters. Extra salt would be transferred into the bottom layers in the Cretan Sea through enhanced entrainment during internal convection and the additional LIW drawn into the Aegean would encourage the salinity redistribution probably at the expense of reduced transport of LIW into the Adriatic and even through the Sicilian Channel. This scenario is proposed as a possible mechanism responsible for the observed changes of deep water formation in the Eastern Mediterranean, and a model simulation of it forms the basis of this paper.

Section 2 describes the experiments and presents results for comparison with observations from the two *Meteor* cruises reported by Schlitzer et al. (1991), Roether and Schlitzer (1991), Roether et al. (1996), and Klein et al. (1999). A process analysis is presented in section 3, which looks in detail at how the altered water masses are formed in the model. In section 4, we discuss the relevance of this simulation to the actual events between 1987 and 1995 in the Eastern Mediterranean. Section 5 concludes the paper.

2. A simulated deep water renewal

a. The model

The numerical model employed is a Mediterranean version of the Geophysical Fluid Dynamics Laboratory modular ocean model (MOM). The current version is based on the model used by Wu and Haines (1998) and the reader is referred there for a general description of the equations. This version of the model has a horizontal resolution of $0.125° \times 0.125°$, with a vertical resolution of 41 levels. The vertical spacing of model levels adopts method 2 from MOM2 (see Pacanowski 1995), with a smooth vertical stretching as described in Treguier et al. (1996). The vertical grid points have also been rearranged such that a w point is in the center of the vertical velocity box but tracer points are off-centered. This ensures that $\partial T/\partial z$ is calculated at the depth where w is evaluated. In this way the vertical advection of tracers has a higher accuracy and avoids possible negative diffusivity caused by the numerics. The subgrid mixing is biharmonic in the horizontal and diffusive in the vertical with horizontal momentum coefficient $A_h = 3 \times 10^{17}$ cm^4 s^{-1}, horizontal tracer coefficient $K_h = 2 \times 10^{17}$ cm^4 s^{-1}, and vertical momentum coefficient $A_v = 1.5$ cm^2 s^{-1}, and vertical tracer coefficient $K_v = 0.3$ cm^2 s^{-1}.

The model has spherical coordinates with a bathymetry shown in Fig. 2. The model domain is closed with rigid lateral boundaries. Outside Gibraltar there is a small box of Atlantic water allowing water exchanges, and within this Atlantic box temperature and salinity are relaxed on a 1-day timescale at all levels towards climatology. The model is forced with monthly varying wind stresses from a 9-yr average of the National Meteorological Center (now known as the National Centers for Environmental Protection) wind field as derived by Roussenov et al. (1995), and a Haney (1971) relaxation on surface T and S toward a modified monthly dataset based on the MED5 version of MODB data (Brasseur et al. 1995). The forcing repeats each year. A hard relaxation of surface temperature with a 2-h timescale is applied to the top layer (10 m) to ensure winter con-

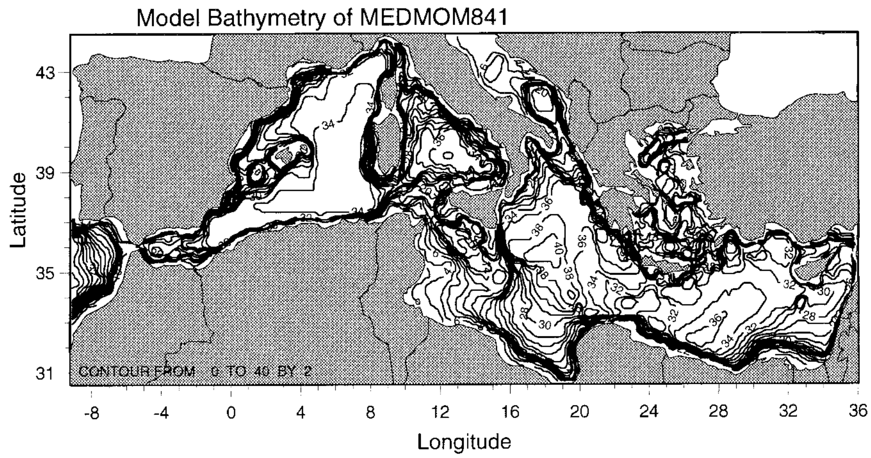

FIG. 2. Model bathymetry of the Mediterranean Sea at a horizontal resolution of 0.125° × 0.125° and 41 vertical levels from the surface to 4000 m. Contours are labeled by model levels with an interval of 2.

vection. For salinity, however, a much weaker relaxation (5-day timescale) is used everywhere except in the Levantine, thus permitting deep convection to be influenced by the in situ hydrography, that is, the subsurface salinity distribution which is controlled by circulation.

An earlier version of this model has been run at a coarse resolution (0.25° × 0.25°, 19 levels) for various studies of the Mediterranean (Wu and Haines 1996, 1998). The higher resolution version used here is necessary because we are interested in details of the bottom circulation and its intertaction with the bathymetry around the Aegean which is very complicated. A further addition to the previous model version is the addition of a passive tracer. This has no effect on the circulation but is carried and diffused in an identical manner to T and S. This passive tracer is used to simulate the distribution of CFC12 using code to define the solubility and surface concentrations over time provided to us by Roether (1998, personal communication). We will not discuss the tracer results in detail here because further work is required, but they are compared to observations and they effectively indicate the presence or absence of new waters at the bottom of the Eastern Mediterranean produced from different sources.

b. The experiments

Experiments are designed to reproduce a deep-water renewal process similar to the 1987–95 event in the Eastern Mediterranean based on the scenario described in the introduction. The question to be answered is: Could extra cooling in the Aegean have triggered, and be largely responsible for, the observed new water masses? First the model is tested for its ability to reproduce the pre-1987 deep-water formation process, which is a repeated run of Wu and Haines (1998) but at higher resolution. The model is then integrated further with extra cooling applied in February, mainly in the northern Aegean.

1) EXPERIMENT A: CLIMATOLOGICAL FORCING

Experiment A has the same surface forcings as Wu and Haines (1998). The model was initialized with the final state of the 100-yr simulation by Wu and Haines (1998) which was linearly interpolated to the higher resolution grid and then run for 20 more years to permit adjustments to the higher resolution. After this 20-yr period the CFC12 tracer was initialized to zero everywhere and surface concentrations were defined for year 1970. The model was then run for 17 more years with surface CFC12 evolving, to compare with 1987 *Meteor* cruise data.

2) EXPERIMENT B: EXTRA COOLING IN THE AEGEAN

Experiment B is designed to simulate the impact of cold winters in the Aegean, with the minimum sea surface temperature forced to fall by 1°–2°C. This change is introduced as a smooth northward reduction of SST in the Aegean Sea, toward which the model is relaxed. This change is only introduced once per year during the coldest month, February. Figure 3a presents the climatological February SST over the Aegean and surrounding areas from MODB MED5. Figure 3b shows the new February SST field. Apart from this extra cooling, all the other model parameters remain the same. It should be emphasized that the surface salinity relaxation remains unchanged and the results, therefore, concentrate on the effect of cold winters. Any changes to $E - P$ are a consequence rather than a triggering factor in the event and in any case are shown to play a more minor role in the transformation. The extra cooling is

