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Decadal oscillations in the Mediterranean Sea: a result of the overturning circulation variability in the eastern basin?

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ABSTRACT: We studied the decadal variability of Mediterranean overturning circulation, induced by freshwater flux forcing in an eddy-permitting oceanic general circulation model (OGCM), using a statistical approach. Analysis of the modes of variability was carried out by means of the empirical orthogonal functional analysis (EOF) method. We demonstrated that the eastern basin is sensitive to switching from a restoring boundary condition on salinity to a freshwater flux. This study suggests that variability in the eastern basin drives the response of the whole Mediterranean Sea. We put forward the hypothesis that the eastern basin undergoes internal oscillations of bimodal amplitude.

KEYWORDS: Eastern Mediterranean Transient · Overturning circulation · Mediterranean Sea · Decadal oscillations

1. INTRODUCTION

Due to its relatively small dimensions, the Mediterranean Sea promptly responds to atmospheric variability (Demirov & Pinardi 2002, Stratford & Haines 2002). Indeed, variability in the production and transformation of water masses in the Mediterranean Sea has been mainly attributed to interannual changes in the strength of external atmospheric forcing (Pinardi & Navarra 1993, Mertens & Shott 1998, Malanotte-Rizzoli et al. 1999, Castellari et al. 2000). Nevertheless, theoretical and idealized as well as realistic modelling studies have shown that feedbacks between the rate of deep water formation and the strength of overturning circulation are potential sources of internal variability, independent of changes in atmospheric forcing (Welander 1982, Lenderink & Haarsma 1994, Yin & Sarachik 1995, Rahmstorf 1996). Such mechanisms of variability can be effectively studied inside the Mediterranean Sea, where processes are expected to occur on time scales that enable both close experimental monitoring and affordable numerical simulations.

The Mediterranean Sea is strongly density-stratified, mainly due to vertical gradients of salinity. It exchanges water with the Atlantic Ocean through the shallow and narrow Strait of Gibraltar, and is composed of 2 sub-basins of comparable size: the Western and Eastern Mediterranean Sea (WM and EM hereafter), separated by the Sicily Strait. Through the Strait of Gibraltar, the comparatively fresher superficial Atlantic inflow compensates evaporation and the outflow at depth of denser, saltier Mediterranean water into the Atlantic. The incoming Atlantic Water forms a surface water layer 100 to 200 m thick that flows eastward. The Atlantic water, moving eastward, is heated and mixes with the saltier surface Mediterranean water, forming the Modified Atlantic Water (MAW). In the Levantine Basin, Levantine Intermediate Water (LIW) of relatively high temperature and salinity is formed during winter. The LIW circulates cyclonically across the eastern and western Mediterranean sub-basins before reaching the Atlantic Ocean through the Strait of Gibraltar. The LIW is usually observed at a depth between 200 and 800 m. Deep water is produced

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at different locations in the Mediterranean Sea: in the Gulf of Lions (WM), in the South Adriatic and in the Levantine sub-basin (see Malanotte-Rizzoli et al. 1999).

Recently, the attention of the scientific community has been drawn to the observational characterization of Mediterranean circulation on decadal time scales (MEDAR Group 2002). The changes in the disposal of water masses that have occurred in the EM since the beginning of the last decade (1990s) have been documented by several studies based on field observations (Roether et al. 1996, Klein et al. 1999, Lascaratos et al. 1999 among others). Observations conducted over the last decade demonstrate that water of Aegean origin, which used to be observed at intermediate depth, substituted Adriatic Deep Water (ADW), filling the deep layers in the Ionian and Levantine basin. This bottom water of recent formation is warmer and saltier than the 'old' bottom water of Adriatic origin, which was lifted by several hundred meters (Roether et al. 1996). Consequently, the eastern deep layers became warmer and saltier while the intermediate layers, affected by the lifting of the old deep water, become fresher, colder, relatively poor in oxygen content and rich in nutrients. This event is know as the Eastern Mediterranean Transient (EMT, Theocharis et al. 1999, Klein et al. 2000). The change in the vertical distribution of water masses was accompanied by a change in the dispersal path of the relatively fresher MAW and of the relatively saltier and warmer LIW and Cretan Intermediate Water (CIW). The modification of surface and intermediate circulation caused the Aegean Sea to become a favourite site for the production of dense water, owing to its augmented salinity and to the contemporary decrease in the inflow of salty intermediate water into the Adriatic (e.g. Malanotte-Rizzoli et al. 1999).

Several studies have addressed the relative influence of altered meteorological conditions on the observed changes in circulation and vertical stratification in the EM (Roether et al. 1996, Samuel & Haines 1999, Theocharis et al. 1999, Wu et al. 2000, Demirov & Pinardi 2002, Josey et al. 2003, Rupolo et al. 2003a). One open issue is whether the EMT is to be considered 'unique' or whether it is connected to some internal variability of the EM. Indeed, by 1961 Wüst had analysed historical data of deep and bottom water in the EM and claimed that "it seems probable that some smaller influences come from the Aegean Sea by occasional overflow through the channels between Crete and Rhodes. But because of the small number of observations, the conditions of this overflow cannot yet be sufficiently examined". Forty years later and basing their analysis on the Medatlas database, Theocharis et al. (2002) documented great variability in the characteristics of Mediterranean water masses during the last century. In particular, there were at least 2 distinct events characterized by an increase in both salinity and temperature in the Ionian, Cretan and Levantine basins, one in the early 1970s and the other from the mid to late 1990s. The second episode is clearly related to the EMT and its evolution has been monitored up to now. As to the first event, less data are available, and we can only conjecture on its appearance/disappearance mechanisms.

In the WM, there is evidence that the newly formed deep water has exhibited strong variability since 1959 (e.g. Lacombe et al. 1981, 1985). Moreover, some studies claim that there is a positive trend in deep water temperature and salinity (Bethoux et al. 1990, Leaman & Schott 1991, Rohling & Bryden 1992). The contemporary observed changes in dense water formation and in deep water characteristics of the Ionian Sea and Levantine Basin (Roether et al. 1996) have provided further evidence for the existence of a trend in heat and freshwater budgets over the entire Mediterranean. The question now is to understand if the observed variability is anthropic (e.g. the construction of the Aswan High Dam on the Nile River, see Boscolo & Bryden 2001), or if it is at least partly due to the natural variability of the thermohaline circulation.

There is definitely a demand for and explanation of the dominant feedbacks inside the ocean, as both observations and model simulations have revealed the importance of the role of saline water at intermediate depth in regulating deep water formation (Wu & Haines 1996, Bethoux & Gentili 1999, Rupolo et al. 2003b). Although air-sea interaction remains the principal driving force of Mediterranean circulation, purely oceanic processes might prove to be crucial in determining what equilibrium will be reached.

This study aims to explore the natural variability of the Mediterranean circulation on decadal time scales, and of the basin's response to fixed atmospheric forcing, in view of the fact that deep and intermediate water formation through convection is one of the key processes taking place in this basin. We compare the 2 different circulations that result from running an oceanic general circulation model (OGCM) under different surface forcing, and look for integral variables that can provide synthetic information on the mean state of the basin.

This study is organized as follows: Section 2 describes the numerical model and the experiments we conducted. Results are presented in Section 3: a first analysis of the variability observed in the model simulation is presented in Section 3.1; in Section 3.2 we propose a tentative characterization of mean circulation and of its variability by decomposing it in empirical orthogonal functions (EOF); in Section 3.3 we examine the possible relevance of our results in connection with

episodic phenomena such as the EMT. Finally, in Section 4 we summarize our results and outline future developments.

2. MODELLING THE MEDITERRANEAN THERMOHALINE CIRCULATION

2.1. Model configuration

The Geophysical Fluid Dynamics Laboratory Modular Ocean Model (MOM) (Pacanowski et al. 1991, Roussenov et al. 1995, Wu & Haines 1996) was adjusted to properly simulate the Mediterranean circulation, and a realistic representation of coast lines and topography was included. A rigid lid approximation was applied at the surface. We chose a horizontal grid spacing of 0.25° × 0.25°, 19 vertical levels (see Table 1) and a time step of 1 h as the best compromise between the spatial resolution necessary for the correct simulation of the Mediterranean circulation and a time step long enough to allow climatological integrations. The non-linear UNESCO equation for density was used. The Gibraltar Strait is open (two velocity grid points) and the model domain extended into the Atlantic to 13° E, with a buffer zone in the western part, where temperature and salinity were relaxed at all levels to the Levitus (1982) climatology with a time scale of 5 d.

Dissipative terms in the momentum equations were parameterized through the biharmonic operator in the horizontal and the Laplacian operator in the vertical, with constant values for eddy viscosity, A_h and A_v of 3×10^{18} cm^4 s^{-2} and 1.5 cm^2 s^{-1}, respectively.

