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Simulation of the Mediterranean Sea circulation from 1979 to 1993: Part I. The interannual variability

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Abstract

The interannual variability of the Mediterranean (MED) circulation from 1979 to 1993 is studied with a $\frac{1^\circ}{8} \times \frac{1^\circ}{8}$ resolution OGCM. The surface forcing used is 6 hourly ECMWF (European Center for Medium Range Weather Forecast) reanalysis data. Two different periods in the surface forcing and model variability are identified during 1981–1993: the first, Period I (1981–1987) and the second, Period II (1988–1993). Changes in the model response between the two periods are driven by corresponding differences in the surface forcing, which presumably are a result of the decadal scale changes of the Northern Hemisphere (NH) atmospheric regimes, related to the intensification of North Atlantic Oscillation (NAO) at the end of the 1980s. During the second period (1988–1993), the Mediterranean circulation reveals an overall weakening of the kinetic energy in the Western Mediterranean (WMED) basin and significant changes in the structure of circulation in the Eastern basin. In the latter region, the anticyclonic activity increases in the southern Ionian and in the southern area of the mid-Mediterranean Jet. These anticyclonic eddies, present during different years of Period II with variable intensity, have an important impact on the transport of Modified Atlantic Waters (MAW) and Levantine Intermediate Waters (LIW) in the Eastern Mediterranean (EMED). This change of the circulation modified the salinity and the amount of Levantine Intermediate Waters transported towards the Aegean Sea and the Adriatic Sea, which are important factors for deep water formation processes there. These model results are in good agreement with available observational results. The deep water formation event observed in the Aegean Sea [Science 271 (1996) 333] is produced by the model, but at a shallower depth. We interpret this event as the result of circulation changes between Period I and Period II and anomalous surface atmospheric forcing over the Aegean Sea.

Keywords: Mediterranean Sea; Atmospheric forcing; Circulation; Deep water masses; Intermediate Water Masses; Interannual variability

1. Introduction

The Mediterranean (MED) Sea is a semi-enclosed region consisting of two main parts—Eastern (EMED) and Western (WMED)—which communicate through the relatively shallow Strait of Sicily (Fig. 1). The basin variability is influenced by a relatively large range of processes and their interactions (Robinson and Golnaraghi, 1994). Physical processes typical of the open ocean, such as deep water formation and the overturning thermohaline circulation play an important role for the overall dynamics of the basin. At the same time, the dynamics of its regional seas (e.g., the Adriatic and

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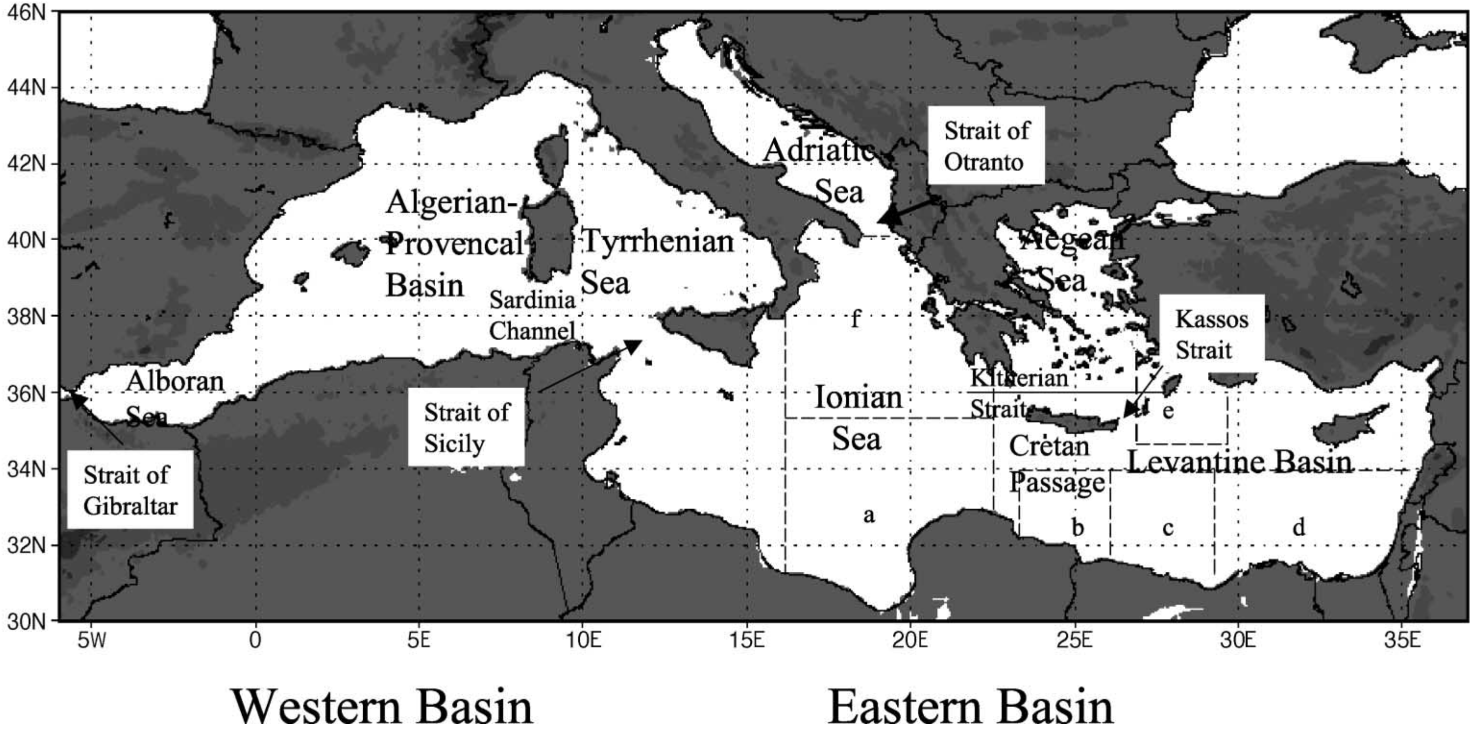

Fig. 1. The Mediterranean Basin configuration. The boxes a, b, c, d, e and f indicate the areas of averaging of salinity in Fig. 14. The line across the Cretan Basin shows the section for Fig. 13.

Aegean) and the transport through the relatively shallow and narrow straits (Gibraltar, Sicily, Otranto, Kassos) are of primary importance for the circulation variability observed in the deep part of the MED.

The MED Sea is connected with the North Atlantic through the relatively narrow and shallow (400 m) Strait of Gibraltar (Fig. 1). The inflow of Atlantic surface waters drives the so-called Atlantic Stream System (ASS) in the Alboran Sea, which transports relatively fresh waters of Atlantic origin to the easternmost part of the EMED influencing the properties of the surface waters in the whole basin. Beyond the Alboran Sea, the ASS becomes the Algerian Current (AC). In the eastern part of the WMED, the AC splits into two parts, with a northern branch entering the Tyrrhenian Sea and a southern branch entering Strait of Sicily. In Strait of Sicily and Ionian Sea the flow is known as Atlantic–Ionian Stream (AIS), which travels in a zonal direction across the basin and enters the Levantine Basin, where it gives rise to the mid-Mediterranean Jet (MMJ). The MMJ flows eastward between the Rhodes gyre on the north and the Mersa–Matruh gyre and the area of the Shikmona gyre on the south. Beyond Cyprus a branch of the MMJ becomes the Asia Minor Current (AMC), which flows along the Turkish coast. The Modified Atlantic Water (MAW) transported by the AIS, MMJ and AMC reaches the easternmost part of the basin, mixing continuously with the surrounding waters and deepening to about a 50- to 100-m depth in the Levantine Basin. The salinity of the waters entering through Gibraltar is about 36.15 psu, while the MAW in the Levantine Basin becomes 38.6 psu (see Ozsoy et al., 1989).

Wintertime intermediate convection in the Levantine Basin produces Levantine Intermediate Waters (LIW), which are transported westward in the layer between 300 and 500 m towards the Strait of Sicily and then towards Gibraltar. The surface transport of waters of Atlantic origin by the ASS and the westward propagation of LIW at intermediate levels are the main parts of the overturning thermohaline cell of the Mediterranean Sea (Robinson and Golnaraghi, 1994).

Two meridional thermohaline cells form in WMED and EMED driven by the processes of deep water formation. The deep waters of the WMED (WMDW) form in the Gulf of Lions by deep convection processes during cold winters (Leaman and Schott, 1991). The presence of a cyclonic gyre in the Gulf of Lions is an important preconditioning factor for this winter deep convection process. The Eastern Mediterranean Deep Water (EMDW) forms in the Adriatic (Roether and Schlitzer, 1991) or the Aegean Sea (Roether et al., 1996) and then, through the relatively narrow and shallow straits, sinks into the deeper parts of the basin.

Recent observational studies in the MED presented evidence of noticeable large interannual variability in the 1980s and the beginning of the 1990s. Hecht (1992) found significant changes in salinity of the surface and intermediate water masses in the EMED during 1982. Brankart and Pinardi (2001) showed that at 200 m, a strong abrupt cooling occurred during period 1981–1983 around the Aegean Sea. The LIW changed approximately by 0.4 °C between the decades 1971–1980 and 1984–1993 due to changes in the atmospheric forcing. Korres et al. (2000) found that in 1987 the Ionian Sea surface circulation reversed with respect to the 1981 pattern due to the slowly varying atmospheric circulation.

In addition, significant interannual variability was also found in the deep water formation processes and water structure. Bethoux et al. (1990) showed that from 1959 to 1988, the WMDW temperature and salinity had increased by 0.12 °C and 0.03 ppt, respectively. Roether et al. (1996) found that during the period 1987–1995, the source of EMDW shifted from the Adriatic to the Aegean Sea, when, the outflow of deep waters from the Cretan Sea (Cretan Sea Outflow Water, CSOW; Klein et al., 2000) filled the deep part of EMED at a rate of about 1 Sv, replacing partly the Adriatic waters (EMDW). These changes in the deep water characteristics and formation areas were accompanied by a change in the deep and intermediate layers circulation structure as discussed by Malanotte-Rizzoli et al. (1999).

In this work, we investigated the nature and causes of the interannual variability of Mediterranean Sea for the period 1979–1993 using ocean circulation model simulations. Previous model studies of the Mediterranean during this period or sub-periods of it were done by Samuel et al. (1999), Nittis and Lascaratos (1998), Lascaratos et al. (1999), Korres et al. (2000) and Castellari et al. (2000). These studies reproduced some of the major elements of the known variability of the Mediterranean Sea. However, they were made with coarse resolution models or for the EMED only. At the same time Malanotte-Rizzoli et al. (1999) showed that the mesoscale variability had a major role in the transition from 1987 to 1995 in the EMED and that the MAW distribution and propagation in the EMED changed.

The aim of the present work is to study the related variability of atmospheric forcing, circulation and water mass formation processes, for the whole MED in 1980s and beginning of 1990s using an eddy permitting model in an attempt to verify and test different hypothesis on the causes of the observed variability. We do this by using an OGCM with a horizontal resolution of $\frac{1^\circ}{8} \times \frac{1^\circ}{8}$ and high-frequency (6 hourly) surface atmospheric forcing to simulate the variability of the MED during the 1980s and 1990s. Section 2 describes the model set up. In Section 3 we introduce the external forcing used in the simulations. In Section 4 the model results and comparison with existing data and the origin of major transition of the Mediterranean circulation between the periods 1981–1987 and 1988–1993 are discussed. The last section offers the conclusions.

2. The model

The model used is the Modular Ocean Model (MOM), which was adapted to the Mediterranean Sea by Roussenov et al. (1995). The model grid has 31 vertical levels, as in Korres et al. (2000), and a horizontal resolution of $\frac{1^\circ}{8} \times \frac{1^\circ}{8}$, which does not include the northern part of the Adriatic Sea. Horizontal turbulent mixing is biharmonic with tracer coefficients equal to 1.5×10^{10} m^4/s and momentum coefficients equal to 5×10^9 m^4/s. Vertical turbulent processes are parameterized by a constant turbulent diffusion coefficient set equal to 0.3×10^{-4} m^2/s, a viscosity coefficient equal to 1.5×10^{-4} m^2/s. A standard convective adjustment procedure (Cox, 1984) is applied when static instability appears in the water column. This vertical mixing scheme choice has been studied in the previous works of Korres et al. (2000) and Castellari et al. (2000).

Transport through Strait of Gibraltar is parameterized by extending the model area westward of Gibraltar to a longitude 9.25°W. In this Atlantic box, lying between latitudes $33°30' \mathrm{N} < \varphi < 37°\mathrm{N}$, the surface forcing is switched off and temperature and salinity are relaxed towards annual mean climatological fields.