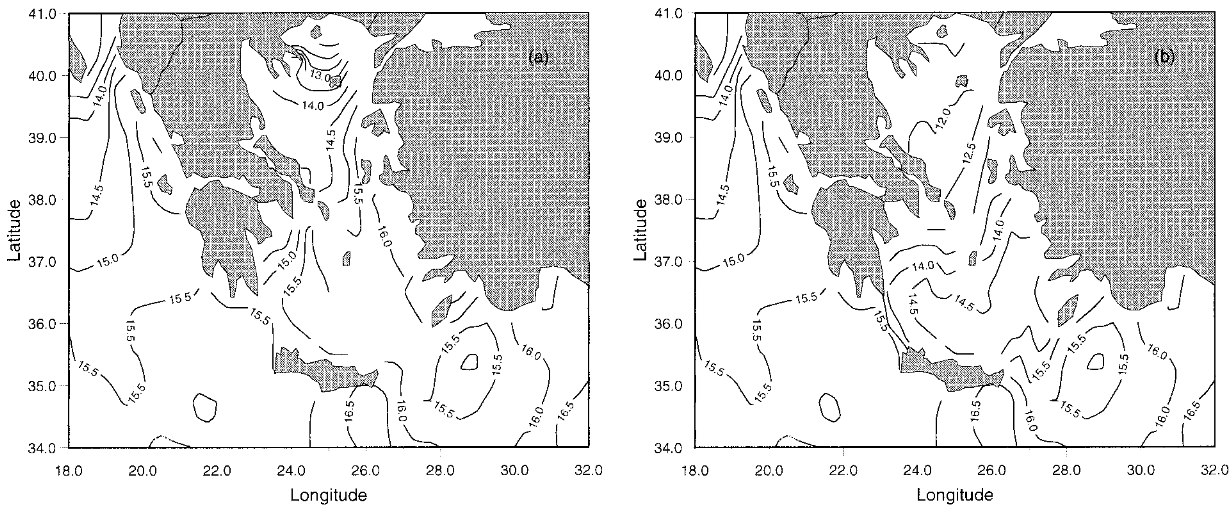

FIG. 3. Sea surface temperature comparison between experiments A and B for the coldest month, February. (a) is taken from MODB MED5 and (b) is a modification of (a) by a smooth reduction of temperature in the Aegean. This is the only difference between the two experiments in surface forcing.

applied from year 18 onward for a further 8 years to simulate *Meteor* cruise M31 conditions in 1995.

c. The results

Figure 4 shows vertical cross sections of temperature, salinity and CFC12 distributions in the Eastern Mediterranean for two model snapshots on 15 December, year 17 (1987 conditions), and 15 December, year 25 (1995 conditions). First of all, one can see dramatic differences between the left and the right columns in Fig. 4 showing the two distinctive regimes. From year 17 to year 25, a large amount of warmer and saltier water with high CFC12 concentrations has replaced the old EMDW. The relatively colder and fresher old EMDW, with low CFC12, has been lifted up, forming an identifiable layer between 600 and 1500 m that is especially visible from the CFC12 in the Levantine. The LIW layer west of Crete has thinned with a much reduced core salinity. In year 17 this water has salinity above 38.9 psu in the central Ionian, while 8 years later it has reduced to just above 38.8 psu. This resembles the changes from 1987 to 1995 reported by the two *Meteor* cruises M5 and M31.

The left column of Fig. 4 shows a classical three-layer water mass structure consisting of the MAW, the LIW, and the EMDW, which is most clearly seen from the salinity. The range of water properties, in the EMDW can be described in a *T–S* scatterplot as shown in Fig. 5a, which includes only grid points below 1400 m. Figure 5a shows the homogeneity of the EMDW in the model after 37 years of high-resolution integration. The potential density, σ_θ, lies in a very narrow range between 29.16 and 29.18, with temperature between 13.2° and 13.7°C and salinity between 38.65 and 38.75 psu, close to the 1987 measurements (see Klein et al. 1999). The upper right of the scatterplot represents older waters in the far east with higher temperature and salinity. Toward the bottom left are the younger waters with less homogeneity, slightly higher density but lower temperature and salinity.

We can see that the source of the EMDW at this stage is located in the Adriatic. Some of the water formed in the Aegean does reach the bottom of the Cretan Sea but its effect outside is limited to depths above 1500 m. This can be seen from the vertical cross sections of *T*, *S*, and most clearly from CFC12. It is clear that all the CFC12 at this time is in the western Ionian, mostly against the western wall, broadly consistent with the distribution described by Schlitzer et al. (1991). The bottom water properties in year 17 are shown in Fig. 6, where *T*, *S*, σ_θ, and CFC12 values are plotted whenever the bottom is deeper than 1500 m. Figure 6 should be compared with Figs. 15a,b,e,f of Klein et al. (1999). The simulated bottom-water properties over most of the domain are very close to M5 measurements. In the far east the water is slightly warmer and saltier in the model because in this region the water has not been replaced from that produced in the $\frac{1}{4}°$ spinup. The CFC12 distribution confirms this and there is still no CFC12 found in the deep Levantine Basin. The $\frac{1}{8}°$ EMDW has improved properties over the $\frac{1}{4}°$ simulation results.

The right column from Fig. 4 shows the cross sections on 15 December, model year 25, for comparison with the 1995 observations. After 8 years with the extra Aegean winter cooling we see a large volume of new water with high CFC12 concentrations in the deep basins of the Eastern Mediterranean to the south and east of Crete. This new water is characterized by higher temperatures and salinities compared to the previous waters in this region. Comparing with Fig. 3 of Klein et al. (1999), we conclude that the model has reproduced many fea-

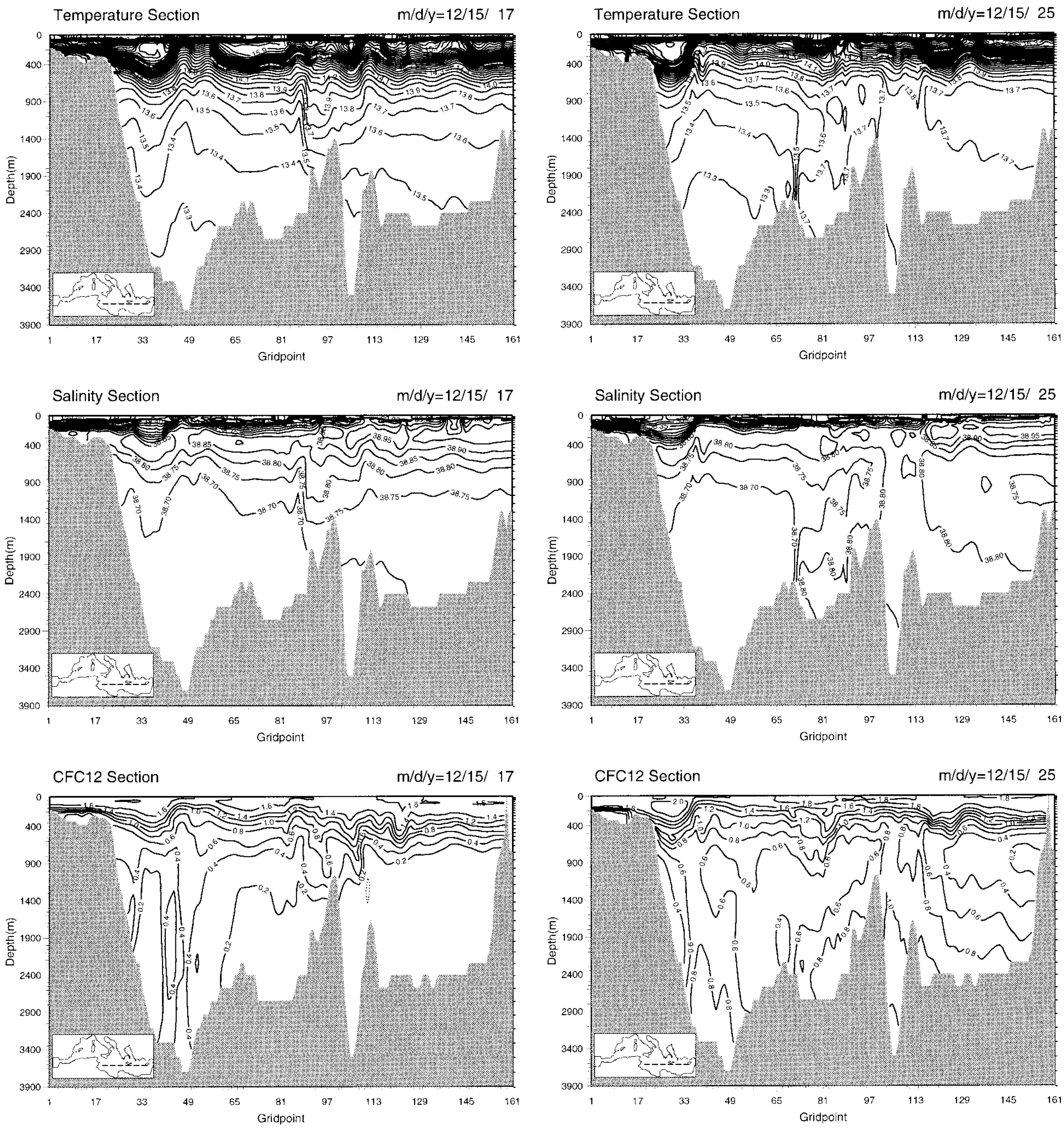

FIG. 4. Cross sections of temperature (top panel), salinity (middle), and CFC12 (bottom) from model snapshots on 15 Dec. The left column (year 17) should be compared with the 1987 state as revealed by the *Meteor* M5 cruise, and the right column (year 25) should be compared with the 1995 state as revealed by the *Meteor* M31 cruise.