Table 1. Depth, thickness and vertical diffusivity coefficient (K_v) for each model level

Level	Depth (m)	Thickness (m)	K_v (cm^2 s^{-1})
1	5	10	3.0
2	15	10	1.2
3	30	20	0.3
4	50	20	0.05
5	80	40	0.02
6	120	40	0.01
7	160	40	0.01
8	200	40	0.01
9	240	40	0.01
10	280	40	0.01
11	340	80	0.01
12	420	80	0.01
13	500	80	0.01
14	620	160	0.01
15	850	300	0.01
16	1250	500	0.01
17	1750	500	0.01
18	2250	500	0.01
19	2750	500	–

Vertical diffusion in the tracer equations was represented by a Laplacian operator with a vertical diffusivity profile similar to that used by Wu & Haines (1998). In order to mimic the physics of the mixed layer, rather high values (from 3 to 1 cm^2 s^{-1} in the first 80 m) were assumed in the upper layers, which rapidly decreased to very low values for depths greater than 100 m (0.01 cm^2 s^{-1}, see Table 1).

In order to improve the sub-grid mesoscale field description and to better represent fluxes through the straits, many authors (e.g. Wu & Haines 1996) adopted a variable horizontal eddy diffusivity K_h, which is higher in the straits than in the open ocean. In this study, a space and time dependence in the horizontal eddy diffusivity was introduced by using the parameterization proposed by Babiano et al. (1987), which relates the long-time diffusion coefficient K_h to the values of the enstrophy (Z) and eddy kinetic energy (E) by means of $K_h \propto E \times Z^{-0.5}$.

The values of eddy kinetic energy and enstrophy were computed by taking into account the time and space eulerian decorrelation lengths, typically 10 d and 100 km. The diffusivity field was computed and updated every 10 d. It shows large space and time variability, attaining maximum values at the straits. The model sensitivity to such a parameterisation was studied by Rupolo et al. (2003b), who showed that it leads to significant improvements in the representation of frontal structures and deep water formation.

2.2. Air-sea surface fluxes

2.2.1. Restoring and mixed boundary conditions

Pioneering attempts in constructing OGCMs followed practical considerations in the choice of surface boundary conditions. Because the use of heat, freshwater and momentum fluxes from climatology led to very unrealistic thermocline structure, generally due to the scarcity of information included in climatological data sets, are usually prescribed surface values of temperature and salinity rather than surface fluxes. Although very effective in reproducing experimental observations, and therefore today's climate, this choice is often regarded as arbitrary.

The relaxation to the climatological temperature, as parameterization for air-sea heat flux, is justified with regard to Haney (1971). His formulation is based on a set of empirically justified assumptions on the properties of the heat flux H, which were inserted in a Taylor expansion for $H \times T_0$, where T_0 is surface temperature.

Although lacking any physical ground, restoring boundary condition for salinity is still widely used in practice with the aim of obtaining realistic salinity

fields. Moreover, both evaporation and precipitation are impossible to parameterise in terms of oceanic variables. In particular, the only oceanic variable that explicitly appears in the expression for evaporation is surface temperature and not surface salinity itself, while precipitation is a very complicated function of atmospheric processes.

The lack of physical grounds for the relaxation of salinity is evidenced by the inability of ocean models to develop multiple equilibria under such a boundary condition, which on the contrary may be expected due to the non-linear nature of the governing equations. In this view, relaxation appears to reduce the degrees of freedom of the system.

An effective improvement was introduced when Bryan (1986) designed a technique in which restoring boundary conditions were applied both to temperature and salinity, until the ocean model reached an equilibrium state where salt conservation was achieved. At that point, the equivalent mean salt flux S_f, necessary for maintaining the salinity balance, can be computed as

$$S_f = \beta(S^* - \overline{S}) \quad (1)$$

where β is the relaxation time scale, S^* is the employed climatology and $\overline{S}$ is monthly surface salinity averaged over decadal time scales.

The model-diagnosed surface salt flux was then used to force the integration of the model from that point onwards, keeping a restoring boundary condition for temperature. The salt flux computed according to Eq. (1) can be converted into an equivalent freshwater flux by

$$E - P = \frac{S_f}{\overline{S}} \Delta z_1 \quad (2)$$

where Δz_1 is the thickness of the first model layer (10 m in our case).

Such a combination of 'Haney'-type temperature boundary condition and fixed equivalent salinity flux has become known as 'mixed' boundary condition.

2.2.2. Surface momentum and heat fluxes

The presence of high-frequency variability in surface forcing was proven to be a key feature for a correct representation of the water mass formation process in OGCM studies of the Mediterranean Sea (Artale et al. 2002). To retain the high frequency variability of the heat and momentum surface forcing, we use simultaneous daily mean values of European Centre mid-range weather forecast (ECMWF) winds and satellite derived sea surface temperature (SST) of a particular year, iterated perpetually.

The surface temperature was relaxed in the whole domain to daily values of 1988 SST from satellite data (D'Ortenzio et al. 2000) using a wind dependent relaxation time: the stronger the wind, the smaller is the surface temperature relaxation time, which ranges from 2 d to 2 h for winds exceeding 10 m s^{-1}. The resulting mean time scale slightly exceeds 1 d: small values (O[1 h]) are confined in the region of the Mistral wind, while in the rest of the domain assumes rather constant values (O[1 d]); see Artale et al. [2002] for a full account of such parameterization). The year 1988 was arbitrarily chosen because of the closeness of SST to climatological values during this time (e.g. see Marullo et al. 1999). However, a negative bias of about 1°C was added to the local field in the Adriatic Sea in February in order to get a realistic water mass formation rate.

ECMWF daily mean surface wind fields of the same year (1988) were mapped on the model grid and then used to compute wind stress and temperature relaxation time.

2.3. Numerical experiments: Expt 1 and Expt 2

Following Bryan's (1986) technique, we carried out 2 experiments that differed in the formulation of the surface boundary condition for salinity. In Expt 1, restoring boundary conditions were applied both to temperature and salinity.

The model was run for 200 yr with the surface momentum and heat fluxes described in Section 2.2.2. Surface salinity was relaxed to monthly climatological data from the Mediterranean Ocean Database (MODB) (Brasseur et al. 1996). At the end of Expt 1 the virtual salinity flux was diagnosed according to Eq. (1), where $\overline{S}$ was taken as the monthly surface salinity field averaged over the last 10 yr of simulation. In Fig. 1 we show the annual mean of the diagnosed equivalent surface freshwater fluxes $E - P$ (Eq. 2), whose spatial distribution and spatial average (83 cm yr^{-1}) are very similar to the climatology and to those obtained in Myers & Haines (2000) and Artale et al. (2002).

In Expt 2, we switched to mixed surface boundary conditions by prescribing the virtual salt flux diagnosed in Expt 1 and maintaining the same momentum and heat fluxes as in Expt 1. The model was run for 300 yr.

When analyzing the results, we skip the first 100 years (spin-up) of both Expts 1 and 2. Consequently, in the following we refer to Years 101 to 200 as Expt 1, and Years 301 to 500 as Expt 2. To summarize the model integration performed in this study, we present in Fig. 2 the time evolution of salinity (depth 280 m) in the Aegean Sea. The sudden jump in salinity marks the switch to mixed boundary condition.

3. RESULTS

3.1. Detection of interannual variability

During the numerical experiments we stored the average vertical profiles of temperature and salinity at selected locations (Fig. 1) every 5 d. These sites were chosen so as to spot deep water formation in the eastern basin and to trace LIW advection into the Western Basin. In particular, Sites 2 and 3 indicate the areas of the Adriatic dense outflow through the Otranto Strait and the area of dense water formation in the South Aegean Sea (Cretan Sea), respectively. Sites 4 and 5 were monitored in order to investigate the role of the Sicily Channel in determining mass, heat and salt exchanges across sub-basins.

In Fig. 2 (salinity at Site 3), it is evident that after switching to mixed surface forcing the model undergoes an adjustment process, at the end of which a new dynamical equilibrium is achieved that is different in terms of mean value, amplitude of oscillation and time scales of variability in hydrological properties. Under restoring (Expt 1) the model does not exhibit oscillations on scales longer than a few years. The oscillation has an amplitude well within the noise induced by variability on shorter scales. On the contrary, under mixed surface forcing conditions (Expt 2) oscillations over decadal time scales are excited. Spectral analysis confirms that in Expt 2, much more energy is involved in processes over decadal time scales than it is in Expt 1 (Fig. 3). A similar spectral peak was observed by Myers & Haines (2000). However, they used flux boundary conditions for both temperature and salinity and the spectral peak corresponding to interannual variability was shifted to slightly lower frequencies. The time-evolution of temperature at the same site and the corresponding spectral analysis lead to similar results.