The surface forcing is computed in an interactive way with 6 hourly ECMWF atmospheric reanalysis fields and Sea Surface Temperature (SST) from the model. The different components of the surface net heat flux are computed on the basis of the parameterizations of Reed (1977), for the surface solar radiation flux, of Bignami et al. (1995) for outgoing long-wave radiation and of Kondo (1975) for sensible and latent heat fluxes. The bulk formulation of Hellerman and Rosenstein (1983) is used in the wind stress computation. Descriptions of the implementation and test of the surface momentum and heat flux parameterizations can be found in Castellari et al. (1998, 2000). The meteorological data used are atmospheric temperature and humidity at 2 m and wind components at 10 m. Cloud cover is taken from the monthly mean COADS data (Da Silva et al., 1994). The sea surface water flux is parameterized with a salt flux given by relaxation of model sea surface salinity towards a new climatology called MED6 (Brankart and Pinardi, 2001). The relaxation constant is everywhere 2 m/days. As we mentioned in the introduction, the aim of the present study is to diagnose the interannual variability of the MED circulation from a high-resolution model simulation, forced with high-frequency realistic surface momentum and heat fluxes. At the same time, the parameterization of the surface water flux does not contain the important interannual component of the surface forcing. The drawback of this boundary condition shows up in the model sea surface salinity. Comparison of MED6 climatology with independent observations of Russian *R/V Gakel* in the Northeastern Levantine and Aegean basins, show that the MED6 climatology underestimates sea surface salinity during winter 1988 by about 0.1–0.15 psu (Gertman and Popov, 1989). The MED6 climatology tends to underestimate also salinity in the most eastern and southeastern parts of the Levantine Sea (A. Hecht, personal communication).

The model is first run for 7 years with perpetual monthly mean forcing to reproduce the seasonal cycle in the basin, following the same procedure as in Roussenov et al. (1995). This the solution is used then to initialize the simulations starting at January 1, 1979.

3. External forcing variability

The simulated circulation variability in the model is driven by three major types of forcing: inflow–outflow through the Strait of Gibraltar, surface wind stress and buoyancy fluxes. All three inputs are prescribed in an interactive way i.e., they are dependent upon external parameters (climatological temperature and salinity in the Atlantic box and atmospheric parameters) and the model solution. The discussion of the simulation results starts with a short description of the external forcing variability and comparison with existing data and model results.

The model used in our simulations has a rigid lid, and the inflow and outflow through the Strait of Gibraltar are balanced at each time step. The mean volume transport for the whole period is 0.85 Sv. The amplitude of annual variability is relatively small and has a maximum of 0.3 Sv in 1981 and a minimum of 0.05 Sv in 1987. The model transport through the Strait of Gibraltar is slightly higher than the recent observations made by Bryden et al. (1994) and Tsimplis and Bryden (2000). According to the latter study, the inflow through the Strait of Gibraltar is 0.78 ± 0.47 Sv. The error in these estimates depends on the method of calculation of the integral transport and the uncertainty in the position of the interface position. As in the model simulations, the annual cycle of the transport through the Strait of Gibraltar evaluated from the data is relatively low, namely, 0.03 Sv for the outflow and 0.12 Sv for the inflow. Thus, the inflow/outflow mechanism at Gibraltar is not sensitive to the interannual changes occurring in the circulation and cannot be considered as a forcing for the interannual variability of the basin circulation.

3.1. Surface meteorological forcing

Surface meteorological forcing over the Mediterranean during the 1980s and the beginning of the 1990s has been recently investigated because the changes in meteorological conditions are considered to be an important cause of circulation and water mass variability (Pinardi et al., 1997; Korres et al., 2000; Wu et al., 2000; Castellari et al., 2000). The available observations and model results suggest that there are three changes in atmospheric forcing in EMED during this period, as follows.

(a) According to Tselepidaki et al. (1990) and Lascaratos et al. (1999), the period 1988–1993 was dry with low precipitation values.

(b) According to Korres et al. (2000), 1981 was exceptionally cold and a year of strong winds. Brankart and Pinardi (2001) showed that the years 1981–1983 were cold over the Levantine basin. Some of the years after 1987 were exceptionally cold over the Aegean Sea (see Lascaratos et al., 1999; Wu et al., 2000).

(c) The winter wind regime over the EMED and the WMED for the period 1980–1987 was different from that during 1988–1993. Samuel et al. (1999) showed that from the first period to the second the intensity of Mistral winds decreased, while the Aegean wind regime became much stronger.

The time series of the monthly mean basin averaged surface wind stresses (Fig. 2) for the period 1979–1993 reveals a strong seasonal cycle. The summer minima do not change significantly from year to year, however, the winter time wind stress maxima exhibit strong interannual variability, which is different in the EMED and the WMED. The months with strong wind stress in the WMED (Fig. 2b) are January 1981, December 1981 and January 1986 while in the EMED (Fig. 2c) they are January 1981 and December 1991. In 1981 the wind forcing is very strong in both parts of the basin and reaches an amplitude of 1.7 dyn/cm^2 in the EMED and about 1.4 dyn/cm^2 in the WMED. During 1982 and 1986, the wind forcing is relatively strong only in the WMED. From 1988 to 1990, the wind over both the WMED and EMED has a tendency to decrease, with anomalous weak winds occurring in 1990.

The basin averaged wind stress curl (Fig. 3a) is predominantly positive, giving rise to a net cyclonic vorticity input in to the basin. It changes sign during summer, when it can achieve negative values. Over the whole Mediterranean (Fig. 3a), the highest winter value of the wind stress curl occurs in 1981 and the lowest in 1990. The months with highest values of the wind stress curl in the WMED (Fig. 3b) are December 1981, January and February 1986, January and February 1987 and December 1990. During these months the wind stress curl in the EMED (Fig. 3c) is relatively weak. In the EMED, a relatively high wind stress curl occurs during January 1981, December 1988 and December 1991. A relatively low wind

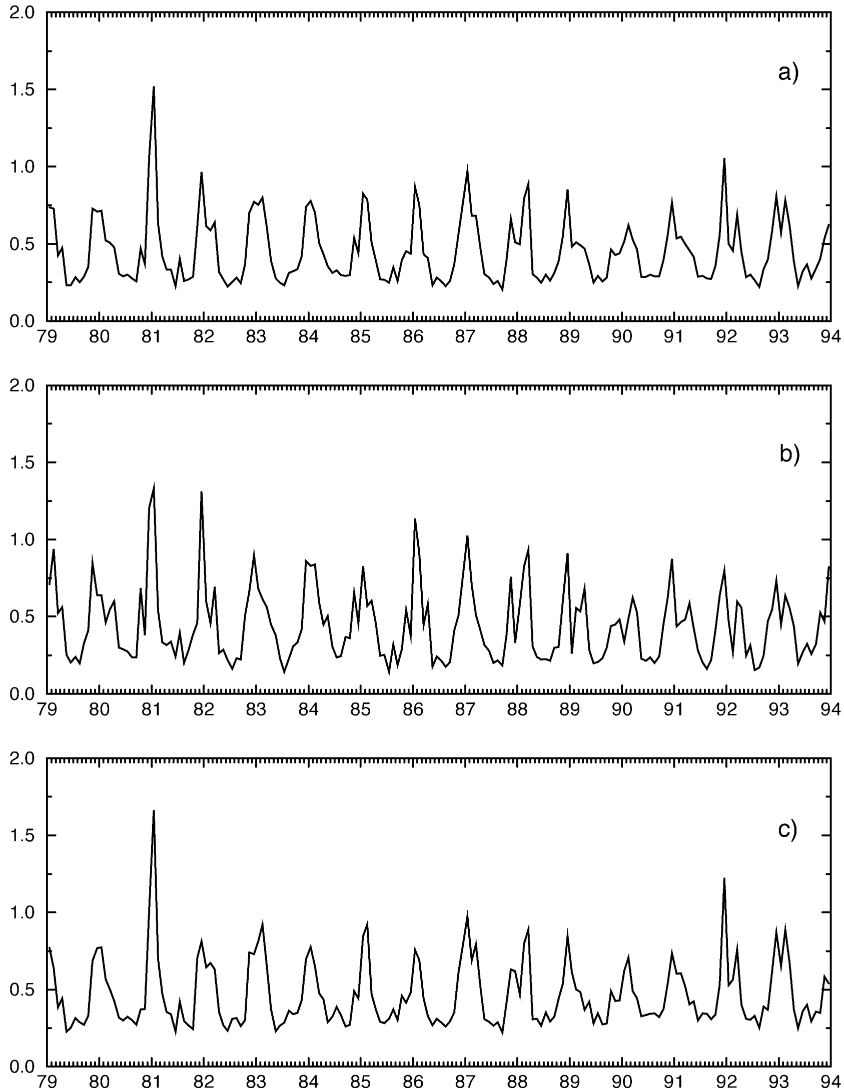

Fig. 2. Surface averaged monthly mean wind stress (dyn/cm^2) calculated for (a) the whole Mediterranean Sea; (b) Western Mediterranean; and (c) Eastern Mediterranean.

stress curl occurs during the whole year in 1989 and 1990 in the WMED, and in 1990 and 1991 in the EMED.

The monthly mean heat fluxes over the Mediterranean (Fig. 4a) reveal three relatively warm summers—1985, 1986 and 1987. During the remaining years in the period 1979 to 1993, the summer maxima do not vary significantly from year to year. On the other hand the interannual variability of the winter cooling is much more pronounced. Over the whole Mediterranean the coldest years are 1981, 1982, 1989, 1991 and1992. In the WMED (Fig. 4b), the coldest winters are in 1981, 1987, 1989 and 1991, while in the EMED (Fig. 4c) these are in 1981, 1982 and 1992. During most of the years the coldest months are December, January and February. There are, however, some exceptional winters, when strong surface cooling can be observed also in March, such as in 1987. Leaman and Schott (1991) documented strong cooling in February–March 1987 in the WMED, which is also

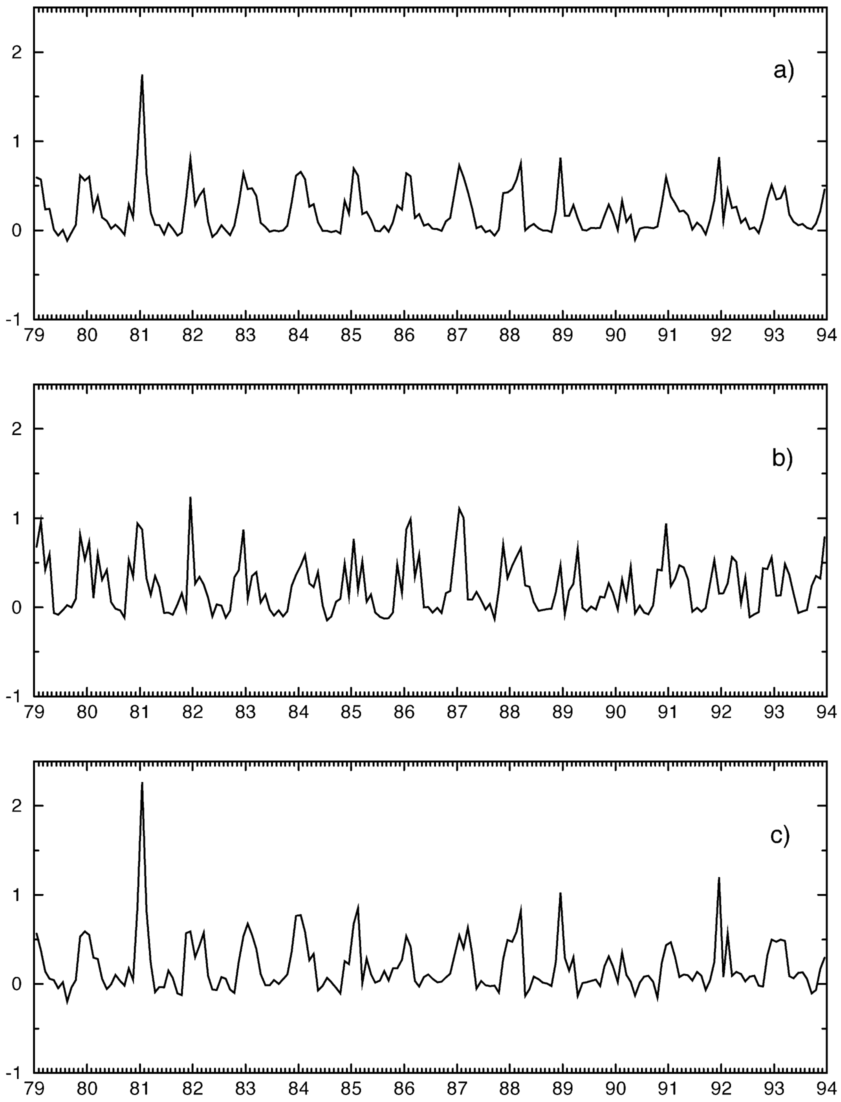

Fig. 3. Surface averaged monthly mean wind stress curl (10^6 dyn/cm^3) calculated for (a) the whole Mediterranean Sea; (b) Western Mediterranean; and (c) Eastern Mediterranean.