tures of the new situation. The new properties of the Eastern Mediterranean deep waters below 1400 m are shown in scatterplot Fig. 5b. A lot of new water has been added in the upper right of the scatterplot with higher density, temperature, and salinity than anything seen in Fig. 5a. The greater scatter here also indicates the younger age of the water and the evolving formation conditions. The newly added water has a potential density around 29.2 or higher, a temperature between 13.6° and 13.8°C, and salinity between 38.7 and 38.86 psu. The very densest waters with σ_θ greater than 29.3 are waters from the bottom of the Cretan Sea before they exit into the main basin of the Eastern Mediterranean.

Corresponding to Fig. 6, the distributions of bottom water properties for the new regime are plotted in Fig. 7. Large contrasts of all water properties between the

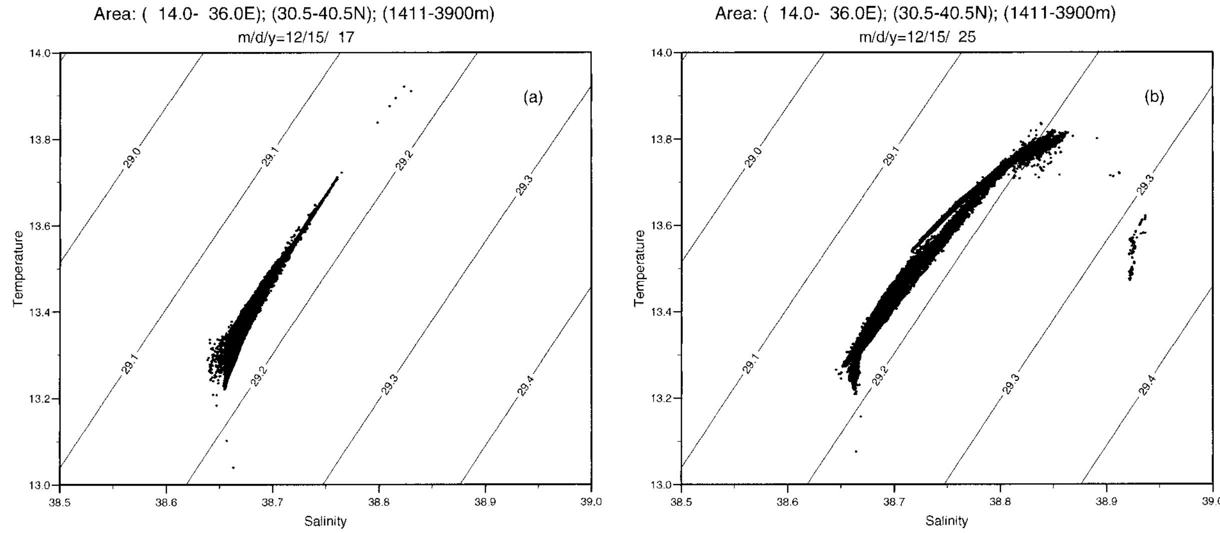

FIG. 5. Scatterplots of *T–S* for the eastern Mediterranean deep water below 1400 m for (a) 15 Dec, year 17 and (b) 15 Dec, year 25.

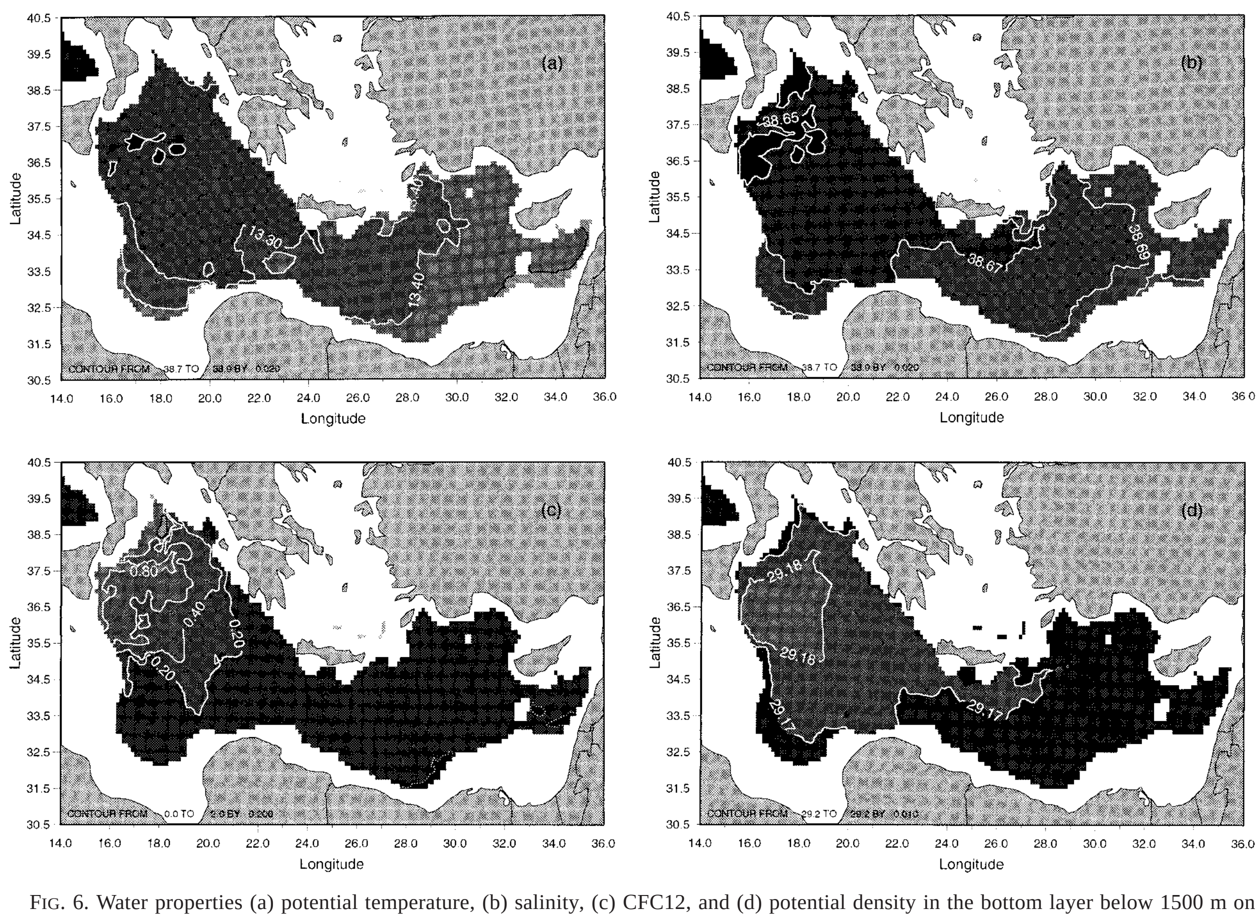

FIG. 6. Water properties (a) potential temperature, (b) salinity, (c) CFC12, and (d) potential density in the bottom layer below 1500 m on 15 Dec, model year 17. They should be compared with the 1987 *Meteor* M5 cruise measurements shown in Klein et al. (1999) their Figs. 15a,b,e,f.