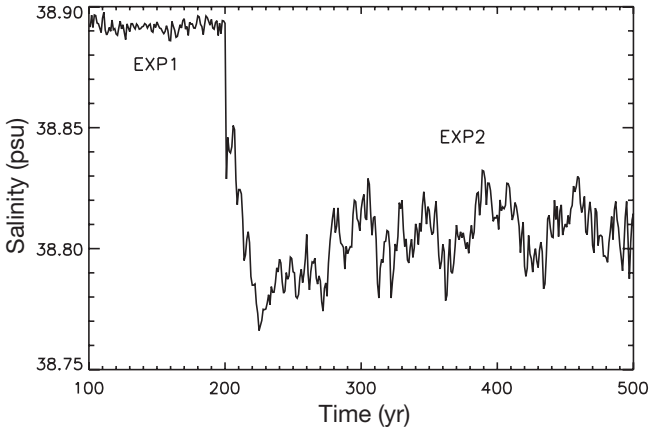

Fig. 2. Time evolution of salinity at 280 m depth in the Aegean Sea (Site 3). The first 100 years of spin-up are omitted. Fast Fourier Transform (FFT) was separately applied to time series of Expts 1 and 2, and the signal truncated to eliminate components with periods <1 yr

This effect impacts on the characteristics of the water that is advected from the eastern to the western sub-basin; Figs. 4 & 5 show the variability of the salinity profile at the entrance of the Sicily Channel (Site 4) and west of Sardinia (Site 6), both in Expts 1 and 2. It is evident that the intermediate water flowing across the Sicily Channel also exhibits a great interannual variability in Expt 2 that is not present in Expt 1, and that this 'oscillating' behaviour must be held as a robust characteristic of western Mediterranean intermediate water in Expt 2. Moreover, a great interannual variability in the strength of the convection is also observed in the Gulf of Lions (Fig. 6), where we report for both experiments the time series of the convection depth, which is defined as the maximum depth z at which $\rho(z) < \rho(0) + 0.01$, where $z = 0$ is the sea surface and $\rho(z)$ is the density obtained from mean profiles of temperature and salinity from Site 7. Note that this underestimate can be considered only as a proxy (lower bound) of the maximum convection depth simulated by the model, since the vertical profiles are averaged over a box (4 × 4 grid points) that is larger than the area where the simulated convection takes place. Deep

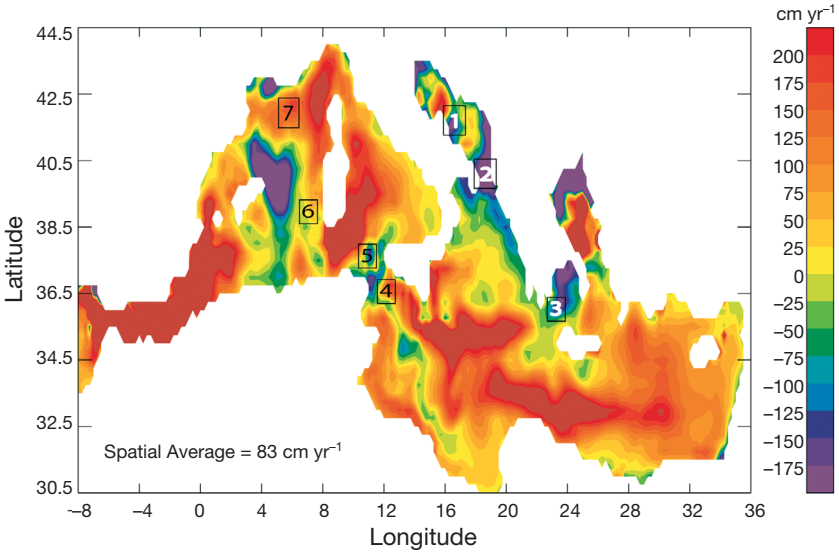

Fig. 1. Expt 1. Mean surface freshwater fluxes diagnosed over Years 190 to 200. Spatial average was computed over the area enclosed in the Mediterranean Sea. Numbers refer to sites (discussed in Section 3.1) where vertical profiles of temperature and salinity were collected every 5 d

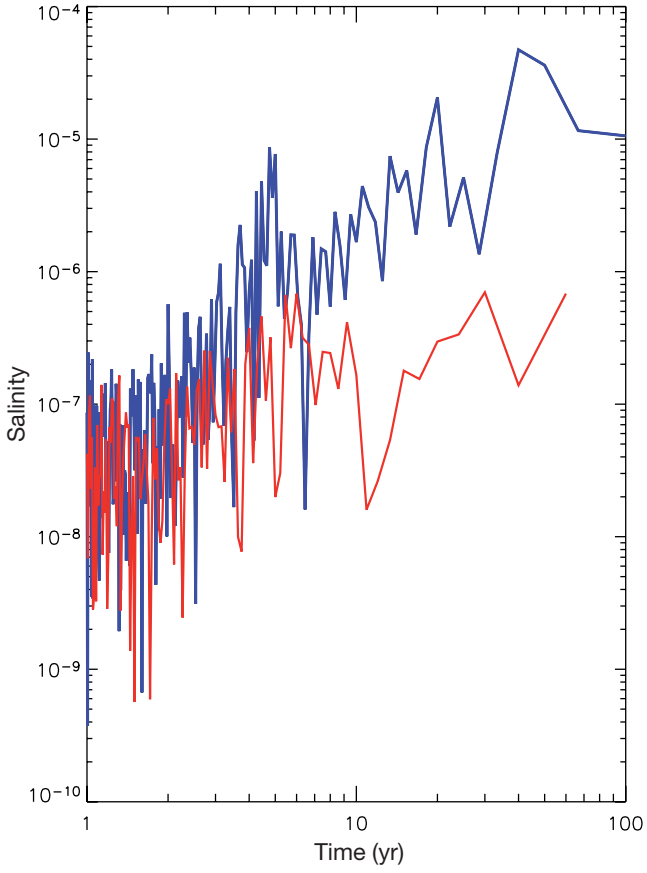

Fig. 3. Spectra of evolution of salinity in the Aegean Sea (280 m) for Expt 1 (red) and Expt 2 (blue)

convection events occur in February–March for both Expts 1 and 2, but the latter experiment shows a definite greater interannual variability in convection depth, even if we did not find a clear correlation time lag between the deepest convection events of Expt 2 and the salty impulse in the Sicily and Sardinia Channel shown in Figs. 4 & 5.

The results of Expt 2 raise 2 main questions: (1) Is it possible to detect an increased natural variability in the dynamics at basin scales, and to assess its impact on events that are mainly a consequence of occasional atmospheric forcing (i.e. the EMT)? (2) Does the LIW transported across the Sicily Channel, carrying with it the interannual variability acquired in the eastern basin, play a role in enhancing mixing processes in the Gulf of Lions, or are these solely a response to the change in surface forcing? In the present study, we mainly focus our attention on the first question.

3.2. Dynamical characterization of sub-basin circulation

In order to separately characterize the dynamic behaviour of the 2 Mediterranean sub-basins, we attempted a representation in the phase space defined by kinetic energy and surface heat flux. Figs. 7 & 8 provide evidence that the EM undergoes a seasonal cycle in both experiments, whereas the Western Basin does not. The main difference between the 2 sub-basins is that the EM hosts the principal processes of deep and intermediate water formation, and therefore acts as a transformer of potential to kinetic energy. The WM does not allow the same synthetic dynamical description. This difference in behaviour is probably a consequence of the unique properties of each basin. For example, the morphology of the WM does not allow the circulation to organize itself on basin scales. Therefore, the exchange of energy with the atmosphere affects mainly oceanographic processes at a local scale (i.e. convection in the Gulf of Lions). In contrast, in the eastern basin, overflows from the Adriatic and Aegean Sea corresponding to water mass formation processes imply a reorganization of the circulation at a basin scale.

The Mediterranean as a whole appears as a superposition of the 2 sub-basins, with the eastern basin functioning as the engine that sustains the overall circulation.

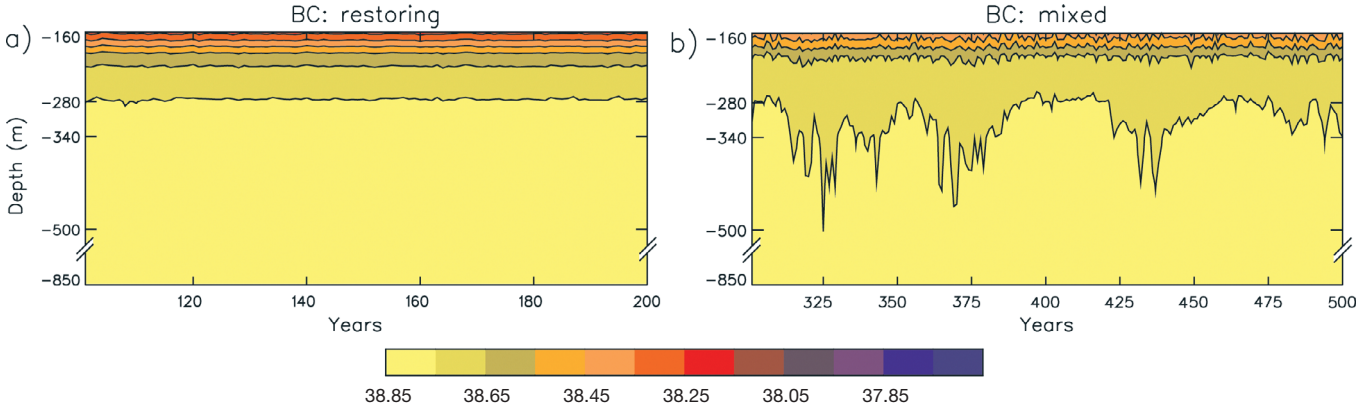

Fig. 4. Time evolution of salinity profile (psu) in the Sicily Channel (Site 4, Fig. 1) for (a) Expt 1 and (b) Expt 2. BC: boundary condition