present in our data (Fig. 4b). Gertman et al. (1987) describe cooling over the northeast Levantine in March 1987, which is manifested by the two peaks in 1987 in Fig. 4c. In order to represent the integral effect of the whole winter cooling, we calculate the parameter

$$Q_C = -\frac{1}{2T}\int_T (Q(t) - |Q(t)|)\mathrm{d}t$$

where $Q(t)$ is the net surface heat flux and the time integration is made over the whole year. The integral surface heat loss in the MED (Fig. 5a) is strongest during the years 1981, 1983, 1987 and 1992; in the WMED (Fig. 5b)—1981, 1987 and 1991; and in the EMED (Fig. 5c)—1981, 1983, 1987 and 1992. In addition, it is possible to argue that for the whole period (1989–1993), the EMED shows relatively high heat losses. Fig. 6 presents the time series of the same parameter Q_C but for three areas where deep water

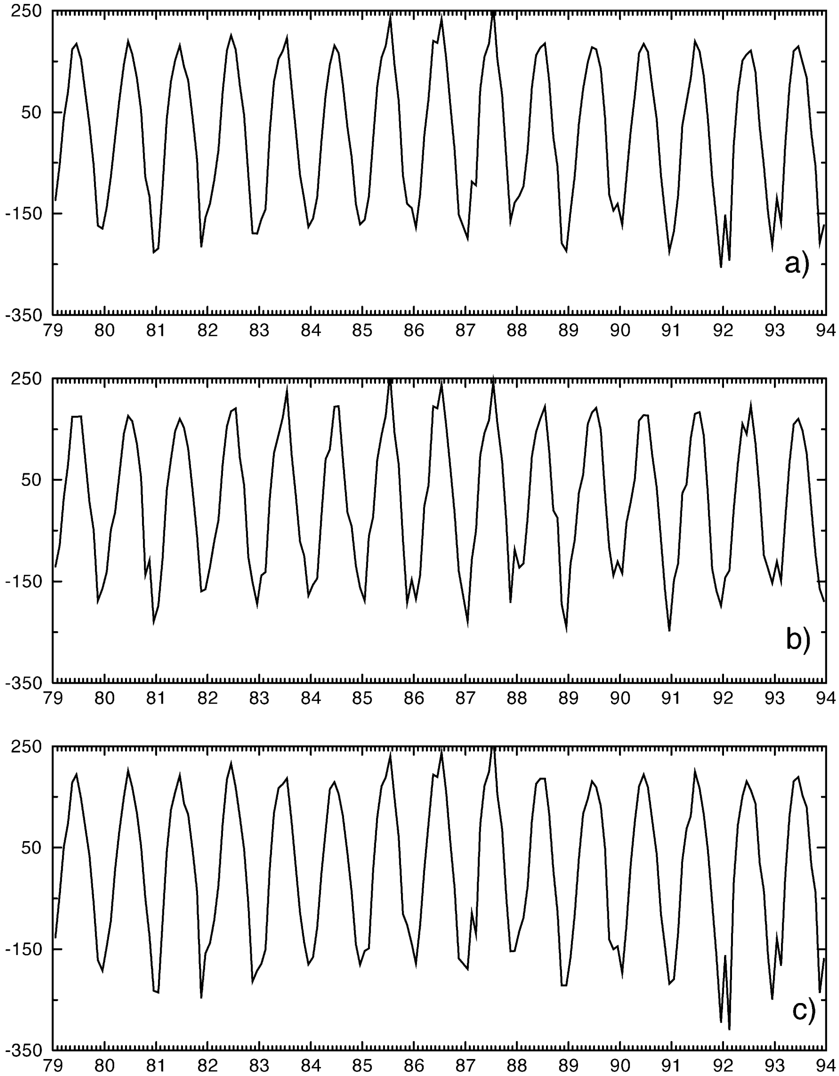

Fig. 4. Surface averaged monthly mean net heat (W/m^2) calculated for (a) the whole Mediterranean; (b) Western Mediterranean; and (c) Eastern Mediterranean.

formation occurs, i.e., the Gulf of Lions, the Southern Adriatic and the Aegean Seas. The coldest years in the Gulf of Lions are 1981 and 1987; in the Adriatic, 1981, 1987 and 1992; and in the Aegean, 1987, 1990, 1992 and 1993. In general, we can conclude that during the period 1979–1993 there is a shift of the areas of enhanced cooling from WMED and Gulf of Lions (years 1981 and 1987) towards the Aegean Sea (1987, 1990, 1992–1993). It is thus possible to identify two periods, which will be called hereafter Period I and Period II. The first extends from 1981 to 1987, and the second from 1988 to 1993.

The winter (December, January, February—DJF) wind distributions for the two periods, 1981–1987 (Fig. 7a) and 1988–1993 (Fig. 7b), are very similar to the results of Samuel et al. (1999). During Period I, the Mistral winds over the WMED are relatively strong, and the wind over the central and eastern parts of the EMED has a predominant zonal component. The mean winds over the Aegean and Adriatic are

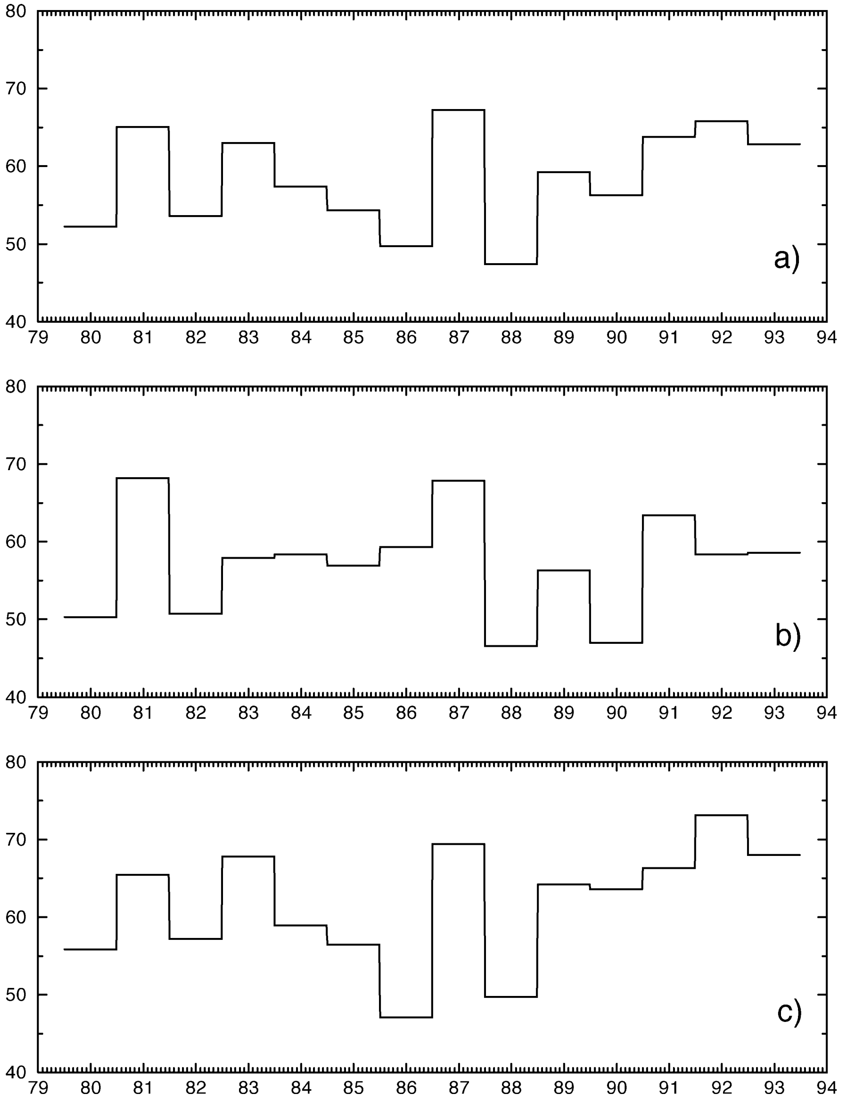

Fig. 5. Surface averaged annual mean net cooling (W/m^2) calculated for (a) the whole Mediterranean Sea; (b) Western Mediterranean; and (c) Eastern Mediterranean.

relatively weak. During the Period II, the Mistral weakens while the Aegean winds (also called the Etesian winds during summer season) become much stronger.

The distribution of the mean wind stress curl for the winter season (Fig.7a,b) has similar elements for both periods—cyclonic vorticity input over the northern Ligurian–Provencal basin, Tyrrhenian Sea, northern and central Ionian and Eastern Levantine and anticyclonic input in the Eastern Catalan Sea, Algerian Basin, Southeastern Mediterranean and Western Aegean Areas. There are (see Pinardi and Navarra, 1993) four areas of "wind curl dipoles" where the wind changes in a relatively small distance from cyclonic to anticyclonic vorticity input: the first one is in the Gulf of Lions, the second is between Sardinia and the Northern African coasts, the third is in the Ionian Sea and the fourth is over the Aegean Sea and

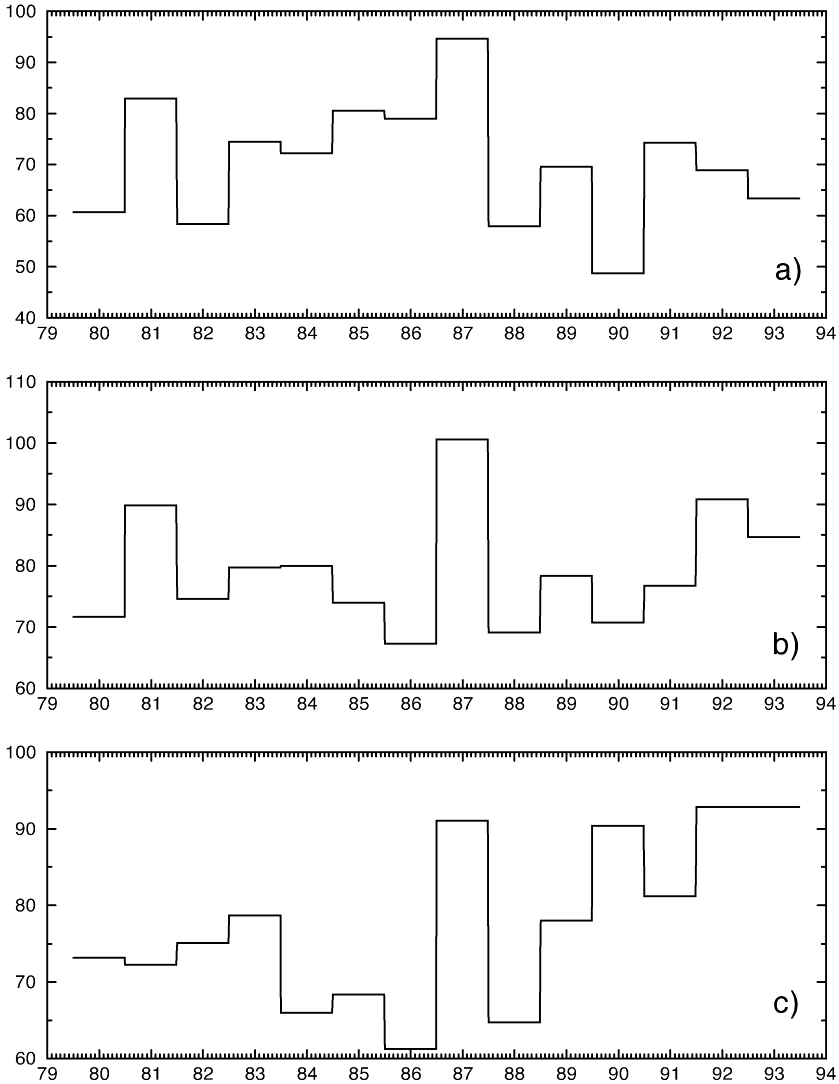

Fig. 6. Surface averaged annual mean net cooling (W/m^2) calculated for (a) Gulf of Lions; (b) Adriatic Sea; (c) Aegean Sea.

Northern Levantine basin. During the second period, the intensity of the dipoles decreases in the Gulf of Lions and south of Sardinia and increases in the Ionian and Aegean Seas.