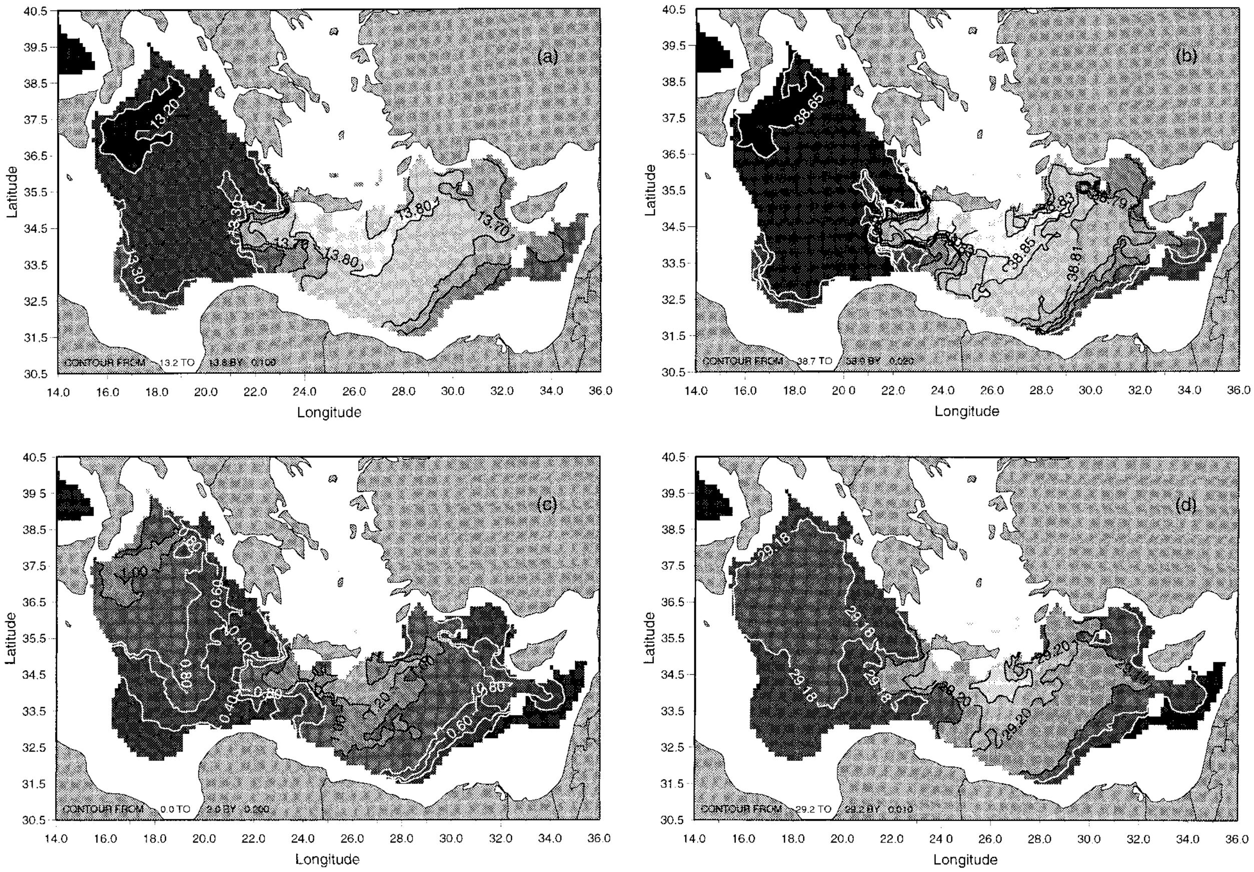

FIG. 7. Water properties (a) potential temperature, (b) salinity, (e) CFC12, and (d) potential density in the bottom layer below 1500 m on 15 Dec, model year 25. It should be compared with the 1995 *Meteor* M5 cruise measurements shown in Klein et al. (1999), their Figs. 14a,b,e,f.

Levantine Basin and the Ionian Sea are seen in Fig. 7, indicating two different deep-water sources. The newly added water in the Levantine is denser, warmer, and saltier, with higher CFC12 concentrations compared to the water in most of the Ionian. It is not difficult to trace the origin of the new water back to the exit of the Kasos Strait. Just outside the straits, southeast of Crete, one sees the core of the new water with a maximum temperature of 13.8°C, salinity of 38.86 psu, and σ_θ of 29.21. These values are very consistent with observations in Figs. 14a,b,e,f of Klein et al. (1999). The spreading of the new water westward into the Ionian Sea across the Cretan sill is also evident in Fig. 7. The modeled bottom CFC12 values are higher than observations, which might be because the atmospheric data for CFC12 was only available up to 1991 corresponding to model year 20. After year 20, the model was repeatedly forced with the 1991 data, which may be higher than in later years since atmospheric concentrations have been falling (Roether et al. 1996; Klein et al. 1999). Insufficient mixing may also contribute to the higher modeled CFC12 values.

3. A process analysis

In the previous section, we have shown simulations of two instantaneous states of water mass structure for the Eastern Mediterranean. Compared with the two *Meteor* cruise datasets presented by Klein et al. (1999), our model simulations resemble the observations in considerable detail. Here we show how this change comes about in the model.

Figure 8 shows T and S at the bottom of the Aegean in the winter to summer of year 18, the first year of new Aegean water formation, indicating the path of the newly formed Aegean deep water from the north of the Aegean Sea into the Cretan Sea. The first cold year is chosen because the contrast with the older waters enables circulation paths to be more clearly seen. During the coldest time, February (top plots), surface convection reaches the bottom of the northern Aegean but not the Cretan Sea (see Fig. 9). The densest water formed has a temperature about 12°C and salinity about 38.7 psu. Farther south, the bottom water is warmer (12.5°C) but saltier (38.8–38.9 psu). In April (middle plots), the new water has begun to flow down into the north Cretan

