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"Sensitivity of the Mediterranean Thermohaline Circulation to Vertical Diffusivity in OGCM"

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SENSITIVITY OF THE MEDITERRANEAN THERMOHALINE CIRCULATION TO VERTICAL DIFFUSIVITY IN OGCM

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RIASSUNTO

Un modello numerico derivato dal Cox è usato per studiare la circolazione termoalina del Mediterraneo. Il modello è aperto a Gibilterra ed include una zona di buffer nel Nord Atlantico.Un studio sulla sensibilità della circolazione termoalina alla variazione del coefficiente di diffusione verticale K_v mostra un proporzionalità inversa tra K_v e l'intensità della cella termoalina. Il trasporto medio di massa a Gibilterra e nello Stretto di Sicilia varia con K_v ed assume valori, rispettivamente, $0.97 < Q_g < 1.19$ e $1.36 < Q_s < 1.63$ [Sv] per K_v compreso tra 0.0-1.0 cm^2/s. Questa proporzionalità inversa è investigata studiando I flussi aria-mare di galleggiamento (buoyancy) in funzione della densità superficiale dell'acqua. I risultati ottenuti dal modello vengono confrontati con quelli sperimentali. Da ciò risulta che il flusso di sale "numerico" nel bacino Levantino, dovuto ad un eccesso di evaporazione sulla precipitazione, è strettamente correlato con l'intensità della circolazione termoalina e agisce da "motore" principale. Inoltre è possibile ottenere stime quantitative della formazione di Acqua Levantina Intermedia (LIW) e verificare il diverso ruolo del miscelamento verticale nei bacini Est ed Ovest. Viene infine proposto uno schema sintetico del bilancio, dalla sorgente superficiale Levantina al flusso uscente a Gibilterra.

ABSTRACT

A derived Cox model applied to the Mediterranean Sea is used to study the thermohaline circulation in the Mediterranean Sea. The model is open at Gibraltar and includes a buffer zone in the North Atlantic.
A sensitivity study of the thermohaline circulation in the Mediterranean Sea vs. the vertical diffusivity shows an inverse proportionality between the tracer vertical diffusivity and the strength of the thermohaline cell. The averaged mass transport at the Gibraltar and Sicily straits depends on the vertical diffusivities and assumes respectively the values of $0.97 < Q_g < 1.19$ and $1.36 < Q_s < 1.63$ [Sv] for K_v ranging to 0.0-1.0 cm^2 / s. This inverse proportionality is rationalised studying the surface air-sea buoyancy fluxes as a function of a surface water density. The results from the model are compared with the experimental ones. It is shown that the numerical "salt flux" in the Levantine basin, representative of an excess of evaporation over precipitation, is strictly correlated to

thermohaline circulation strength and acts as its real engine. Moreover it is possible to obtain some estimate on the annual rate of Levantine Intermediate Water formation and to investigate the different role of vertical mixing in the eastern and western sub basins. A synthetic budget scheme from the Levantine surface source to the Gibraltar outflow is proposed.

1 INTRODUCTION

In the last years the greater time computing availability and the considerable progress obtained by the scientific community make practical the possibility of simulate the oceanic circulation within primitive equations numerical models. This is true in particular for the case of the Mediterranean sea whose principal circulation features are by now substantially reproduced by these numerical model (see Zavatarelli and Mellor 1995, Roussenov et al. 1995, Wu and Haines 1996).

The Mediterranean is a concentration basin that is bathymetrically distinct from the North Atlantic; the net result of the air sea interactions in the entire basin consists in an outflow at Gibraltar of a salty water mass and in an inflow of fresher water. But despite this simplicity, the thermohaline circulation (THC hereafter) on the Mediterranean sea remains complex. In fact, within the Mediterranean system itself there are further bathymetric divisions in different subbasins, and most of them are distinguished by meteorological characteristics that induce a negative THC, i.e., the surface fluxes give rise to a formation of dense water.

The main feature of the thermohaline circulation in the Mediterranean, is the formation and the spreading of the Levantine Intermediate Water (hereafter LIW), characterised by a salinity relative maximum, in the Levantine basin (Wust, 1961). The core depth of this water mass (about 300 m.) permits the LIW to spread in almost all the subbasins of the Mediterranean, where its hydrological properties are changed by mixing with other water masses, before flowing out through Gibraltar. The Gibraltar outflow is then the result of a complicate ensemble of time dependent formation and mixing processes, rather than a simple surface flux balance (Tzipermann and Speer,1994, hereafter TS).