In the EMED, the wind stress direction and curl reveal some significant changes from the first period to the second. During Period I the wind stress curl is positive (cyclonic) over the entire area of the Central and Northern Levantine and Northern Ionian Seas (Fig. 7a). During Period II this area is divided into two—one over the Eastern Levantine, where the wind stress curl increases with respect to Period I and a second over the South Adriatic and Northern Ionian (Fig. 7b). In general, the cyclonic vorticity input decreases in the WMED, Sicily Strait, Ionian and Western Cretan Passage (Fig. 1), as shown in Fig. 7c where the difference of the wind stress curl between Period I and Period II is reproduced.

In Fig. 8, the wintertime heat flux for both periods is presented. The tendency of the heat flux between Period I (Fig. 8a) and Period II (Fig. 8b) to increase in the WMED and Ionian Sea is evident. In addition, the

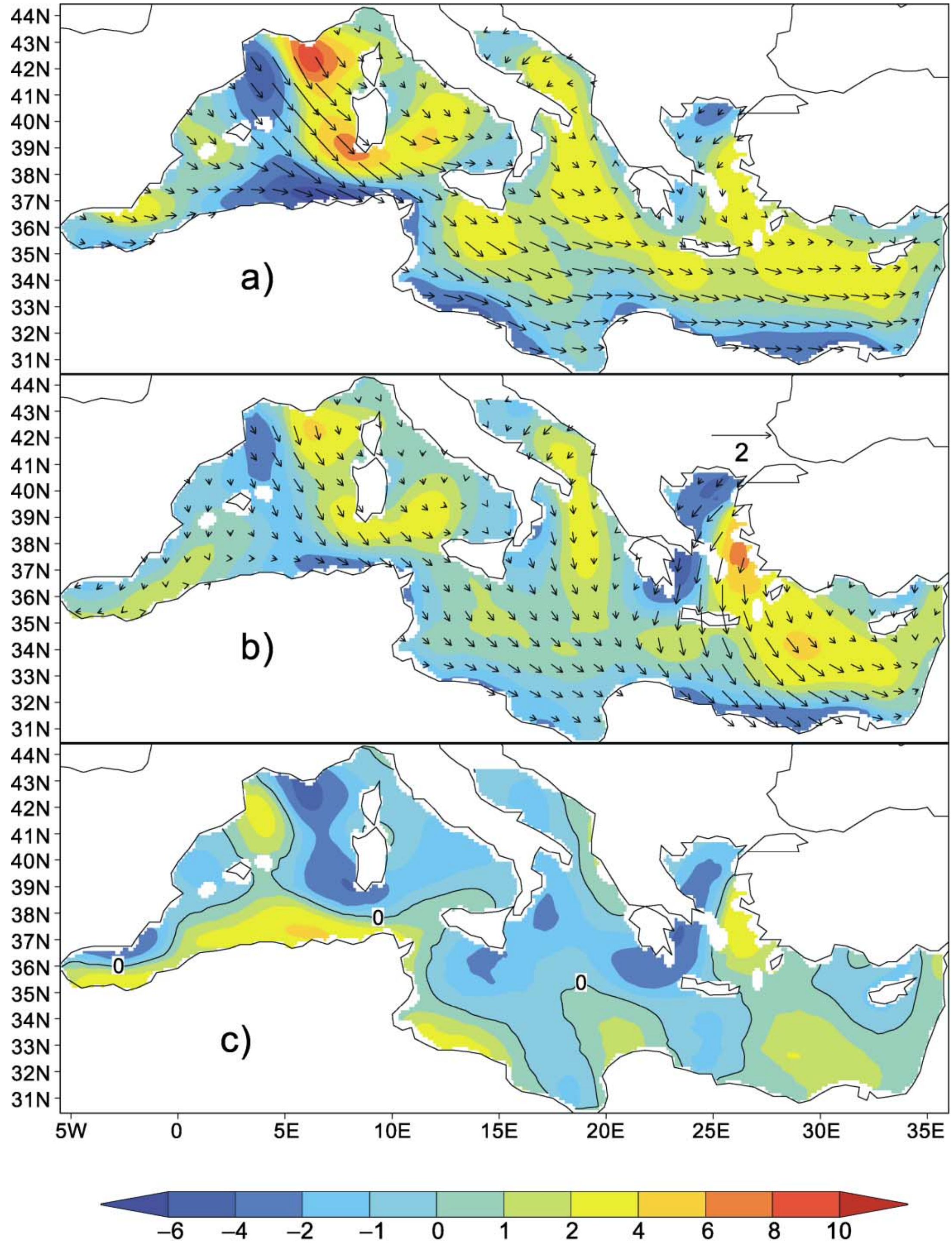

Fig. 7. Winter (DJF) wind stress (dyn/cm^2) and wind stress curl (10^6 dyn/cm^3) (a) averaged for Period I (1981–1987); (b) averaged for Period II (1988–1993). (c) Difference between the wind stress curl of Period II (1988–1993) and Period I (1981–1987) (10^6 dyn/cm^3).

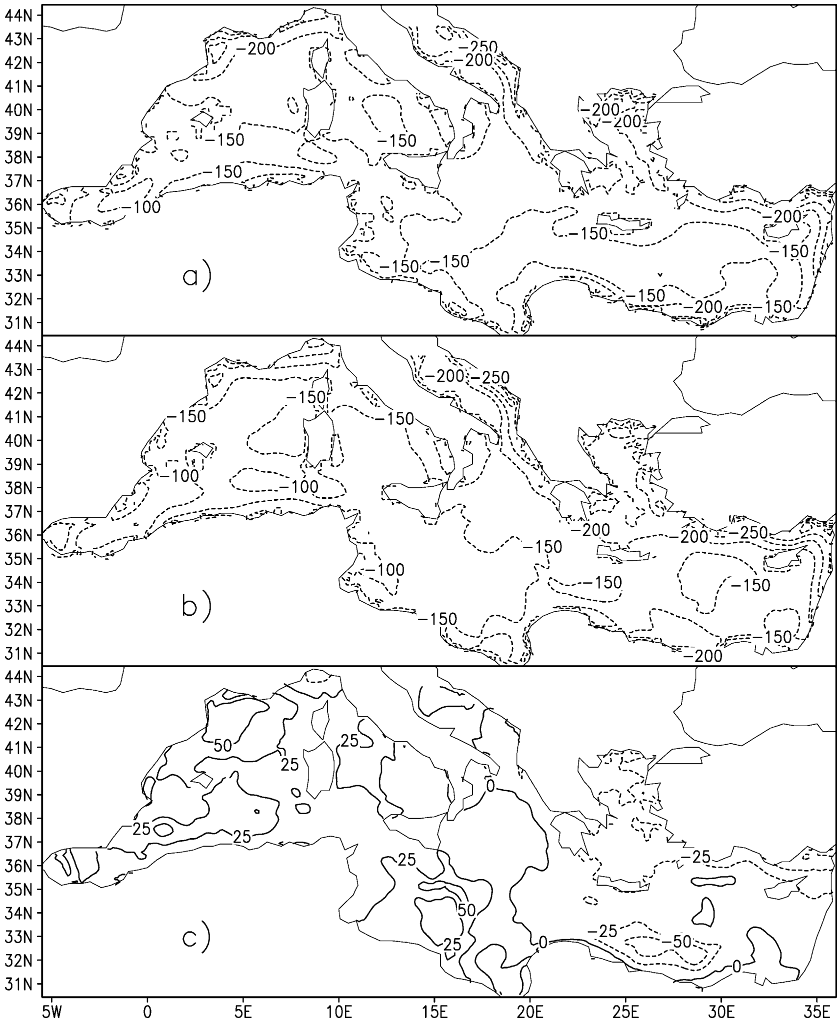

Fig. 8. Winter (DJF) surface net heat flux (W/m^2) (a) averaged for years 1981–1987; (b) averaged for years 1988–1993. (c) Winter mean net heat flux differences 1988–1993 mean less 1981–1987 mean.

heat flux decreases over the Levantine and Aegean Seas. The areas where the mean winter surface cooling decreases significantly during Period II (Fig. 8c) are the Gulf of Lions, Eastern Algerian Basin and southern Ionian Sea. The areas of increased winter cooling are the Mersa Matruh area, the Northeastern Levantine and the Central Aegean Sea.

3.2. Period I and period II teleconnections with NAO

The brief description of ECMWF reanalysis data given in Section 3.1 suggests that the changes in momentum and heat fluxes are part of more general changes in meteorological conditions, which occurred over the whole MED. The ECMWF data show that the interannual variability in the atmosphere between Period I and Period II was at least as strong over the WMED as over the EMED. Period I coincides with high winter heat losses and a large amplitude Mistral wind stress over the WMED; Period II consists of a general weakening of all momentum and heat fluxes over the WMED and enhanced heat losses and wind amplitudes in the EMED. Recent atmospheric studies of the Northern Hemisphere (NH) circulation regime demonstrated that during the second half of eighties the atmospheric conditions over Europe, the Mediterranean area and Middle East were connected to a large-scale change in atmospheric circulation. In particular they showed that, after the winter 1988/1989, there was an abrupt change in the general circulation of the NH with a persistent intensification of the northern polar vortex (Walsh et al., 1996) and a weakening of the subtropical jets over the Pacific and Atlantic (Watanabe and Nitta, 1999). Correspondingly, the atmospheric temperatures increased over the regions from Eurasia to the North Pacific and from North America to Europe, and decreased over central Eurasia and the Labrador Sea. These decadal changes and especially their abruptness are explained by the phase matching of some long-term variations in the NH regimes, namely the North Atlantic Oscillation (NAO), the Pacific–North America Pattern (PNA) and the Eurasian Pattern (EU) (Watanabe and Nitta, 1999). According to Watanabe and Nitta (1999) and

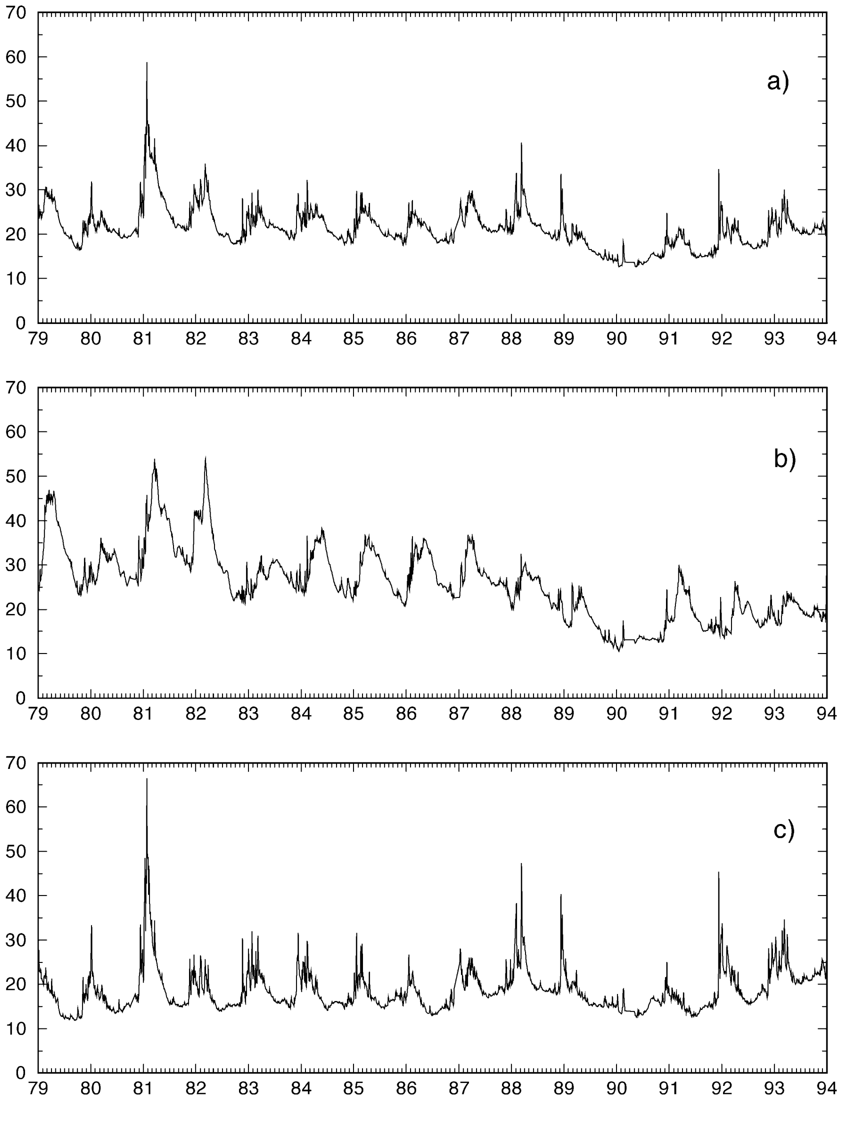

Fig. 9. Total kinetic energy per unit mass (cm^2/s^2) for (a) whole Mediterranean, (b) Western Mediterranean; and (c) Eastern Mediterranean.