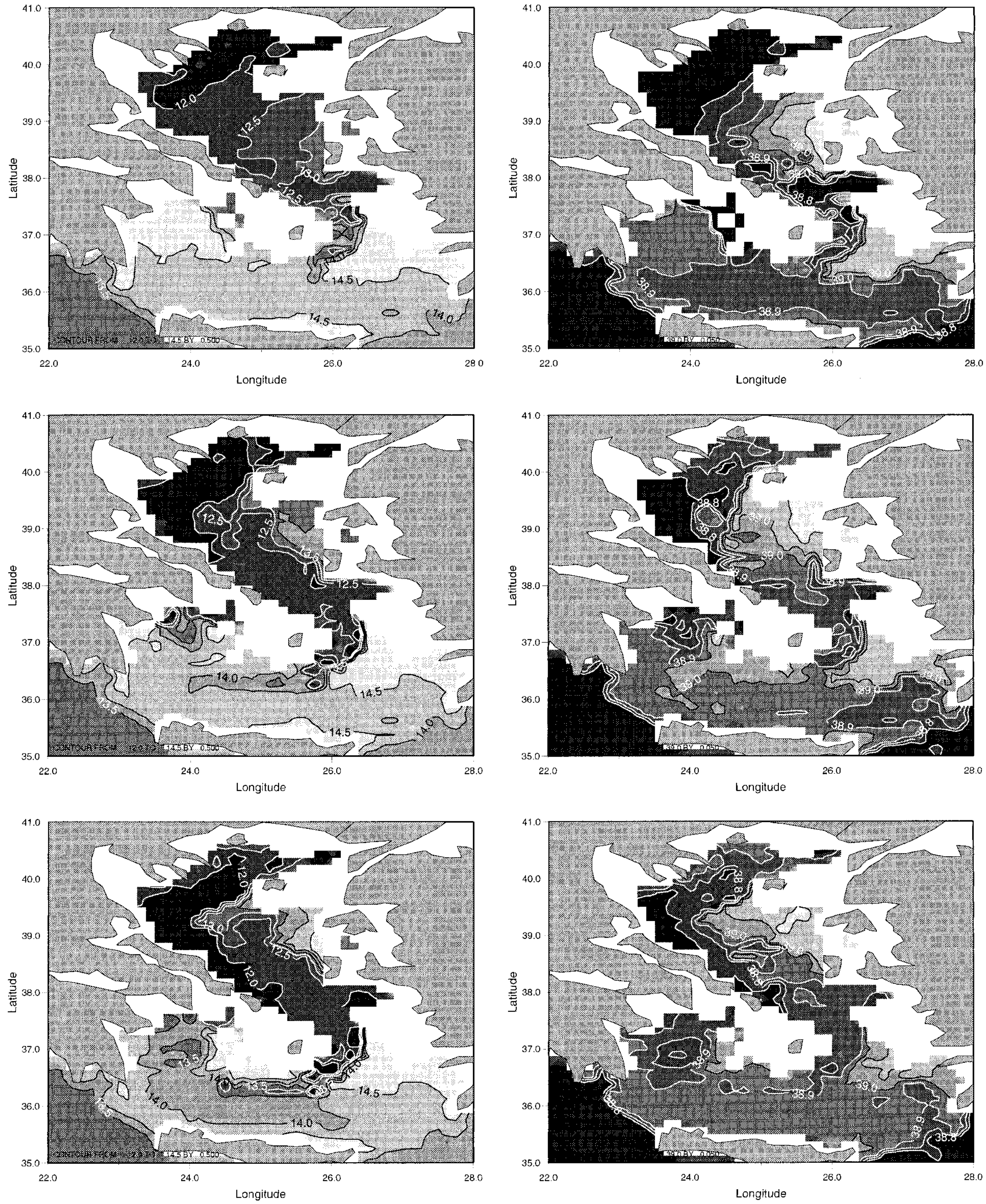

FIG. 8. Water properties (left, potential temperature; right, salinity) at the bottom level (deeper than 150 m) for Feb, Apr, and Jun of model year 18, the first winter of new water formation, showing the outflow of the Aegean water from the north to the Cretan Sea along the east coast of Greece. Notice the increase of temperature and salinity on the way out due to mixing with LIW above.

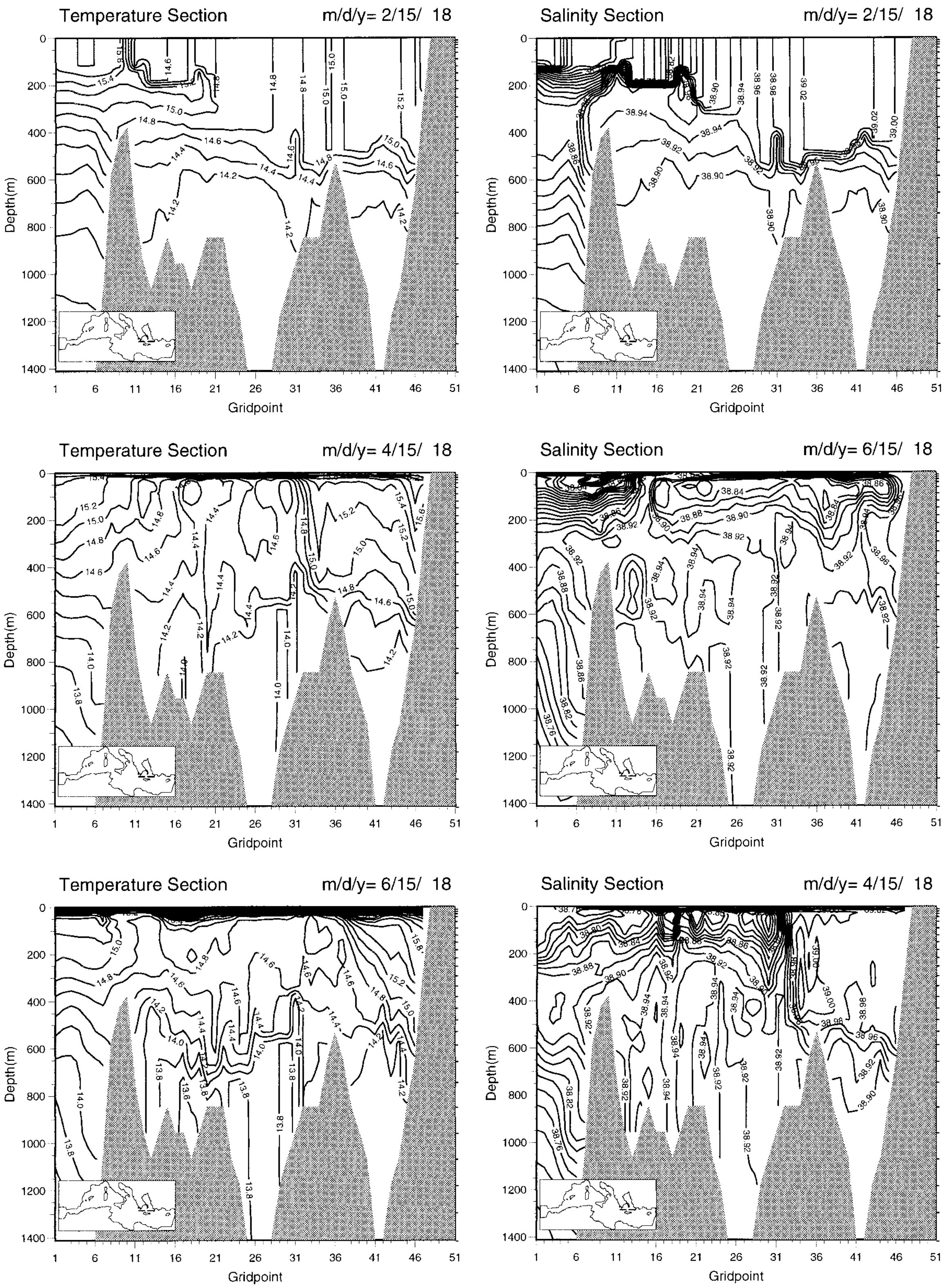

FIG. 9. East–west vertical cross sections of temperature (left) and salinity (right) in the Cretan Sea for Feb, Apr, and Jun of model year 18, the first winter of new water formation, showing the outflow of the Aegean water and its vertical mixing. Notice that convective mixing occurs below the bottom of the winter mixed layer and the decrease of temperature and increase of salinity later in spring in the deep layers of the Cretan Sea.