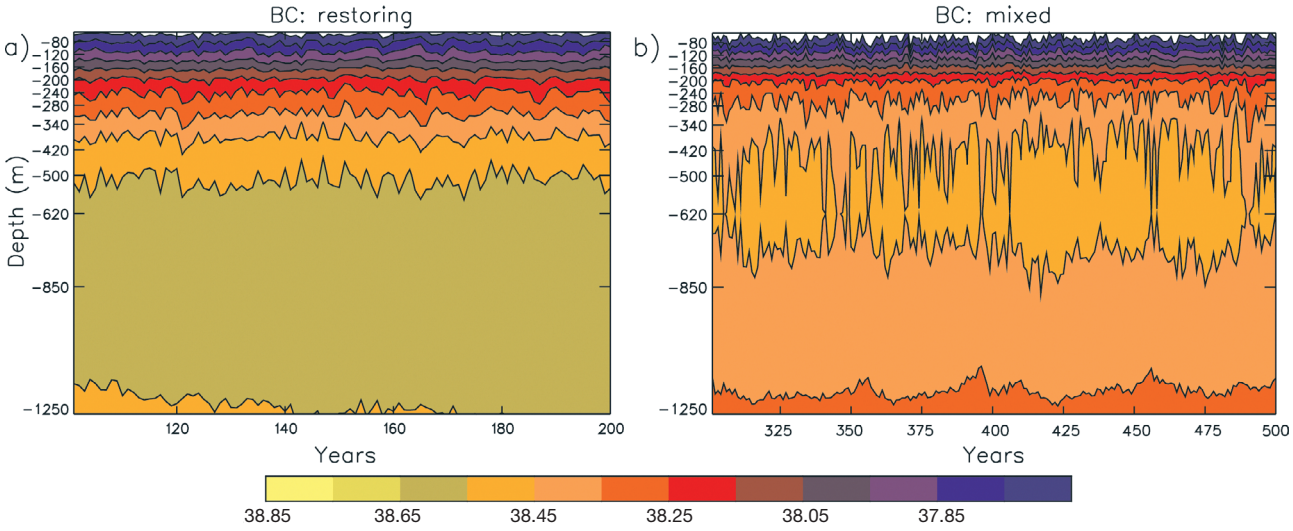

Fig. 5. Time evolution of salinity profile (psu) west of Sardinia (Site 6, Fig. 1) for (a) Expt 1 and (b) Expt 2

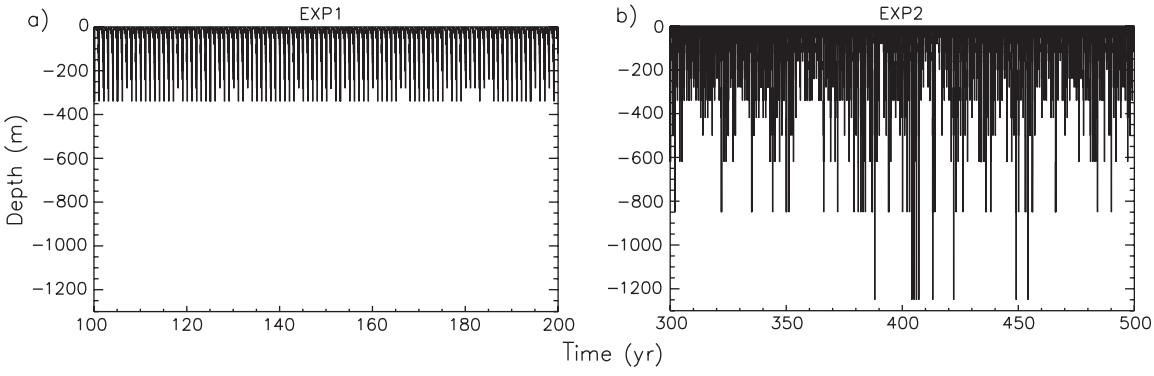

Fig. 6. Convection depth (see definition in Section 3.1) in the Gulf of Lions as a function of time for (a) Expt 1 and (b) Expt 2

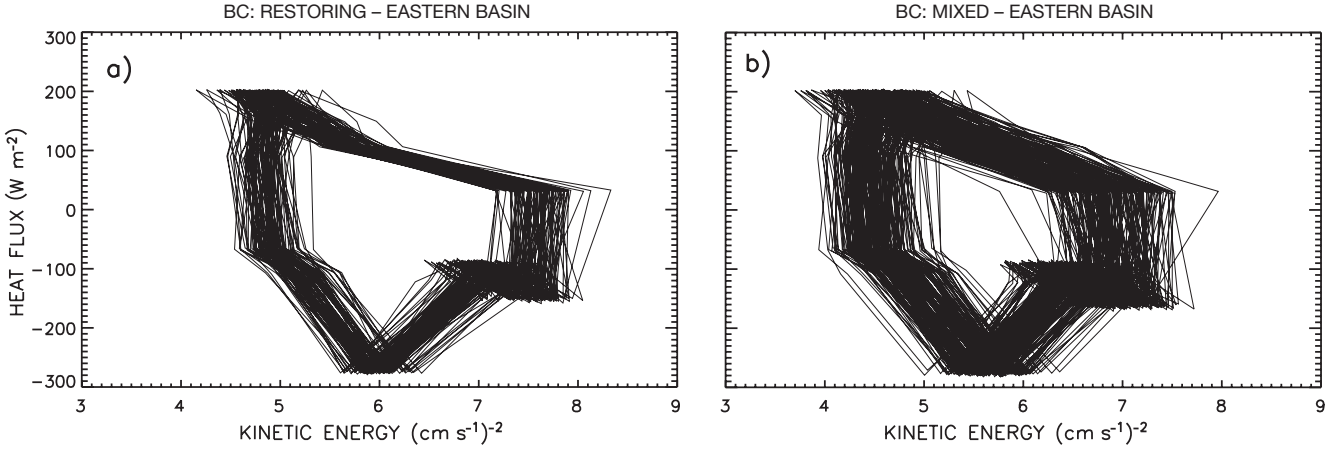

Fig. 7. Phase diagram of kinetic energy against heat fluxes for the eastern basin: (a) restoring and (b) flux boundary condition

The major dynamic response to the change in surface forcing appears to be the increased variability evident in Expt 2, which is consistent with the oscillation detected in the time series of temperature and salinity. In order to verify if this variability can be tracked at basin scales, we chose zonal and meridional transports (stream functions) as the variables that might give the best synthetic information on the dominant dynamics (see Figs. 10 & 11).

The averages of monthly means over the whole duration of Expts 1 and 2 were calculated. Fig. 9 shows

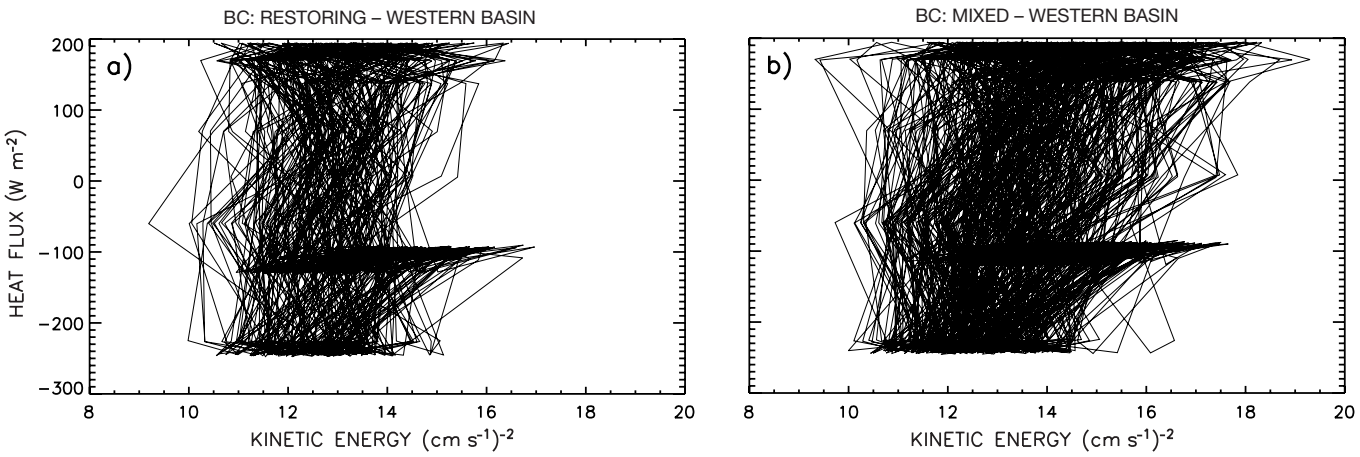

Fig. 8. Phase diagram of kinetic energy against heat fluxes for the western basin: (a) restoring and (b) flux boundary condition

the mean global zonal transport fields for the period from 15 June to 14 July in the case of restoring (Fig. 9a) and mixed (Fig. 9b) boundary conditions. This is the month for which the 2 experiments exhibit the greatest differences. In this selected extreme example, differences at intermediate and surface levels can be appreciated in the EM, whereas the structure of circulation in the WM does not appear to be affected. In particular, in the EM we observed a weakening of the intermediate cell, accompanied by an intensified inflow of MAW at the surface. When meridional transport is considered, the mean circulation in the western basin does not appear to be significantly modified by the change in boundary conditions (figure not shown). Instead, the situation changes substantially when we examine the EM, where the simulated mean meridional circulation appears to be definitely weakened at the intermediate levels during the whole period starting approximately in July and up to the end of December (Fig. 10).