Ting et al. (1996), the change over the Atlantic Ocean and Eurasian continent are presumably induced by an intensified NAO.

The NAO (Wallace and Gutzler, 1981) is a teleconnection pattern that exhibits (a) a negative correlation between the severity of winters over the Greenland and the Labrador Seas and (b) a negative correlation between sea level pressure in the Icelandic Low and an east–west high pressure belt centered at 40°N. The intensity and the phase of NAO is measured by the so-called NAO index (Hurrell, 1995): positive values of the NAO index mean a relatively strong Icelandic Low, relatively high pressure along 40°N, strong westerlies across the North Atlantic, low air temperatures over the Middle East and Greenland–Labrador basins and relatively high temperatures over eastern United States and northwestern Europe. The NAO may be important for the atmospheric variability over the Mediterranean Sea since (see Fig. 1 from Wallace and Gutzler, 1981) when the NAO index is high, the air temperatures over the Aegean Sea and Levantine Basin are relatively low, the winds have a strong northerly component, and the winters over the WMED are relatively warm. Honda et al. (2001) show that the NAO index had relatively high values at the end of 1980s and beginning of 1990s, which corresponds well to relatively strong Aegean winds, cold air over the EMED and warm winters over the WMED, which we found from our analysis of the ECMWF forcing for Period II (Section 3.1).

The evolution of the model kinetic energy (KE) (Fig. 9) indicates that an abrupt change occurred after 1987, that was produced by the variability of surface forcing. During 1988 and 1989, the KE in the WMED has a strong tendency to decrease. At the same time, the KE in the EMED increases reaching high values in 1988, 1989, 1992 and 1993. Another remarkable feature of the year 1988 is the relatively high KE in the EMED during summer. Due to the anomalous weak surface forcing, the KE during 1990, attains the lowest values in both basins. In 1991 the KE remains low in the EMED, while it increases in 1992 and 1993. In the WMED, the KE increases in 1991, but for the whole Period II it remains lower than in Period I. A remarkable feature in the evolution of the KE is the decrease of the amplitude of the seasonal cycle of KE in the WMED after 1990 (Molcard et al., in press).

4. The transition of the Mediterranean circulation between Period I and Period II

In this section, we present the model solution for the Mediterranean Sea circulation in an attempt to assess the changes that occurred between Period I and Period II. First we present the model solution for years 1987 and 1991 and compare it with results of Robinson et al. (1991), Roether et al. (1996), Malanotte-Rizzoli et al. (1997) and Malanotte-Rizzoli et al. (1999). We assess how the model represents the variability of the circulation and water mass structure after 1987 and how it describes the main elements of this variability as known from the data.

4.1. Comparison between the 1987 and the 1991 circulations and water mass structure

The period 1979–1993 includes the time of the first (Robinson et al., 1992) and second POEM (Physical Oceanography of the Eastern Mediterranean) experiments (Roether et al., 1996; Malanotte-Rizzoli et al., 1999; Ozsoy et al., 1993). We present results from model simulations for October 1987 and October 1991 in order to identify the changes that occurred in the model solution.

The model results for dynamic height at 5- and 360-m depths, with a reference level of 800 m are shown in Figs. 10 and 11 for October 1987 and October 1991. The figures show the existence of features, which are quasi-permanent and exist during both Period I and Period II, but there are others that are not. In the WMED, the permanent structures are the Alboran Sea western anticyclonic gyre, the Algerian current and eddy system, the cyclonic circulation in the Ligurian–Provencal basin and the western cyclonic and eastern anticyclonic circulation patterns in the Tyrrhenian Sea. The differences between the dynamic topography of the two periods reveal a general decrease during Period II in the intensity of the cyclonic circulation in the Gulf of Lions and weakening of the Algerian current mesoscale activity. The same tendency is present also in intermediate layers (Fig. 11). In 1987, the Algerian current develops a system of cyclonic and anticyclonic gyres, with a size of approximately 100–150 km. In 1991, the mesoscale eddies in the Algerian current are less intense, have smaller horizontal extension, and below

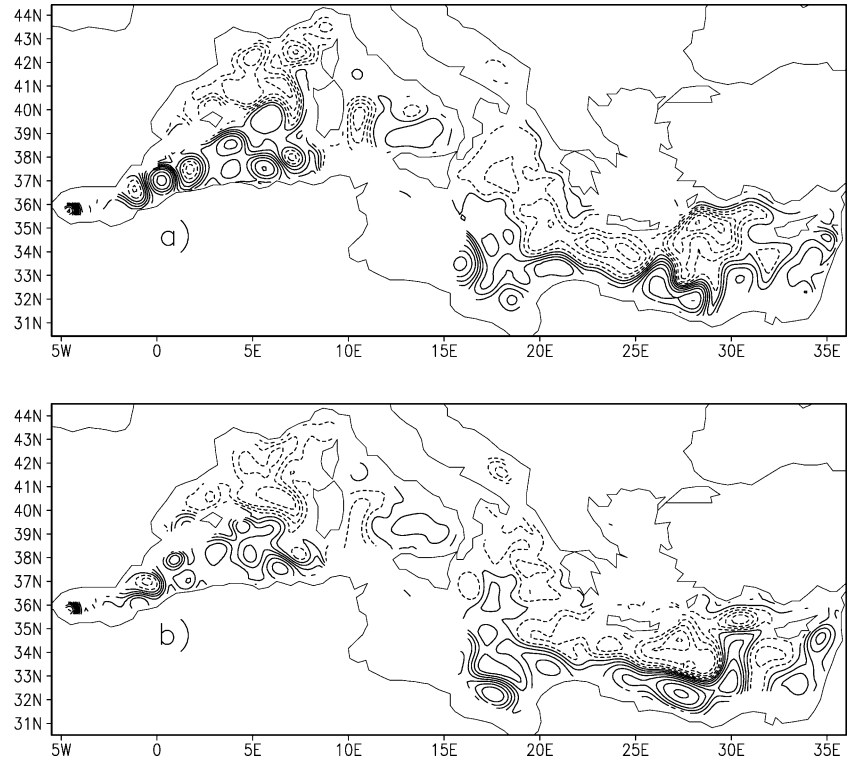

Fig. 10. Monthly mean dynamic height at 5 m with reference level at 800 m. (a) October 1987; (b) October 1991. The contour interval is 2.5 cm.

200 m are very weak and barely present. During this year, one branch of the Algerian Current is strongly meandering around the Balearic Islands.

In 1991 the Tyrrhenian Sea cyclonic circulation is generally weaker at all levels. Thus, we can conclude that the circulation and mesoscale activity in the WMED shows a clear tendency to weaken in 1991 due to the weakening of the surface wind stress and cooling between Periods I and II. In 1987 (Fig. 10), the AIS in the Ionian Sea branches into two parts. The first branch develops into an anticyclonic pattern in the central Ionian. The second and stronger branch of AIS propagates southeastward, strongly meanders and creates a system of three anticyclonic eddies. The first branch has been well studied using the data of POEM by Robinson et al. (1991) and Malanotte-Rizzoli et al. (1997). This branch was also reproduced in the simulations of Korres et al. (2000), who forced a Mediterranean Sea GCM with monthly mean NCEP winds and heat fluxes. On the other hand, the POEM data did not cover the southern part of the Ionian Sea, i.e. the area where the southern branch of the AIS propagates. South of the AIS three anticyclonic eddies developed down to a depth of 200 m, influencing the transport of surface MAW and LIW. The three Southern Ionian eddies were not present in the model solution before 1986 (not shown here). They form in 1987 and remain, with variable intensity, during Period II. Therefore, they are to be considered as part of the transition of the EMED circulation into the Period II. In 1987, we see a well-formed Mersa–Matruh Gyre, an extended cyclonic Rhodes gyre and a western Cretan gyre. The anticyclonic gyre system in the Shikmona area is in good agreement with POEM 1987 results (Robinson et al., 1991). The dynamic topography for October 1991 reveals significant changes in the EMED with respect to 1987. The northern branch of AIS in the Ionian Sea is stronger and the associated anticyclonic circulation extends much farther north in comparison to the 1987 case.

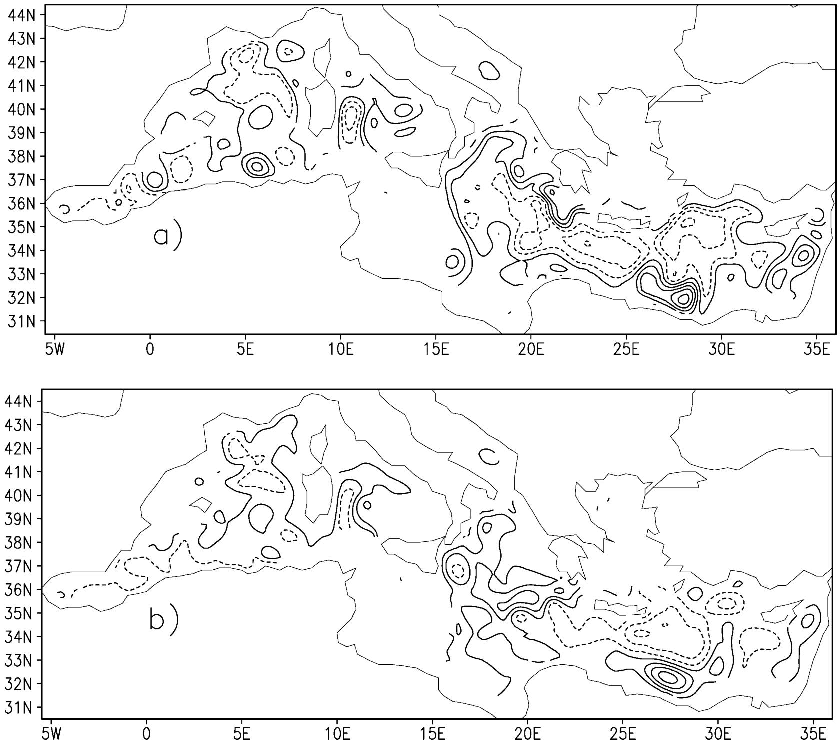

Fig. 11. Monthly mean dynamic height at 360 m with reference level 800 m. Upper panel: October 1987; bottom panel: October 1991. The contour interval is 1 cm.

The southern branch of the AIS is also stronger and the southern anticyclonic gyres have intensified with respect to 1987.

The model solution forms a structure of several anticyclonic eddies in the Levantine basin. It is connected to an enlargement of the anticyclonic Mersa–Matruh gyre northward. At about 29°E, the MMJ turns northward and reduces the size of the Rhodes Gyre to the east. This change was also discussed by Malanotte-Rizzoli et al. (1999) who showed that the major change in the EMED circulation in 1991 with respect to 1987 is the formation of three large anticyclonic gyres in the Levantine. As a result the LIW was blocked in the northeastern part of the basin and was diverted from propagating towards the Ionian Sea. Knowing that the positive wind stress curl was anomalously weak in 1990 and 1991 (Figs. 2 and 3), the model solution for October 1991 (Figs. 10b and 11b) produces the overall weakening of the cyclonic pattern in the Levantine Basin. In 1991 the cyclonic circulation was blocked by a system of large anticyclonic eddies in the Levantine. In the western part of the Cretan Passage, the cyclonic Cretan gyre, observed in 1987 both in the model solution (Figs. 10a and 11a) and the data (Robinson et al., 1991; Malanotte-Rizzoli et al., 1997) is not present in 1991 (Figs. 10b and 11b).