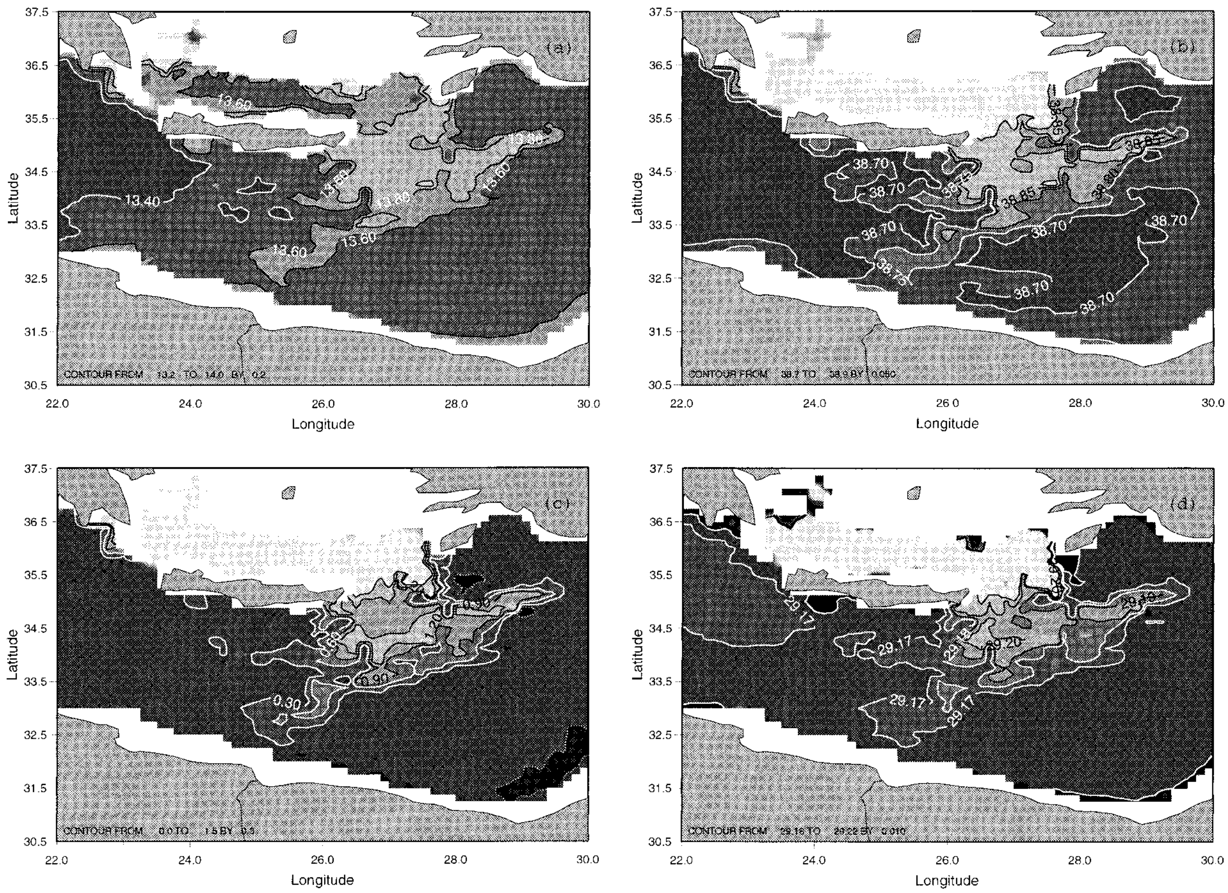

FIG. 10. Water properties: (a) potential temperature, (b) salinity, (c) CFC12, and (d) potential density, at the bottom level (deeper than 500 m) for Dec, model year 19, 22 months after the first new waters are formed, showing the outflow of the Aegean water from Kasos Strait into the Levantine Basin and changes of properties due to convective mixing.

Sea, as indicated by the coherent cold tongue in the bottom temperature. The salinity signature does not show up as well as temperature, but there is an increase of salinity in the bottom layer of the Cretan Sea, which becomes clearer in June (bottom plots). By June, the new water occupies most of the Cretan Sea and, comparing February with June, we see in the bottom layer of the Cretan Sea a decrease of temperature from above 14° to below 14°C (contour interval 0.5°C and an increase of salinity from below 38.9 to above 38.9 psu (contour interval 0.05 psu). The new bottom water in the Cretan Sea is not formed locally by surface convection, but where is the extra salinity coming from?

Figure 9 shows west–east vertical cross sections of temperature and salinity through the middle of the Cretan Sea, corresponding to the same times as Fig. 8. In February, when convection from the surface occurs, the mixed layer is at most 500 m deep, with properties typical of LIW: temperature about 15°C and salinity around 39 psu. The main thermocline and halocline remain intact and convection does not reach the bottom. In April convection in the Cretan Sea is active again, but this time it is not from the surface but starts from the bottom of the main thermocline and convects from there to the bottom. This internal convection begins as the new dense Aegean water arrives from the shelf farther north. By April, we see a decrease of temperature in the bottom layer by 0.2°C and a salinity increase by 0.02 psu. By June, there is a further decrease in temperature to 13.6°–13.8°C and the main halocline has been almost erased as more new Aegean water arrives. By October (not shown), the Aegean deep outflow gradually stops and the main thermocline and halocline begin to reestablish as more LIW is advected into the region. The bottom water of the Cretan Sea ends up with a temperature around 13.7°C and a salinity around 38.91 psu.

After filling the bottom of the Cretan Sea, the new water begins to spill out from the Kasos Strait into the Levantine Basin. Figure 10 shows water properties at the bottom level outside the Cretan Arc, including temperature, salinity, CFC12, and potential density. It is clear that no water is flowing out west of Crete and all the new water exits through the southeast passage between Crete and Rhodes. It is not clear from observations what the relative contributions of the eastern and

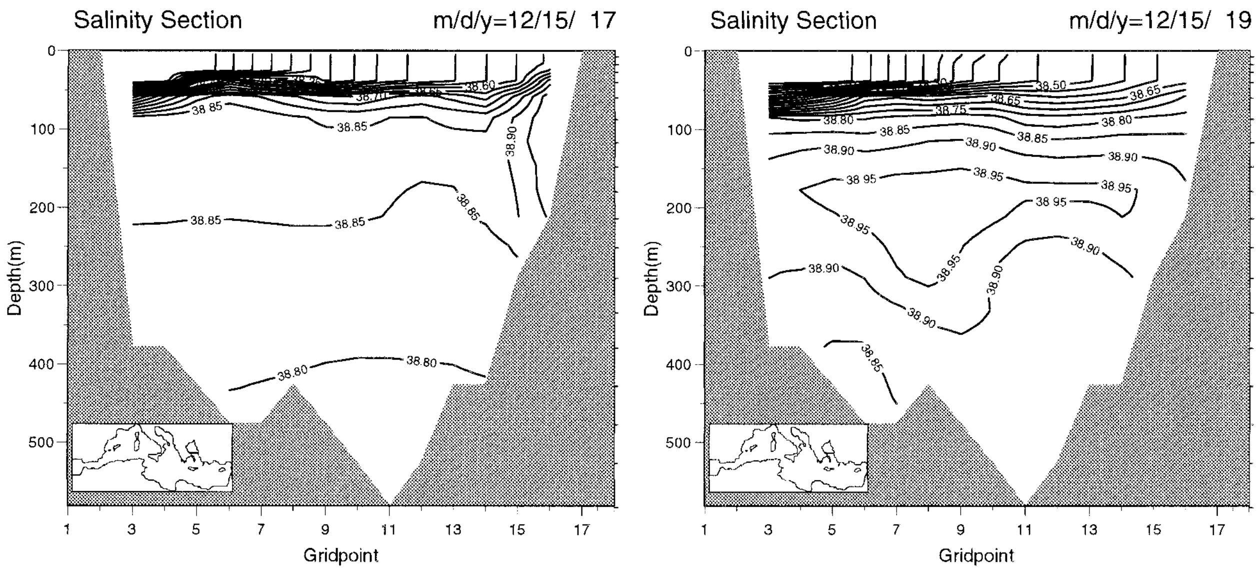

FIG. 11. Vertical cross sections of salinity east–west across the northern Aegean before and after the extra cooling was applied.

western Cretan strait overflows was to the new water mass, but the total lack of deep outflow from the west is not consistent with the current interpretation of the data. This may be the result of too shallow straits in the model, but further work will be needed. The important information provided by Fig. 10 is the change of water properties when it arrives at the bottom of the Levantine Basin. The new water is clearly warmer, but saltier, with higher density and CFC12 concentration. The core of the new deep water outside the Cretan Sea has a temperature of 13.8°C, salinity of 38.85 psu, and σ_θ of 29.2. At the edge of the new water, its properties have been modified as it mixes with the colder and fresher, old EMDW.

The outflow of the Aegean bottom water from the Cretan Sea is compensated by extra upper-layer inflow, which means Levantine surface and intermediate water. Figure 11 compares subsurface salinity prior to extra cooling being applied in the northern Aegean (Fig. 11a), and two years after the extra cooling is first applied (Fig. 11b). We can see a dramatic increase of salinity below the surface in the second figure. In Fig. 11a there is an intermediate layer of salty water, but its maximum salinity is only about 38.85 psu, apart from the narrow boundary current along the Turkish coast. In Fig. 11b the whole layer has expanded, almost doubling in thickness, and its maximum salinity has increased to above 38.95 psu. The bottom water, as a result, has also become more saline, from below 38.8 to above 38.85 psu.

4. Discussion

The comparison between model simulations and observations of the 1987 and 1995 deep-water properties in the Eastern Mediterranean show remarkable consistency, both qualitatively and quantitatively. The change of water masses from one regime to the other is very large, as seen by comparing the right and left columns of Fig. 4. We do not have sufficient data to follow the evolution of the hydrographic structure in the Eastern Mediterranean from the pre-1987 regime to the 1995 state and we will never be able to go back in time to monitor the changeover, even if we have enough resources to monitor any future events. The development of water properties in a model can however be studied. The only experimental difference before and after model year 17 is the lower temperatures (by up to 2°C applied to the Aegean Sea in February, nothing else. Although the surface salinity field used to relax the model remains unchanged throughout the experiment, the surface freshwater fluxes may also vary due to model surface salinity changes brought about by horizontal advection. It is therefore interesting to examine any changes in the freshwater fluxes in the model since changes in $E - P$ have often been quoted as a possible major cause of the observed new deep-water properties.