We attempted a decomposition in EOFs to identify the main patterns of variability. The interannual variability in dynamics was then captured by the time evolution of the principal components (PCs). In order to filter the seasonal cycle out, anomalies were computed with respect to the long term average monthly means. The first 2 EOFs of the Mediterranean zonal transport account for 56% of the variance in Expt 1 and 49% of that in Expt 2. The comparison between first modes in Expts 1 and 2 provide evidence for differences in the East at intermediate levels, together with a slight reduction in the explained variance.

In the WM the 2 main modes account for 70% of the variance, and appear to be similar in terms both of structure and of statistical relevance. Therefore, the EOFs fail to capture the increased variability of the circulation observed in the phase diagram (Fig. 8), which might be a consequence of oscillations in the LIW properties detected in local time series. For this

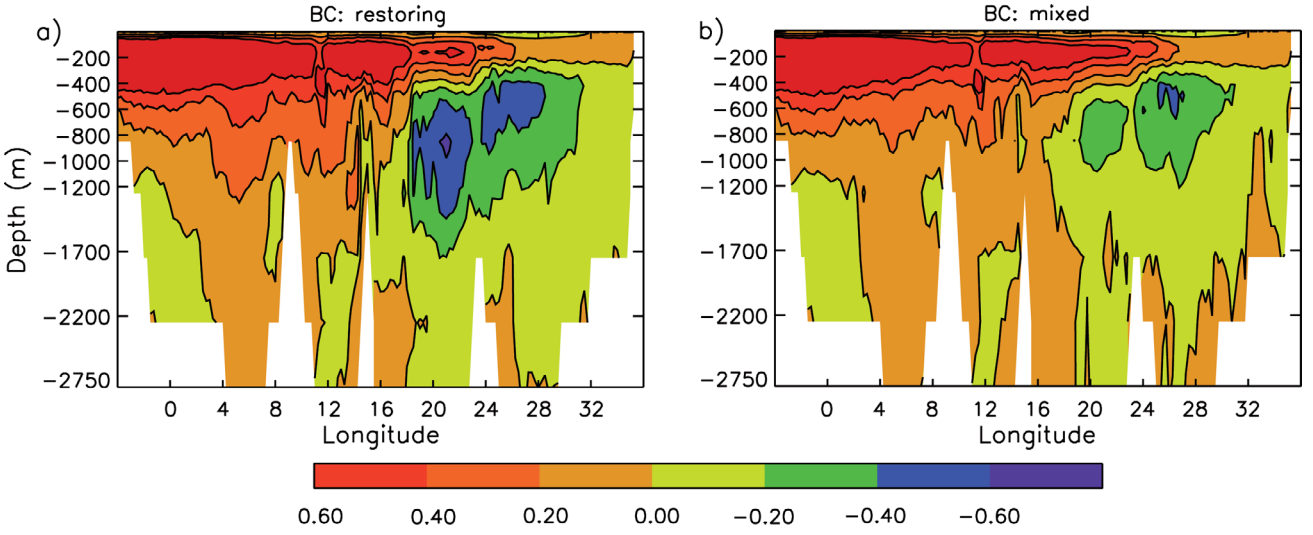

Fig. 9. Mean zonal transport field (Sv) from 15 June to 14 July: (a) Expt 1 and (b) Expt 2. Negative values correspond to clockwise circulation

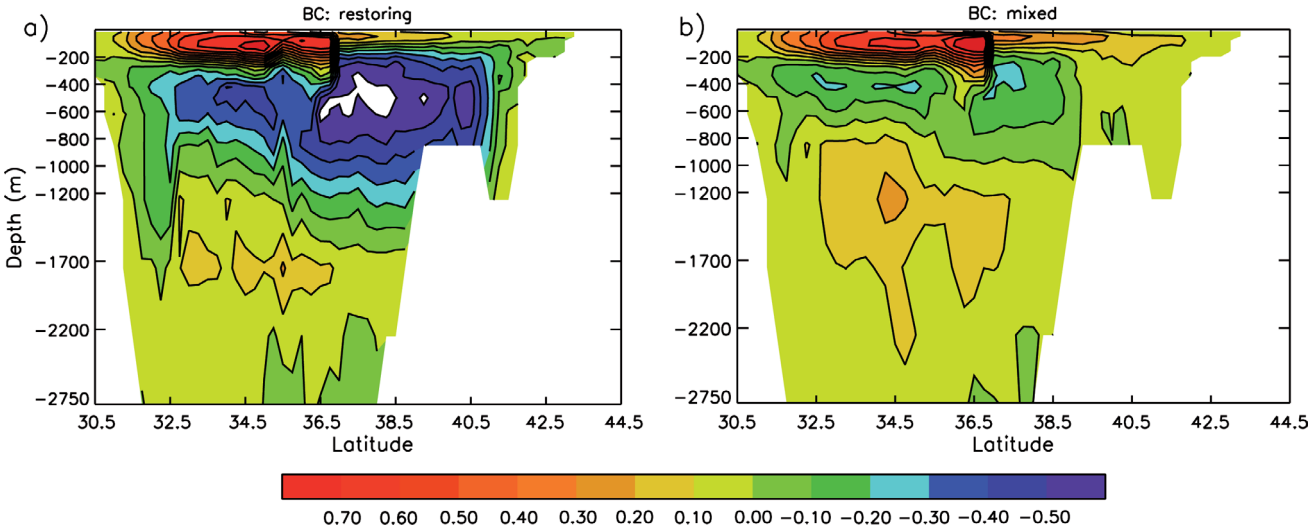

Fig. 10. Mean meridional transport field (Sv) in the eastern basin from 15 July to 14 August: (a) Expt 1 and (b) Expt 2. Negative values correspond to clockwise circulation

reason, we considered the EOF analysis to be irrelevant for the WM and it is not shown. Such inadequacy might be expected in view of the processes that, in the WM, are liable to be affected by LIW variability. In particular, winter convection in the Gulf of Lions is a good instance of the gap that separates the dynamics of the WM, being local and instantaneous if compared to the spatial and temporal scales of the eastern circulation. As the EOF analysis that we applied only serves to isolate coherent correlation structures over time scales longer than 1 yr, its shortcomings are not surprising.

In contrast, the EOF analysis proves to be remarkably effective in capturing circulation variability in the EM: not only do the first 2 modes change structure, but the amount of explained variance is significantly higher in the case of Expt 2, up to 58% compared with 37% of Expt 1. In particular, the statistical relevance of the first mode has almost doubled (46% in Expt 2, 27% in Expt 1, Fig. 11). Moreover, the dominant cell intensifies in Expt 2, and its centre is dislocated deeper and northward. With regard to the second mode (Fig. 12), it is interesting to note its qualitatively different structure between Expts 1 and 2. In Expt 1 a strong and deep positive lobe located at the latitude of Otranto Strait, and therefore related to the overflow of the ADW, coexists with a rather weak southern negative lobe, whose core is located at intermediate depth. In Expt 2 a positive lobe is located inside the Adriatic Sea and extends southward in the Ionian Basin, while a negative lobe appears stronger and deeper, and has shifted northward when compared with that in Expt 1. A possible interpretation of this result is that in Expt 1, the variability