The salinity at 50-m depth in the EMED in 1987 and 1991 (Fig. 12) reveals the influence of the circulation on the salinity distribution. Beyond the Sicily Strait, MAW can be traced by its low salinity values in the southern Ionian Sea. In 1991 subsurface salinity minimum, east of Sicily occupies only the region of the Ionian Sea while its presence in the Levantine is much reduced. This result is consistent with the POEM data, which also found less MAW in this region in 1991 (Malanotte-Rizzoli et al., 1999). This change is related to the change in the transport of MAW by the northern and southern branches of the AIS. A larger amount of MAW is advected northward in the western Ionian Sea as well as trapped in the three anticyclonic eddies in the southern part of the

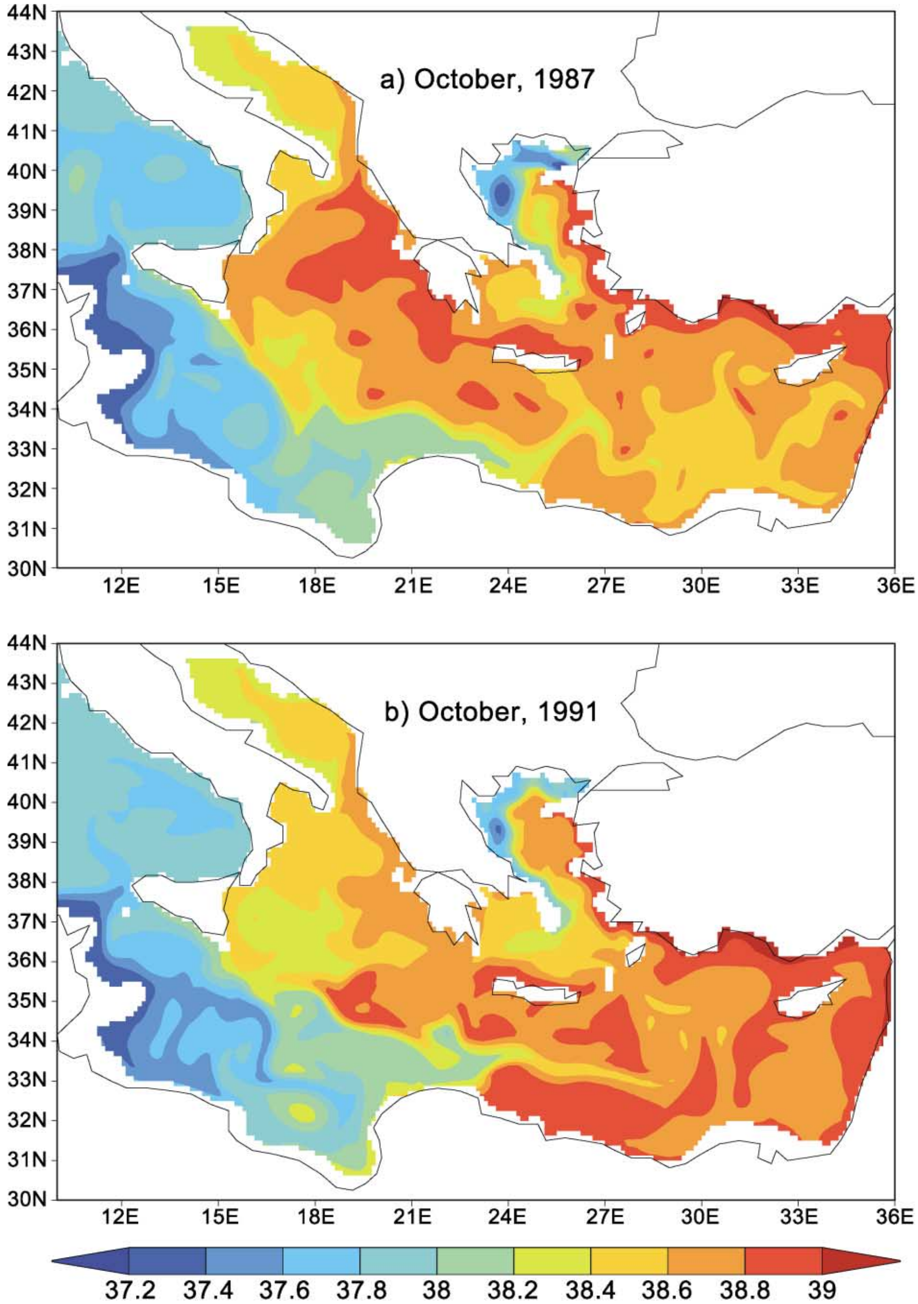

Fig. 12. Salinity distribution at 50 m. 15 October 1987 (upper panel) and 15 October 1991 (bottom panel).

basin. In 1991 the propagation of MAW eastward of this Ionian area is limited to a relatively small filament, much smaller than the flow of MAW in 1987. As a result, in the central part of the Levantine Basin, the salinity is 0.2–0.3 psu, higher in 1991 than in 1987.

In Fig. 13, the salinity and circulation pattern in the intermediate layers are shown. In 1987, a relatively

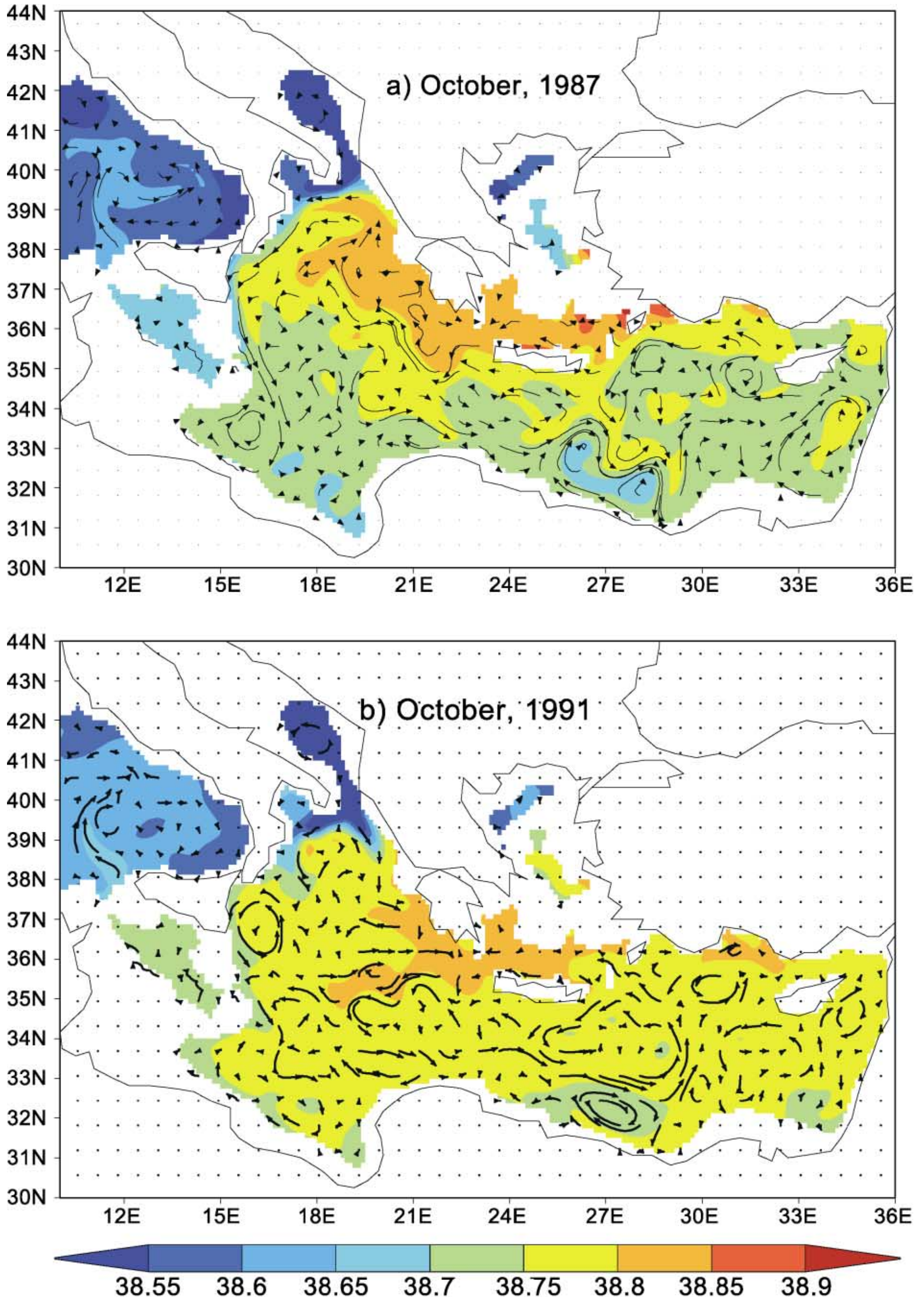

Fig. 13. Salinity and velocity (20 days particle trajectories) distributions at 360 m. 15 October 1987 (upper panel) and 15 October 1991 (bottom panel).

strong westward current originates from the area of the Rhodes Gyre meanders south of Cretan Island and then flows into the Ionian Sea. It transports LIW across the Cretan passage into the Ionian Sea. In 1987, LIW is present west of Crete and south of the Peloponnesus. In 1991 the transport through the

Cretan Passage decreased. The most important thing is that in 1991 the salinity increased almost everywhere in the Levantine Basin. As a consequence, surface (Fig. 12) and intermediate waters (Fig. 13) entering the Aegean Sea had higher salinity in 1991 than in 1987.

Two physical processes can be responsible for this change—deep convection, enhanced by the relatively cold winters after 1987 and the changes in the horizontal circulation, that modify the horizontal transport of MAW and LIW (Figs. 12 and 13). In an attempt to verify these physical mechanisms, we present the changes in salinity in different parts of EMED during Period II. The time series of the horizontally averaged model salinity over different regions of the EMED are presented in Fig. 14. The regions are the Southern Ionian (Fig. 14a), the southern Cretan passage (Fig. 14b), the Mersa–Matruh area (Fig. 14c), the Shikmona area (Fig. 14d), the Rhodes Gyre area (Fig. 14e) and the Northern Ionian Sea (Fig. 14f). They are indicated schematically in Fig. 1.

In the upper 200-m layer, there is a clear indication of a decreased amount of MAW in the Southern Cretan Passage, Mersa–Matruh, Shikmona and Rhodes Gyre areas starting from 1990. The subsurface salinity minimum increases almost everywhere in the whole Levantine basin and reaches the highest values in 1991. This overall decreasing of the influence of MAW in this layer of the Levantine Basin corresponds well to the weak wind cyclonic input in 1990 and 1991. In the regions of the Southern Cretan Passage and Mersa Matruh, the salinity of MAW remains weak to the end of the simulations.

In the Northern Ionian Sea the vertical salinity distribution shows a tendency towards a decrease of the vertical gradient due to the increase of the MAW volume and decrease of LIW. The decrease of LIW volume in the Northern Ionian is concomitant with the formation and growth of an intermediate layer of relatively high salinity (higher than 38.75 psu) in the southern Ionian and Levantine basin. This change starts to occur as early as 1988 in the Southern Ionian but reaches its maximum amplitude during 1990. The depth of this intermediate layer of high salinity, is higher than the maximum winter deep convection depth, suggesting that this layer is formed by the lateral transport of LIW from the Rhodes Gyre formation area (Fig. 14e) to all other regions.

The model solution thus suggests that the change of surface forcing after 1987 resulted in two major changes in the water mass transport in the EMED: firstly, the decreased surface MAW transport into the Levantine Basin and the increased MAW transport into the Northern Ionian Sea, and secondly, the weakening of LIW transport into the Northern Ionian Sea and increase of LIW transport in the Southern Ionian, the Southern Cretan Passage and the Mersa–Matruh areas. These changes in MAW and LIW redistribution produce an increase of the salinity in the surface and intermediate layers in the Levantine Basin. Correspondingly, the LIW approaching the Adriatic Sea in the Northern Ionian decreases. This preconditioning, due to the changes in the circulation, supported the increase in intensity of deep and intermediate water formation in the Levantine and Aegean Seas, where winters of Period II were exceptionally cold (Figs. 5b and 6c) and there was a decrease in the rate of EMDW formation in the Southern Adriatic.

We also have to mention that the increase of LIW into the areas of the southern Levantine favored the processes of local intermediate convection in the Mersa–Matruh Shikmona areas during the winters of 1992 and 1993.