An important problem is to account for the large salinity increase in the deep layers of the Eastern Mediterranean during this period. There are three possible factors contributing to it: an increase of $E - P$ at the surface, an internal vertical redistribution from the layers above, and reduced net salt transport out of the basin. A budget study of heat and salt content in the Eastern Mediterranean shows that from model year 17 to 25 there is an overall heat loss and salinity increase in the basin. The bottom layer (below 1000 m) has a net gain of both heat and salt, but the upper layer (above 1000 m) looses both heat and salt; that is, the bottom layer has been getting warmer and saltier while the upper layer gets fresher and colder. The upper 1000 m decrease in heat content consists of 35% transferred to the deeper layer and the rest to the atmosphere. Alternatively, for

TABLE 1. Comparison of surface fluxes.

Region	Area (10^{12} m^2)	Q (W m^{-2})		$E - P$ (cm yr^{-1})	
		Normal years	Cold years	Normal years	Cold years
West Mediterranean	0.84	−4.9	−4.7	41.4	39.6
East Mediterranean	1.65	−5.3	−7.7	61.3	67.9
Adriatic	0.11	−24.0	−17.1	−33.7	−30.1
Aegean	0.08	−15.7	−55.4	−74.2	−69.6
Total	2.48	−5.1	−6.7	54.6	58.3

every joule of heat energy lost to the atmosphere the upper 1000 m looses 1.5 J while the layer below 1000 m gains 0.5 J of heat energy. The extra heat loss over the Aegean is equivalent to about 40 W m^{-2} on the annual average, which is equivalent to about 2 W m^{-2} extra heat loss averaged over the whole eastern basin. For the eastern basin model salinity budget, 77% of the salinity increase in the layer below 1000 m is due to salt exchange with the upper 1000-m layer, and only 23% from the increased surface $E - P$ over the eastern basin or from net transport through The Sicilian Channel. Alternatively, for every 1 kg of salt added to the eastern basin due to increased $E - P$ or Strait of Sicily transport, the layer below 1000 m gains about 4 kg of salt and the layer above 1000 m looses about 3 kg of salt. These figures need to be compared with detailed observational budgets, which are being presently constructed.

Table 1 provides a comparison of surface fluxes between normal years (climatological forcing) and cold years (extra cooling in the Aegean) for our model simulation. The normal year fluxes are averaged over 10 years and the cold year fluxes are averaged over the first 5 years of application. In normal years, the surface heat loss for the entire Mediterranean is 5.1 W m^{-2}, compared to observational estimates of 7 ± 3 W m^{-2} (Bethoux 1979; Garrett et al. 1993) and more recently, 5.3–6.2 W m^{-2} by Macdonald et al. (1994). The averaged heat flux for the Eastern Mediterranean is 5.3 W m^{-2}, a little more than for the west, 4.9 W m^{-2}. The freshwater loss for the entire basin is 54.6 cm yr^{-1}, a little lower than observational estimates, based on both terrestrial and aerological water cycle budgets, of around 70 cm yr^{-1} by Gilman and Garrett (1994). When the change of SST is turned into heat flux, the eastern basin annually averaged heat loss is increased from 5.3 to 7.7 W m^{-2}, most of the increase occurring in the Aegean. The average heat loss over the Aegean alone for the cold years is about three times that in normal years, similar to the ECMWF fluxes given in Fig. 1. The deep water temperatures in the Cretan Sea, which resulted from the extra cooling, are also comparable to the measurements of Theocharis et al. (1996). Theocharis et al. have shown that the depth-averaged temperature for the deep water below 600 m the Cretan Sea decreased by 0.4°C during the period 1986–95. Figure 12 shows a comparison of temperatures in the Cretan Sea for our model simulation between year 17 before the cold event, and year 25.

It is of great interest to note the change of $E - P$.

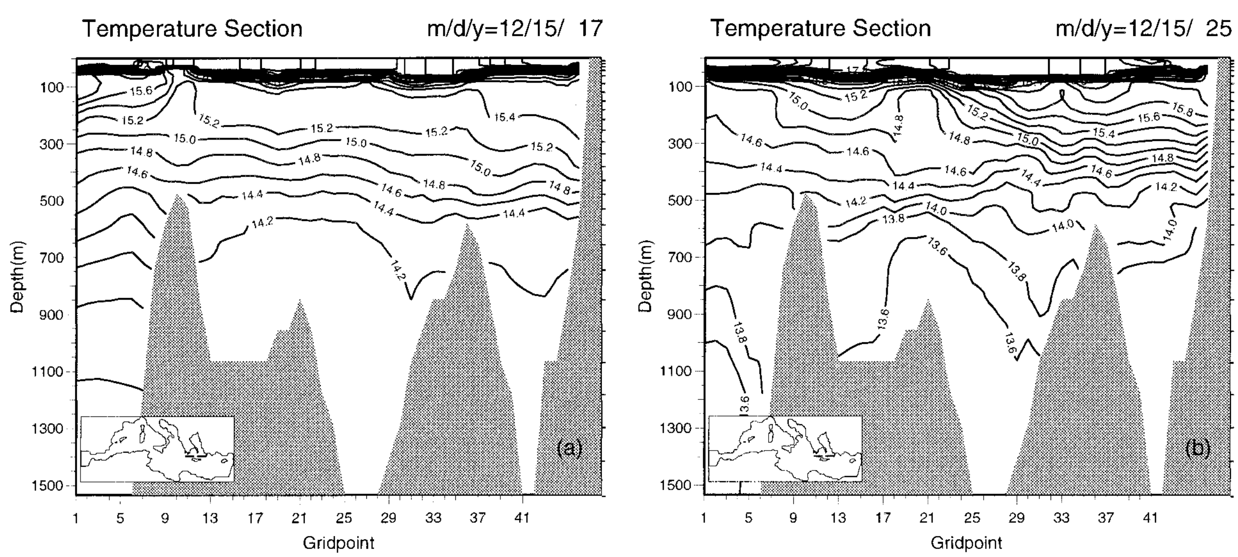

FIG. 12. Temperature cross sections of the Cretan Sea showing a similar amount of decrease in the deep layers as Theocharis et al. (1996)'s measurements: Temperatures (a) in year 17 and (b) in year 25.

There is a small increase of $E - P$ for the entire Mediterranean from 55 to 58 cm yr^{-1} from normal years to colder years. For the eastern Mediterranean as a whole the $E - P$ has increased by 11%, most of which occurs outside the Aegean Sea and the Adriatic. The increase of $E - P$ in the Aegean is only 6%, the majority of $E - P$ increase over the Levantine and the Cretan Sea. This 11% (6.6 cm yr^{-1}) increase of net $E - P$ over the Eastern Mediterranean is much less than the 20 cm yr^{-1}, which would be needed to provide the total salinity addition to the deep water, as estimated by Roether et al. (1996), if the new EMDW was a direct result of increased $E - P$. The fact that in the model most of the salinity increase appears outside the Aegean might suggest an enhanced LIW production. However, Fig. 4 shows the vertical extent of LIW in the Ionian Sea is greatly reduced, together with the maximum salinity. This is further revealed by a plot of salinity on the characteristic LIW isopycnal surface, $\sigma_\theta = 29.0$, shown in Fig. 13. Comparing the bottom (year 25) to the top (year 17), we find there is significantly less LIW in the Ionian and the Levantine where most of the increased freshwater loss occurs. This is actually consistent with the model forcing because to increase the $E - P$ the surface salinity must be lower. There is, however, more LIW in the Aegean and the Cretan Sea. Clearly the missing LIW is being transported into the Aegean Sea where it is added to the deep layers through the route we have described in the previous section.