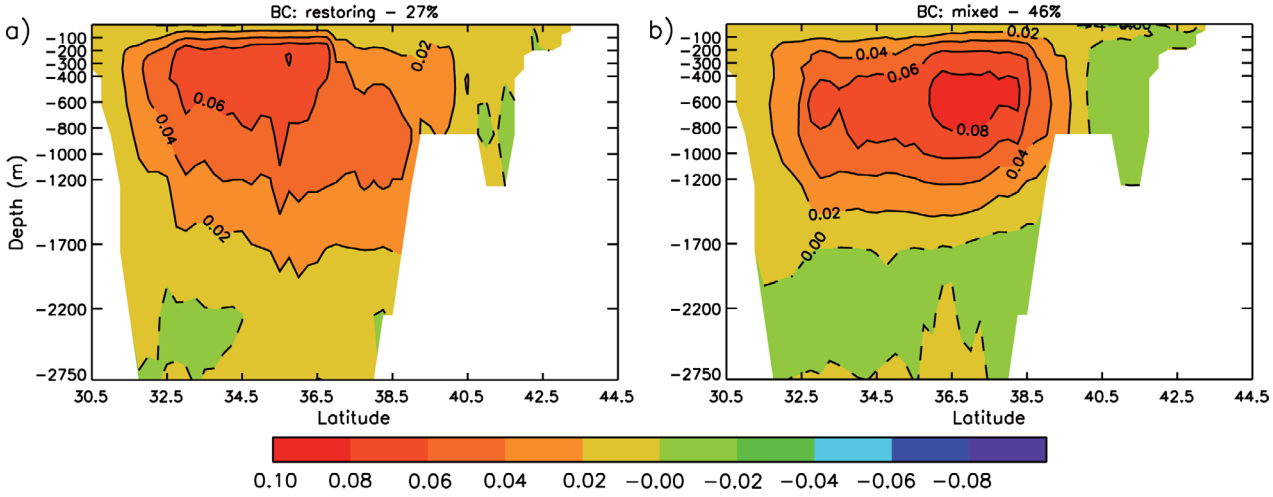

Fig. 11. Meridional transport (Sv) in the eastern basin: first mode of empirical orthogonal function decomposition for (a) Expt 1 and (b) Expt 2

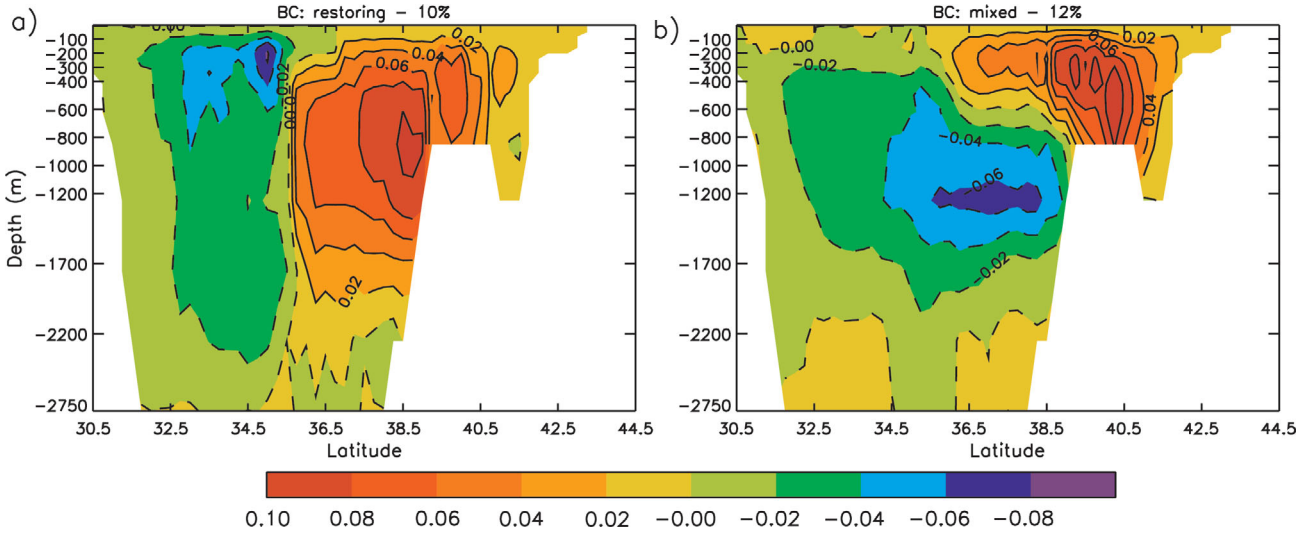

Fig. 12. Meridional transport (Sv) in the eastern basin: second mode of EOF decomposition for (a) Expt 1 and (b) Expt 2

of the meridional overturning cell (Fig. 10) is directly related to the variability of deep water formation in the Adriatic Sea. In Expt 2, the Adriatic Sea is no longer the main contributor to the variability of the overturning cell, and the latter appears to be affected by processes that are located elsewhere in the EM.

The time evolution of the correspondent PCs (Fig. 13, PC 1) shows that in Expt 2, variability has shifted towards longer time scales than in Expt 1, consistent with the spectra of local values of temperature and salinity already shown (Fig. 3). Such a shift is not detectable in the PCs relative to the western basin.

3.3. Variability of mean circulation in the eastern basin

Focusing our attention on the eastern basin, a question arises as to whether the augmented variability in mean transport is the signature of weakening/strengthening of a circulation that does not change structure, or is stabilizes around different equilibria. We consider the time series of monthly mean meridional transport fields in the eastern basin, for the period from 14 June to 15 July. Remarkable differences are observed at intermediate levels (e.g. Years 351 and 359) in the case of mixed boundary conditions (Expt 2, Fig. 14). In the following, we refer to the 2 different circulation patterns as Pattern A (Fig. 14a, Year 351, strong intermediate meridional cell) and Pattern B (Fig. 14b, Year 359, weak meridional cell). In contrast, in the case of restoring boundary conditions, the circulation is found to be quite stable and shows no great differences from year to year, and Pattern A is a persistent feature of meridional circulation.

In order to give a rough indication of the time scales of the oscillation between the 2 patterns, we plotted

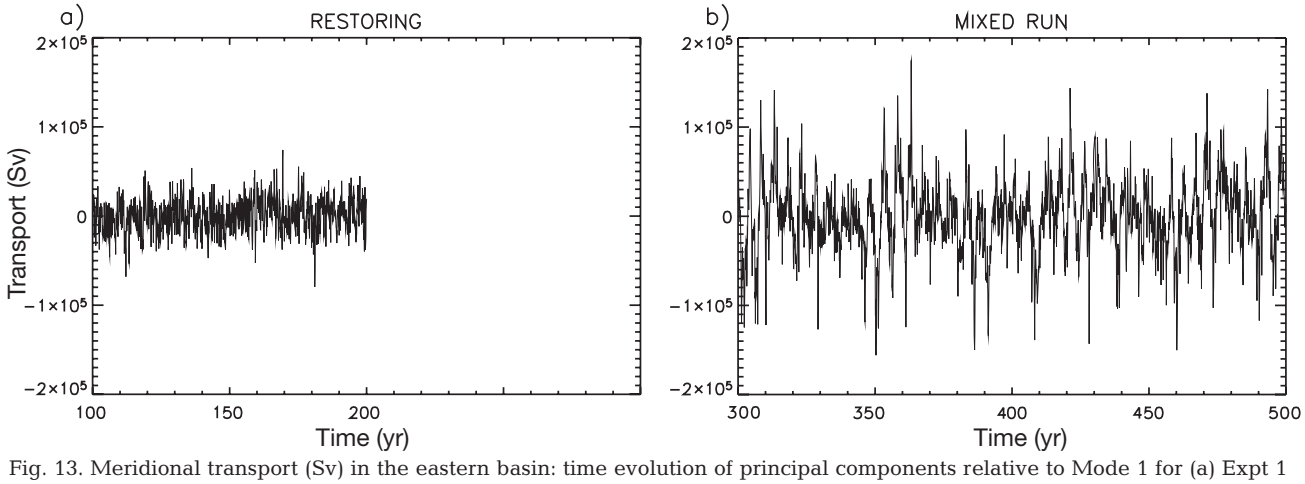

Fig. 13. Meridional transport (Sv) in the eastern basin: time evolution of principal components relative to Mode 1 for (a) Expt 1 and (b) Expt 2