As we have mentioned in the introduction, between 1987 and 1991, the EMED changed from a basin with a single source of deep water formation in the Adriatic to a basin with an additional source of deep waters in the Aegean. Observations (Della Vedova et al., 1998, submitted) and model results (Wu et al., 2000) indicate that the deep Aegean water outflow started much earlier than 1995 when Roether et al. (1996) documented it. Our model results confirm that the outflow started already in 1993, as we will show in what follows. In Fig. 15, we present a zonal salinity section in the Cretan Sea at 36°N latitude (see Fig. 1). The surface and intermediate layers are different in the two periods. While the water column average salinity seems to have decreased between 1987 and 1993 in the upper 200 m, salty waters are confined on the eastern most part of the Cretan Sea and the high salinity surface layer increased below 300 m. Furthermore, in 1993 the model represents a CIW outflow totally absent in 1987. It does not produce deep water outflow due to the weak surface forcing and model drift. The surface water flux being evaluated from monthly mean salinity climatology does not include

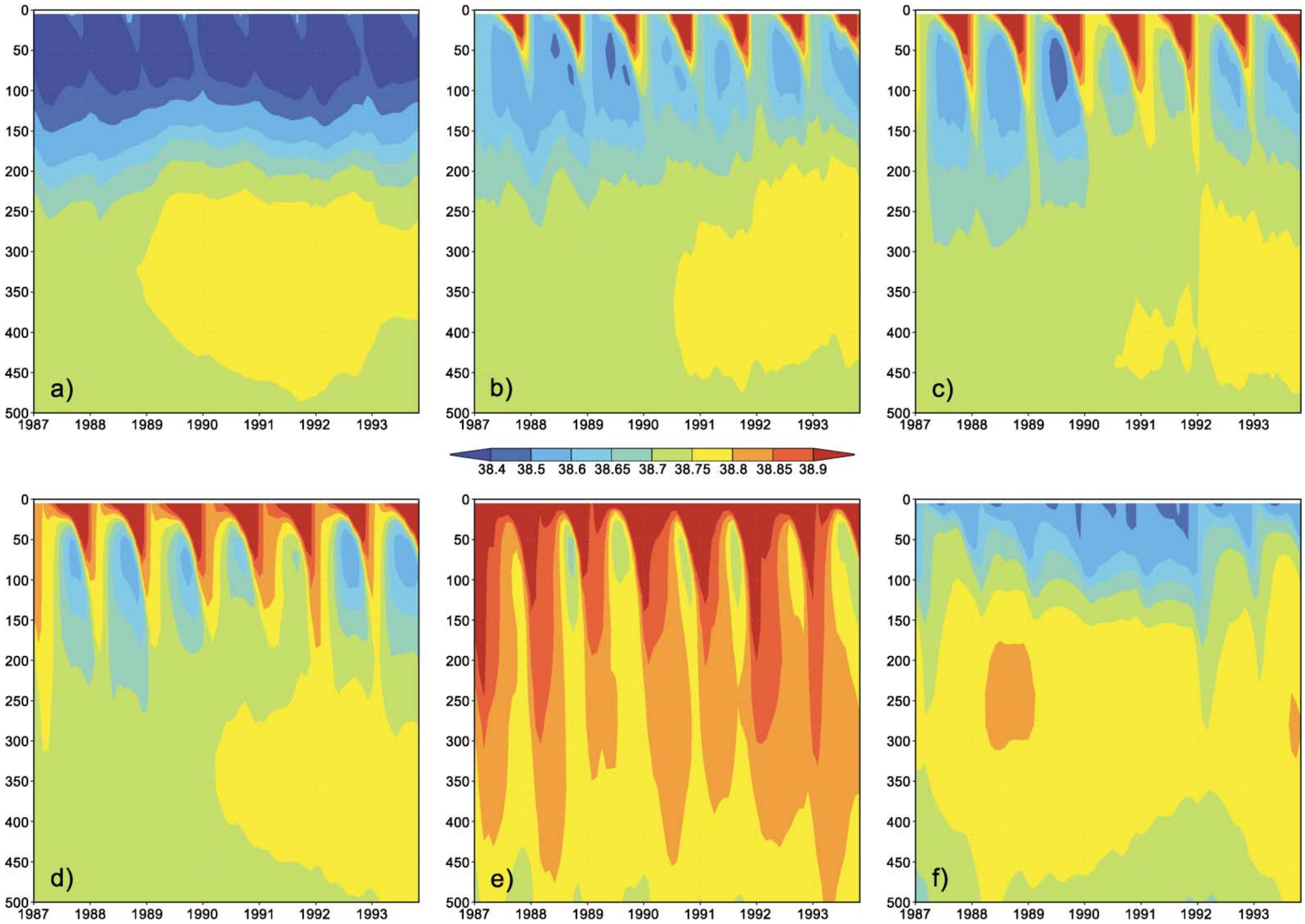

Fig. 14. Time evolution of horizontally averaged salinity in (a) Southern Ionian Sea; (b) Southern Cretan Passage; (c) Mersa–Matruh Area; (d) Shikmona area; (e) Rhodes Gyre area; (f) Northern Ionian Sea.

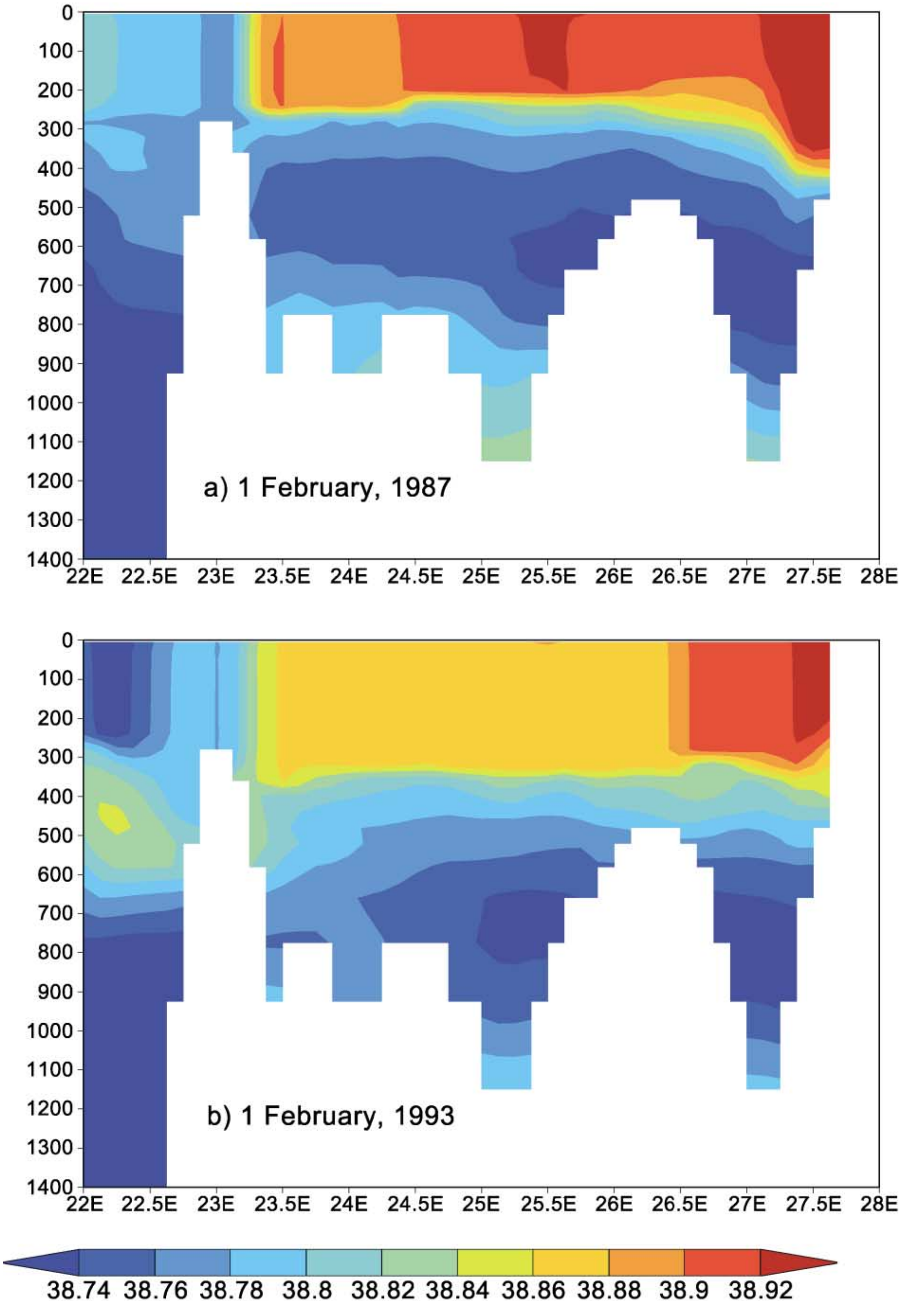

Fig. 15. Salinity cross-section across Cretan Sea. 1 February 1987 (upper panel) and 1 February 1993 (bottom panel).

the important influence of the precipitation and evaporation interannual variability. The model drift even relatively small (about 0.05 °C in the temperature) have an impact on the weak equilibrium between the CIW and EMDW (see Roether and Schlitzer, 1991). However, we claim that the model tries to mimic the

Aegean water outflow by means of CIW outflow. As we have shown before, from 1989 to 1993 the winters over Aegean Sea were anomalously cold (Fig. 6c). In addition LIW is found to be present in the central Aegean Sea (Fig. 13), while it was absent in 1987. The two concomitant factors, the cooling over the Aegean Sea and the overall change in the circulation and water mass structure (Fig. 13) caused the model to produce CIW outflow from the Aegean. Even though results are far from being satisfactory for this event, we claim that the change in the circulation between Period I and Period II in the Levantine and the concomitant cooling in the Aegean Sea could be the overall cause of the Aegean deep water outflow event.

4.2. Mean summer circulation of the Mediterranean sea during Period I and Period II

We would like now to document the dramatic changes in the circulation occurring at the level of mean currents for Periods I and II. The WMED upper layer (Fig. 16) mean circulation is dominated by two main current systems—the AC and the Ligurian–Provencal Current (LPC). LPC is a deep current dominated by wind-driven dynamics (Pinardi et al., 1985), which is present at all depths over the northern shelf and slope of WMED. While LPC weakens at all levels from Period I to Period II (Fig. 16), due to the overall weakening of the surface wind stress over WMED, the AC intensity does not show remarkable changes both in winter (not shown here) and summer (Fig. 16). The persistence in the strength of the AC can be explained by the relatively small interannual variability of the Atlantic inflow through Gibraltar that is a major driving force for the AC. However, the AC path changes from Period I to Period II, forming a stronger eastward current north of Balearic Islands during Period II.

The AC forms a mesoscale eddy in the western part of the Algerian Basin, which is present at all depths up to the bottom and has an horizontal extension of about 150 km between the coasts of Spain and Africa. The eddy position, intensity and type (cyclonic or anticyclonic) are variable in time. In the model solution during most of Period I, it is anticyclonic, while during Period II it is cyclonic, moderating in a different way the path of the AC in the westernmost part of Algerian Basin. The area on the southern flank of the AC is dominated by anticyclonic mesoscale eddy activity. The model solution also suggests that during the winter (not shown here), the anticyclonic eddy activity south of the AC is weaker than during the summer season. In general, the eddy kinetic energy tends to decrease between Period I and Period II in the whole WMED, due to the weakening of the wind energy input after 1987. The mesoscale variability and its energetics are studied in more details in a companion paper by Pinardi and Demirov (manuscript in preparation).

It is interesting to note that a quasi-permanent anticyclonic eddy is observed mainly at intermediate depths during Period I in the area northeast of Balearic Islands. This eddy, known from experimental data (Fuda et al., 2000) is called "WEDDY" and it is very weak and barely present during Period II.

At intermediate depths a meandering current intensified around the eddies is present. Experimental evidence from Wust (1961) and Ovchinikov (1966), as well as numerical results of Roussenov et al. (1995), Wu and Haines (1996), Korres et al. (2000) suggest that at intermediate depths there is a mean westward flow, which we can refer to as the Algerian Counter Current (ACC). The results from the coarse resolution simulations of Korres et al. (2000) suggest that the ACC is more developed during the summer season. Recent experimental results (Fuda et al., 2000) confirm that the ACC is not always present. Our results also suggest that the meandering ACC is not a permanent feature. Due to changes in surface forcing during Period II, the ACC is not present in the model solution while the main current transporting LIW into the Alboran Sea is south of the Balearic Islands (Fig. 16d). Careful inspection of this flow field shows that the LIW transported comes from the Gulf of Lions.

The AC enters the Tyrrhenian Sea through the Sardinia Channel and branches in two parts approaching the Strait of Sicily. The branching of the AC is related to the dynamics of a cyclonic eddy present southeastern of Sardinia. During Period I, it is located close to the relatively narrow Sardinia channel, moderating the inflow of LIW and of Tyrrhenian dense waters (Fuda et al., 2000) into the Algerian Basin (not shown). During Period II the intensity of the cyclonic gyre decreases remarkably during both seasons.