A comparison of Fig. 13 with observations would help to verify this view of the altered salinity budget for the eastern basin. Figure 13a shows the familiar picture for LIW dispersal (Wu and Haines 1996, 1998), with the transport of LIW into the Adriatic along the Greek coast, indicated by salinities greater than 38.85 psu. Even the concentration of LIW within the Mersa–Matruh gyre is visible. It is interesting to compare Fig. 3b with Fig. 4 of Malanotte-Rizzoli et al. (1996), which shows the observed distribution during April 1995. The modeled distribution matches the observations remarkably well. The most obvious feature is that the westward transport of LIW into the Ionian Sea has been dramatically reduced, which has also been realized in Malanotte-Rizzoli et al. (1999) for an analysis of the POEMBC-091 dataset. This then has an important impact in reducing deep-water formation in the Adriatic, as indicated by the reduced heat loss shown in Table 1.

5. Conclusions

A moderate increase of freshwater loss over the Eastern Mediterranean (11% or 6.6 cm yr^{-1} in the experiment), plus some repeated cold winters, which are able to decrease the sea surface temperature in the Aegean by 1°–2°C, would be capable of switching the major EMDW formation site from the Adriatic to the Aegean and altering the EMDW structure from the pre-1987 state to the 1995 regime as revealed by the two *Meteor* cruises M5 and M31. It does not necessarily require a large increase of $E - P$ in the Aegean itself to produce the salty bottom water observed in the Cretan Sea (the South Aegean). An increased transport of LIW into the Aegean would increase the salinity of the Aegean bottom water. However the major salinity increase for the new EMDW may have occurred in the Cretan Sea where new cold, but fresh, Aegean bottom water could meet the LIW and convect to the bottom. Such internal convection causes the temperature to drop but the salinity to rise in the deep layers of the Cretan Sea as reported by Theocharis et al. (1996). When the Cretan deep water flows through Kasos Strait outside of the Cretan Arc, it falls again to the bottom of the Levantine Basin and spreads into the Ionian Sea. During this second course of internal convection the water is diluted further due to mixing with old EMDW.

Two processes should be emphasized here. The mixing of the Aegean water with LIW in the Cretan Sea plays an important role in setting up the new EMDW properties, but it differs from the deep-water formation processes in the Adriatic and the Gulf of Lions. This is because LIW does not have a direct contact with the surface when cooled further in the Cretan Sea, whereas in the Adriatic and the Gulf of Lions it is entrained into the mixed layer with direct contact to surface forcing. The repeated cooling of the Aegean may be an important factor for the massive outflow into the Levantine Basin. The 8 years of cooling applied in the model may be too long, but a single cold year is probably not enough for the large amount of the new EMDW as observed. The recent work by Malanotte-Rizzoli et al. (1999) has shown that the EMDW transition observed by Roether et al. (1996) during the 1995 *Meteor* cruise started actually before 1991.

This scenario is proposed as a possible explanation for the unexpectedly large changes to the deep waters and thermohaline circulation seen in the Eastern Mediterranean over the past decade. The model simulation has shown a successful reproduction of many of the observed features, both qualitatively and quantitatively. The salinity contributions to the model deep water have been shown to come mainly from an internal redistribution (i.e., reduced LIW in the upper layers), a moderate increase of surface $E - P$, and a reduced transport of salt through the Otranto Straits into the Adriatic and through the Sicilian Channel into the Western Mediterranean. We have emphasized the intensive cooling over the Aegean as a triggering factor, however, such extra cooling usually comes with strong winds and consequently increased evaporation. These factors may all have contributed to the actual process that occurred in reality. Changing winds, for example, may be important for thermohaline circulation changes, as suggested by Myers et al. (1998) and Samuel et al. (1999). Our scenario presents one possible mechanism, and we do not exclude contributions from other factors. There may also have been some pre-conditioning and intermediate

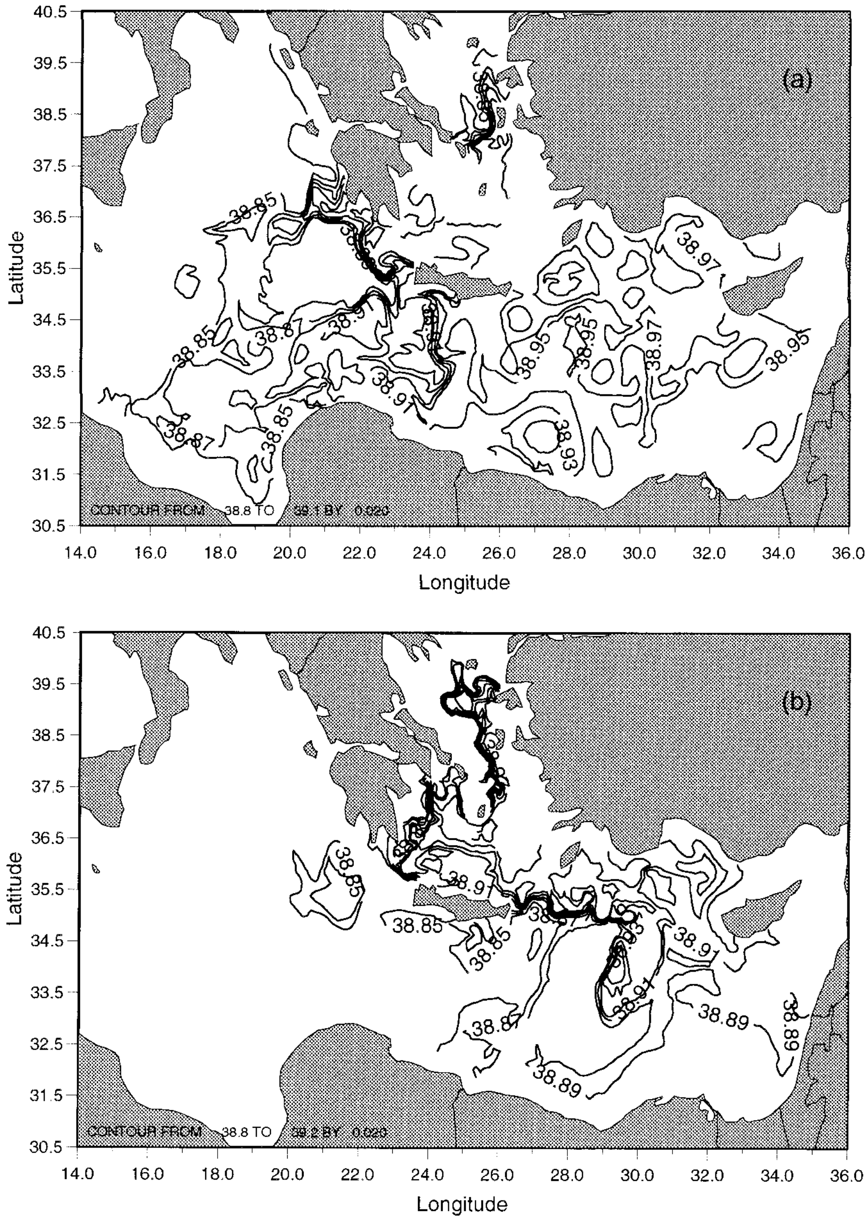

FIG. 13. Salinity on the $\sigma_\theta = 29.0$ isopycnal surface showing the change in LIW distribution in the Eastern Mediterranean before and after the deep water transition on 15 Dec (a) for model year 17 and (b) for model year 25. Only contour values greater than 38.85 psu are plotted, with an interval of 0.02 psu.

processes that we have not considered in our theory, such as the large-scale cooling of LIW during the early eighties as reported by Brankart and Pinardi (1998, manuscript submitted to *J. Phys. Oceanogr.*). A complete understanding and a more realistic simulation of deep-water renewal processes in the Eastern Mediterranean in general, and the 1987–95 event in particular, will require further work with better models and data, particularly meteorological data and air–sea fluxes. The impact of the 1987–95 event on the thermohaline circulation of the Mediterranean as a whole and on the Western Mediterranean, and even possibly on the Gibraltar outflow, should also be an interesting and important topic to investigate.

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