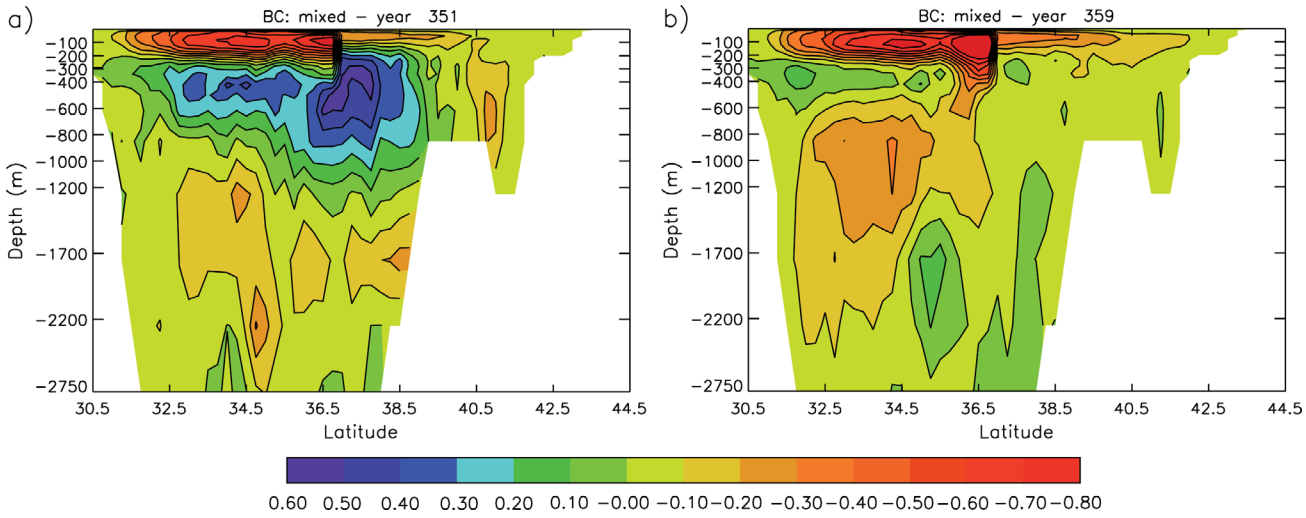

Fig. 14. Expt 2. Monthly mean meridional transport (Sv) in the eastern basin from 15 July to 14 August in (a) Year 351 and (b) Year 359. Negative values correspond to clockwise circulation

the time evolution of the maximum of the meridional streamfunction in the area of the intermediate level cell (Fig. 15), representing Years 351 and 359 with filled circles. Two features are apparent: one is the weakening/strengthening of the overturning over time scales of a few years, the other is the cyclic variation of the amplitude of this oscillation with a period of a few decades. Furthermore, the meridional and zonal overturning are correlated, as shown in Fig. 16. With the aim of providing a synthetic picture of the amplitude variability, we constructed the probability density function (PDF) of the differences between subsequent relative extremes, which is suggestive of a bimodal distribution (Fig. 17). Due to the shortness of the time series, no robust statistical test could be performed on the significance of the PDF. A longer experiment is perfectly feasible, but in the present

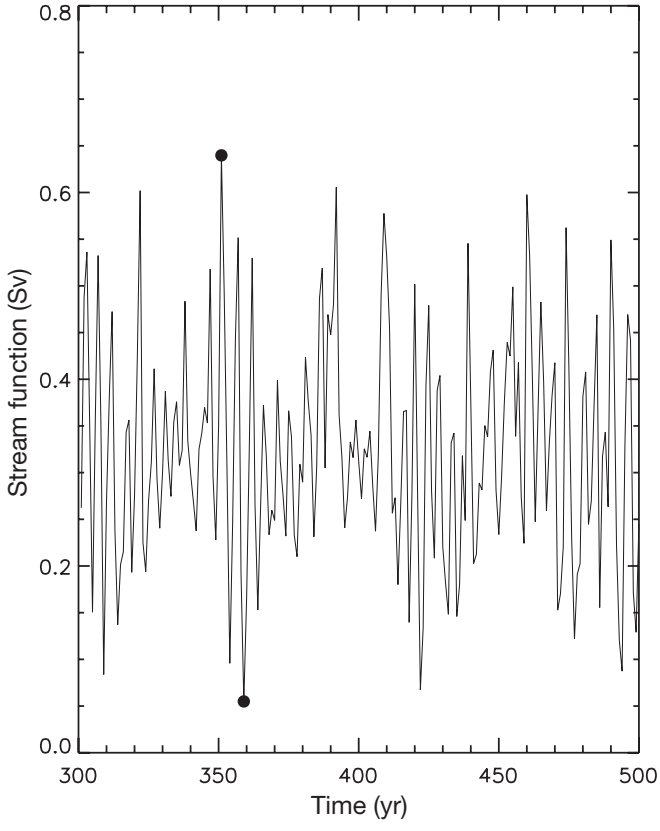

Fig. 15. Expt 2 (July). Evolution of maximum meridional streamfunction (Sv) in the Eastern Mediterranean Sea. ●: Years 351 and 359

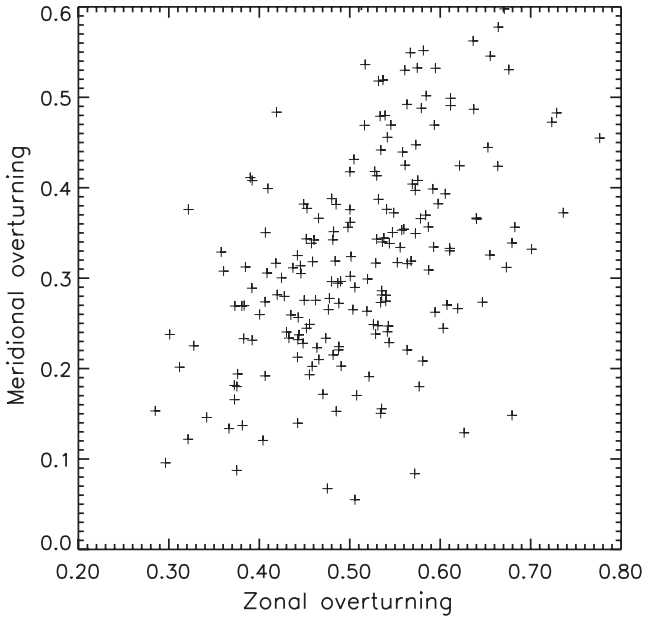

Fig. 16. Scatter diagram of meridional versus zonal overturning. Correlation coefficient r = 0.5

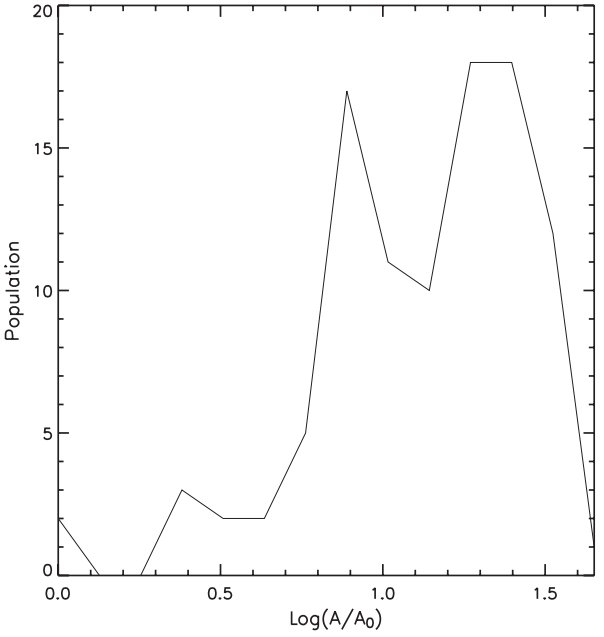

Fig. 17. Probability density function of peak-to-peak amplitude of the oscillation of maximum meridional overturning (Fig. 15). Amplitude A is normalized with a reference amplitude A_0 = 0.01 Sv. Natural logarithm of ratio A/A_0 is considered in order to account for the increase in relative spacing of single peak-to-peak amplitudes in the case of large oscillations

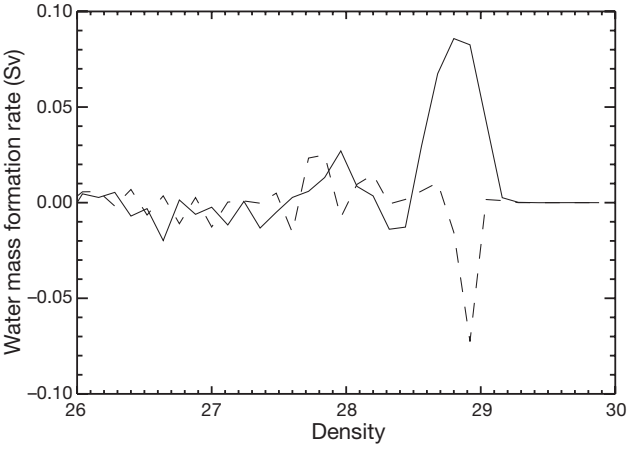

Fig. 18. Difference in yearly water mass formation rate (expressed in Sv as a function of density anomaly) between Years 359 and 351 (corresponding to strong and weak meridional overturning, respectively, in the Eastern Mediterranean Sea). Positive values are indicative of greater formation rates in the year characterized by strong meridional overturning. – – – –: Aegean Sea; ———: Adriatic and north Ionian Seas

study the integration was not designed to capture variability on scales longer than 10 yr. However, when changing the amplitude of the bins used to build the PDF, the bimodal character is substantially unchanged (not shown).

In view of the interest which recently arose regarding the EMT (see Section 1), we attempted to verify if the oscillation in meridional and zonal transport could be related to the time evolution of water mass formation rate in the eastern basins.