In the EMED, the two different mean flow fields are shown in Fig. 17. A major difference in the Ionian

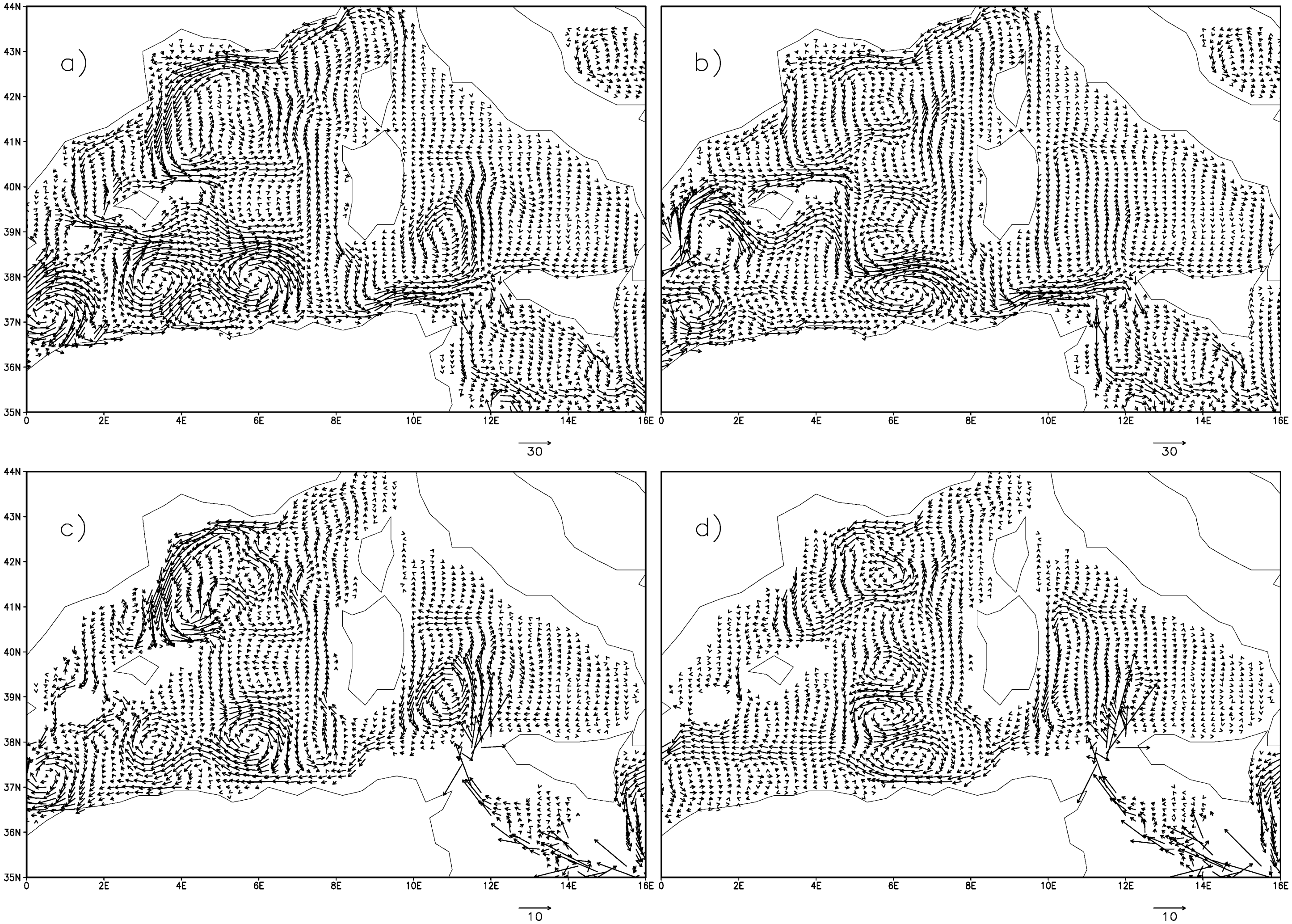

Fig. 16. Summer mean circulation in the Western Mediterranean Sea (a) at 30 m Period I; (b) at 30 m Period II; (c) at 360 m Period I; (d) at 360 m Period II.

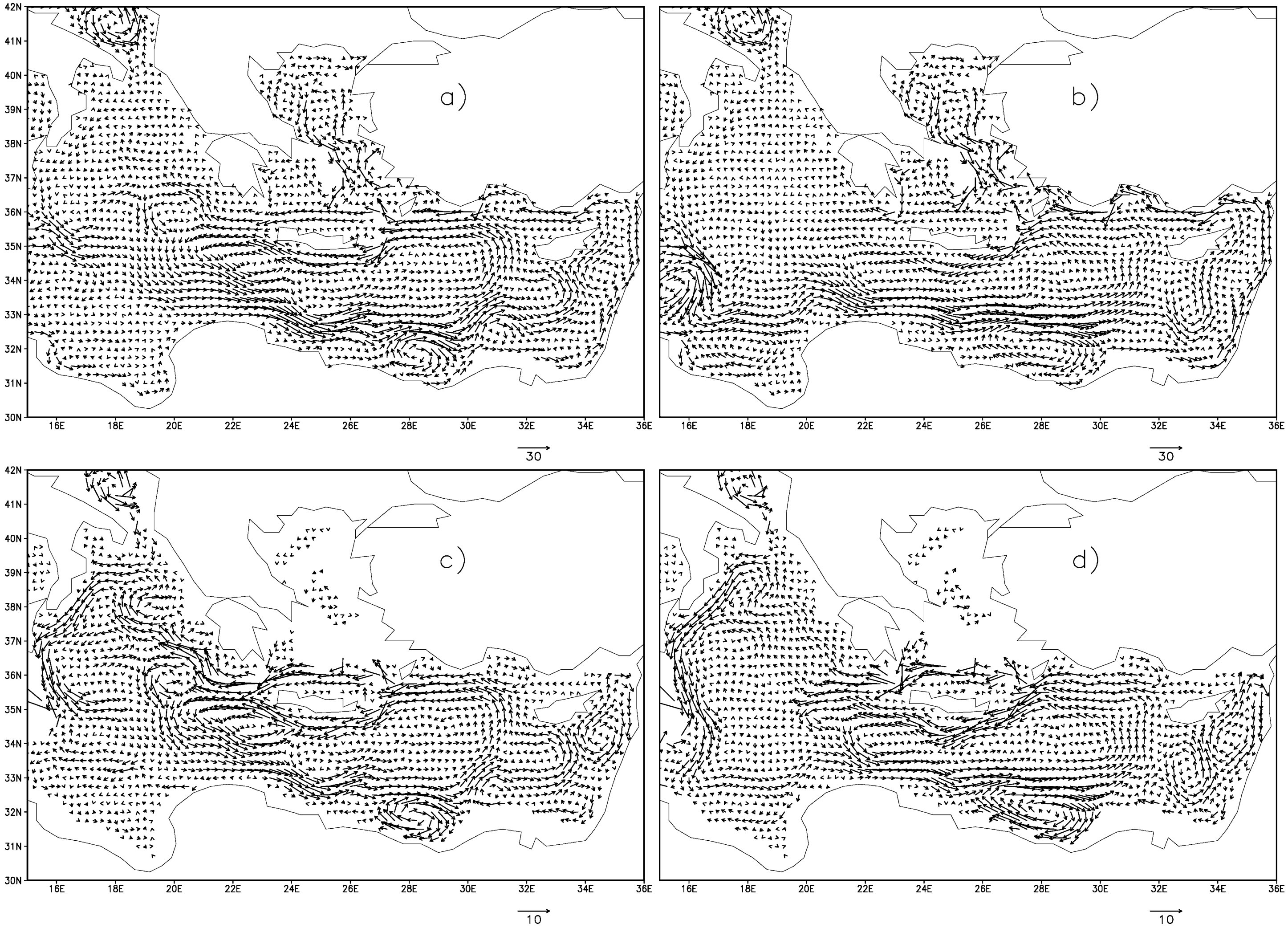

Fig. 17. Summer mean circulation in the Eastern Mediterranean Sea (a) at 30 m Period I; (b) at 30 m Period II; (c) at 360 m Period I; (d) at 360 m Period II.

Sea between Period I and Period II is the path of the AIS. During Period I, it is eastward towards the Levantine Basin. During Period II, AIS meanders around two quasi-stationary anticyclonic gyres along the African coast. As we discussed before, these anticyclones block the propagation of MAW eastward. They are present in the upper 200 m layer. At 360 m, the major difference between Period I and Period II is the weakening of the northwestward current field, which is replaced by a large recirculation of the westward current at the border of region (b) in Fig. 1. This is the current field modification, which produced an increase of high salinity waters in the Southern Cretan Passage and Southern Ionian regions shown in Fig. 14.

Another difference between Period I and Period II in the upper layer, is the increased AMC strength and inflow into the Aegean Sea during the Period II. The intensified AMC encircles all the Levantine basin and extends directly to the central and northern part of the Aegean Sea.

5. Discussion and conclusions

In this paper we show that the ECMWF reanalysis atmospheric forcing fields over the Mediterranean changed dramatically between 1987 and 1993, which can be attributed to long-term NH large-scale atmosphere circulation changes. Over Europe and the Mediterranean, these changes were produced by decadal scale changes related to the NAO. The importance of the NAO in producing Mediterranean Sea variability was discussed in a previous study by Send et al. (1999) with reference to precipitation anomalies. Here, we analyze wind and heat anomalies over the Mediterranean and connect them to the NAO variability.

The surface model forcing computed from ECMWF 1979–1993 reanalysis data should be subdivided in two periods: the first one extended from 1981 to 1987 and the second from 1988 to 1993. The main changes occur during the winter and they consist of: (a) a wind stress and wind stress curl intensity decreased over the whole WMED going from Period I to Period II; (b) wind stress curl and wind stress intensity increase over the Aegean and Levantine Seas; (c) a winter time cooling decrease over the Gulf of Lions and increased over the Aegean Sea and the Levantine Basin.

The experimental data of Lascaratos et al. (1999) indicated that the processes of increase of the Aegean Sea deep waters density in the end of 1980s and beginning of 1990s can be subdivided in two stages: firstly between 1987 and 1992 it was caused by increase in the salinity, and secondly after 1992 it was related mainly to the lowering the temperature. The second stage corresponds well to the relatively cold winters of 1992 and 1993 (see Lascaratos et al., 1999; Fig. 6c).

Two factors may be considered as major reasons for the increase in the salinity of Aegean Sea deep waters between 1987 and 1992. Firstly, as shown by Lascaratos et al. (1999), the surface water flux over the Aegean Sea changed after 1987 as a result of relatively high evaporation and low precipitation. This change in the surface water flux is considered by Lascaratos et al., (1999) as a major factor which influenced the salinity distribution in the MED after 1987 but is not present in our simulation. Due to the lack of good quality data for precipitation, the study of the water flux influence on the transition of the EMED remains beyond the scopes of the present work. The absence of water flux interannual variability in our forcing could be a partly controlling factor for the weak CIW outflow.

The second factor for the increase in the salinity of the Aegean deep waters is the change in the EMED circulation, driven by the atmospheric forcing variability. Our simulation showed that this change was driven by changes in the circulation of the whole MED between Period I and Period II. They are: (a) a general weakening of the WMED circulation; (b) a diversion of MAW in the Western Ionian Sea resulting in a subsequent large decrease of MAW in the Levantine basin; (c) an increased anticyclonic activity in the Southern Ionian basin and Levantine basin; (d) higher salinity at 360 m over the whole basin due to the changes in the transport of LIW from the Rhodes gyre to the Southeastern Levantine. While most of the known features from observations are reproduced in the simulation (comparisons were made with Malanotte-Rizzoli et al., 1999), the Aegean deep water outflow was severely underestimated by the model. However, we argue that the concomitant changes of circulation inducing different LIW distribution in the upper 400 m of the water column, and the anomalous cooling over the Aegean Sea are the likely cause of CIW and CSOW outflow. From our simulations we may say that the whole event, including the preconditioning that started in 1988–1989, make possible the outflow of deep and intermediate waters into the Ionian in 1991.

Our analysis shows that the changes in the surface forcing, which caused firstly the preconditioning of the EMED circulation and salinity distribution and secondly the intensification of deep water formation in the Aegean can be attributed to a large value of the NAO index after 1987. The 1987–1988 change in the state of the whole NH atmosphere was related to a relatively rare case of phase matching between NAO, PNA and EU (Watanabe and Nitta 1999). The present analysis suggests that the MED can be a sensitive area to such atmospheric regime changes.

Appendix A. Acronyms used in the text

MED	Mediterranean Sea
EMED	Eastern Mediterranean Sea
WMED	Western Mediterranean Sea
ASS	Atlantic Stream System
LPC	Ligurian–Provencal Current
AC	Algerian Current
ACC	Algerian Counter Current
AIS	Atlantic Ionian Stream
MMJ	Mid-Mediterranean Jet
AMC	Asia Minor Current
LIW	Levantine Intermediate Water
CIW	Cretan Intermediate Water
WMDW	Western Mediterranean Deep Water
EMDW	Eastern Mediterranean Deep Water
CSOW	Cretan Sea Outflow Water

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