Following Walin (1982), the 1 yr averaged water formation rate per density class, due to buoyancy forcing in the surface layer, $F(\rho')$ is given by:

$$F(\rho') = \frac{1}{\rho'\Delta T}\int \mathrm{d}t \int \mathrm{d}x\mathrm{d}a \left[\frac{\alpha}{C_\mathrm{p}} H(x,y,t) - \rho(x,y,t)\beta \cdot S(x,y,t)\cdot Q(x,y,t)\right]\cdot\delta[\rho(x,y,t)-\rho']$$

where ΔT (= 1 yr) is the total time of integration, $H(x,y,t)$ and $Q(x,y,t)$ are the heat and freshwater surface fluxes, C_p is the specific heat capacity of water, $S(x,y,t)$ and $\rho(x,y,t)$ are the surface salinity and density, and α and β are the derivatives of density with respect to temperature and salinity.

In Fig. 18 we show the difference of yearly water mass formation rate between Years 359 and 351 in the Aegean Seas and in the Adriatic and north Ionian Sea. In this plot a positive value is indicative of a larger water mass formation rate. Therefore, the year characterized by strong meridional overturning is characterized by a stronger formation rate in the Adriatic and north Ionian Seas, which substantially compensates for the reduced water mass formation in the Aegean Sea (dashed line is negative). A parallel inspection of the surface salinity field (Fig. 19) reveals that dense water in the Adriatic Sea is associated with an increase of the surface salinity there. A similar mechanism was discussed in a similar model simulation by Myers & Haines (2000). Because surface flux boundary conditions are applied to the salinity equation in Expt 2, the observed salinity difference can be attributed to advective process. This occurrence is likely to be attributed to the interactions between deep water formation process and advective process in the EM.

We computed the density differences between the water at the sill level in the Aegean Sea and the Otranto outflow of the Adriatic Dense Water (ADW, Sites 3 and 2). Keeping in mind that Expt 1 is characterized by both a stable cell at intermediate levels and by a denser outflow of ADW, we checked whether a correspondence could be found between negative values of the density difference (DD = density of water at the sill level in the Aegean – density of the ADW outflow at the Otranto Strait) and the presence of such a cell. We found that in Expt 1, mean DD is negative and always associated with Pattern A, while in Expt 2 it oscillates around a positive value, and is generally associated with Pattern B or with a meridional cell that is weaker than that of Pattern A. Moreover, the ampli-

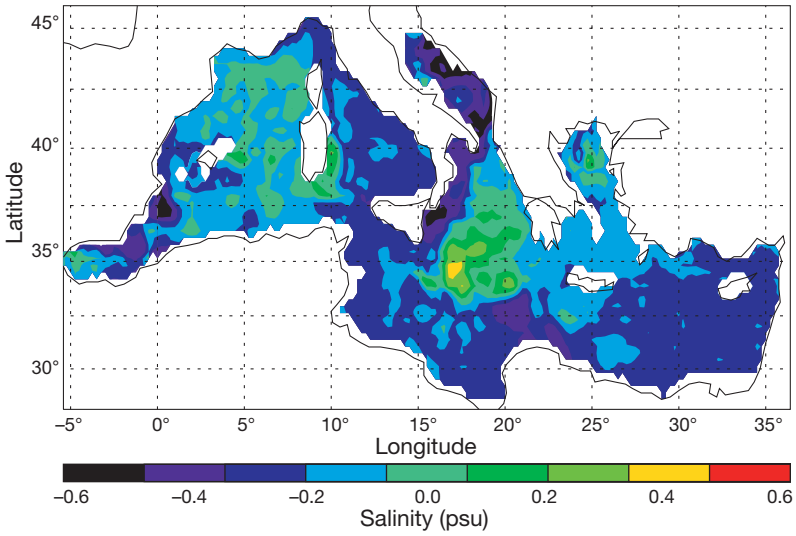

Fig. 19. Salinity difference at surface between Years 351 and 359. Negative values are indicative of saltier water in the year characterized by strong meridional overturning. The same map computed using salt content in the first 100 m (not shown here) shows the same qualitative behaviour

tude of the variability of DD is much greater in Expt 2 than in Expt 1 (figure not shown).

We can observe that switching to mixed boundary conditions only slightly enhances convective processes in the Aegean Sea, and that the overall predominance of dense water of Aegean origin in Expt 2 is due to a smaller formation rate in the Adriatic that in our model is crudely represented. Even so, we think that our results are reliable in the framework of a sensitivity study. It is to be noted that the foregoing results enforce the interpretation of the second EOF of meridional transport in the EM.

Although we are aware that the statistical relevance of such considerations should be enforced by a close analysis of the 3D dynamics, we think that the potential relevance of natural oscillations in determining phenomena such as the EMT deserves attention. From such a perspective, the role of the atmospheric forcing would not be diminished, but it would need to be reconsidered in relation to the state of the basin, for which it necessarily has an impulsive connotation. In particular, it may be argued that if the Aegean and Adriatic Waters had already altered their relative weight due to natural variability, the forcing exerted by the atmosphere would produce an effect that could otherwise be damped out.

The impact of the competing role of the Adriatic and Aegean Seas on the variability and hydrological characteristics of the water masses exported from the eastern to the western basin (see Figs. 4 & 5) is still to be fully understood, and will constitute one direction for future development.

4. SUMMARY, CONCLUSION AND FUTURE DEVELOPMENTS

Our work aims to provide a first characterization of the natural variability of the Mediterranean Sea under an atmospheric forcing that only exhibits a seasonal cycle, through the analysis of a long OGCM simulation.

We found that the Mediterranean Basin does exhibit an internal variability over scales of climatological interest, provided the prescription of boundary conditions that allow the circulation to departure from the equilibrium imposed by restoring to climatology, therefore inducing the generation of convective-advective feedbacks.

In the context of the well-established phenomenological and experimental investigation of the Mediterranean Circulation, our work proposes a description of the dynamic response to surface forcing that attempts to capture the main functional differences between the 2 Mediterranean sub-basins. We looked for synthetic, integral indicators of the state of the circulation that may serve as a guide for more analytical studies of the 3D fields of physical variables. We found that the 2 sub-basins can be qualified in terms of their different representation in phase space. The eastern basin shows an extremely well defined seasonal cycle, and allows a coherent description in terms of integral quantities, probably as a consequence of a circulation that is organized over larger space and time scales in comparison with the western basin. Both sub-basins exhibit a more pronounced variability when mixed boundary conditions are applied. Trace of this variability is lost in the western basin when zonal and meridional stream functions are considered, while the eastern basin definitely oscillates between 2 different patterns (weakening/strengthening of THC), which could be qualitatively related to the density difference of water masses formed in the Adriatic and in the Aegean Seas. The verification of such a result calls for close analysis of the 3D dynamical fields, as statistics cannot constitute a reliable guide in relating the temporal and spatial scales spanned by local and basin scale processes, unless new suitable instruments are developed. The interpretation of the EMT as a purely atmosphere-driven event should be reviewed, and the role of internal variability in relation to advection-convection feedbacks should be reconsidered.

We found that EOF analysis can be a trusted instrument of investigation when coherent processes domi-

nate basin-scale dynamics. In particular, it promises to be efficient in extracting information from climatological simulations of the eastern basin circulation, provided it is applied to variables that are suitably representative of large scale structures. Of course, further investigation is required to give a physical interpretation of such results that are, by definition, of a statistical, inventory nature. Nevertheless, we are convinced that they constitute a worthy guidance for subsequent analysis.

The EOF representation of western basin dynamics is hampered by the lack of a spatially coherent process on basin-scales that might be expected to respond either to the local change in boundary conditions or to the oscillations observed in LIW properties. Nevertheless, a response is observed in local dynamics, such as deep water formation in the Gulf of Lions, where in Expt 2 convective events appear to be enhanced and exhibit greater interannual variability than Expt 1. It is difficult to discern whether this is due to the preconditioning effect of Levantine water of variable temperature and salinity, or determined by the local impact of mixed boundary conditions on convective processes, or both. In fact, any correlation analysis between local time series in the Gulf of Lions and in the eastern basin is undermined by the consideration that the abrupt onset of convection is probably dependent on threshold values of water properties, and also on local factors that are independent of the signal transferred through the Sicily Channel by the advected LIW. Moreover, the climatological signal might be concealed by high frequency noise. Therefore, the analysis of the western basin also demands the development of suitable statistical instruments that are capable of relating phenomena occurring over distinct space and time scales. However, the question of teleconnections cannot be overlooked, as switching to mixed boundary conditions may have an impact on the dynamics of convection through advective-convective feedbacks.

This would account for the different response observed in distinct locations where convective processes take place. Therefore, we cannot exclude an effect of the variability in LIW properties on the preconditioning of winter convection in the Gulf of Lions.

In the near future, simulation under mixed boundary conditions will be prolonged up to 1000 yr. In addition, more sensitivity experiments are required in order to provide a more robust statistical basis for the characterization of multi-decadal natural variability of the Mediterranean Sea, and to firmly found the description of the system in terms of teleconnections between sub-basins and of correlation and interdependence between convective processes and general circulation.

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