In this work we use a GFDL a primitive equations numerical model derived from the Bryan (1969) model and adapted to the Mediterranean case (see section 2). In particular we focus our attention to the sensitivity of the THC of the Mediterranean sea on the vertical diffusivity (section 3). The parametrizations of small scale mixing processes in terms of grid-scale variables, is one of the fundamental problems in modelling the ocean general circulation. Previous work (Bryan 1987, Cummins et al. 1990, Hu 1996) already demonstrated that the THC is particularly sensitive to the way the vertical diffusivity is parametrized. Here we use a Fickian diffusion with a diffusivity coefficient K_v spatially and temporally constant. Other choices are possible: for example to better represents the dyapicnal mixing, the diffusivity coefficient can be assigned to be inversely proportional to the buoyancy frequency N (Hu 1996).

The main results of the previous sensitivity studies consisted in observing a proportionality between the THC and the magnitude of the vertical diffusivity (Bryan 1987, Cummins et al. 1990, Hu 1996). It is to note that all this sensitivity study was performed using rather coarse models representing an idealised ocean basin, to qualitatively represents the climatic circulation in the North Atlantic. At our knowledge the sensitivity of the Mediterranean circulation on the vertical diffusivity was not still investigated. The interest of this study resides in introducing strong topography constraints and more realistic surface fluxes. Moreover this study, in addition of providing useful information about the feasibility of representing correctly critical parameters as the Gibraltar outflow, by using a particularly large Atlantic buffer zone, leads us to a more comprehensive representation of the mechanisms involved in the LIW formation processes. In fact, calculating the air-sea buoyancy flux as a function of the surface water density (TS), it is possible to obtain some estimate on the annual rate of LIW formation and on the importance of the vertical mixing in the eastern and western Mediterranean sub basins (section 4). These results will be resumed and discussed in the final section.

2 THE MEDITERRANEAN MODEL

We employ the widely distributed MOM version of the Bryan-Cox ocean general circulation (Bryan 1969) adapted to the Mediterranean geometry (Pinardi and Navarra, 1993, Roussenov et al. 1995, Wu & Haines 1996) with the same 19 levels vertical resolution. The horizontal resolution is .25° in Longitude and Latitude. The model is open at Gibraltar and includes a large box for the North Atlantic (until to 13°W) (Fig.1). The 3D hydrological properties of the western part (buffer zone) of the Atlantic domain are relaxed toward Levitus (1982) climatology. Hellerman and Rosenstein (1983) wind fields are used to force the model at the surface.

The model is a rigid-lid primitive equations, written in spherical coordinates. The prognostic equations are for the barotropic streamfunction, u',v' (residual velocities with respect to the vertical mean), T, S. ρ is computed by the non linear UNESCO equation. For the surface boundary conditions Levitus (1982) temperature (T^*) and salinity (S^*) monthly fields are used. The surface fluxes are then introduced by adding a nudging term in the prognostic equations for T and S, respectively:

$$\alpha_T(x,y,z)(T-T^*) \qquad \text{and} \qquad \alpha_S(x,y,z)(S-S^*).$$

The time step used is 3600 s for both the tracer and momentum equations. The coefficients of turbulent diffusion are: $A_h = 8\cdot10^{18}$ cm^4/s, $K_h = 2.4\cdot10^{19}$ cm^4/s, $A_v = 1.5$ cm^2/s, $K_v = 0.0$-1.0 cm^2/s. We study the sensitivity of the THC on the tracer diffusivity K_v performing 5 11-years long experiments with various values of K_v (0.0, 0.1, 0.3, 0.6, 1.0 cm^2/s). (We will name these experiments with the letter K followed by two number representing the values of K_v from K00 to K10).

The case of $K_v = 0.0$ is introduced for a purely speculative interest. It is known that in a non equally spaced grid the difference algorithms used by the Cox model (1969) (see Yin et al.,1991) are accurate only at the first order and more, that the resulting inacurateness give rise to a purely numerical diffusive terms. This run performed in order to assess the relevance of this numerical diffusivity and we will discuss its results because they are representative of the case of slightest vertical diffusivity.

Convection is introduced into the model using the convective adjustment proposed by Bryan (1969): when static instability occurs the temperature and salinity of the unstable column are locally homogenised, by the way a weighted average, until neutral stratification is reached. The surface relaxation (Haney 1971) toward Levitus climatology is performed using constant values of $\alpha_T(x,y,0)=\alpha_S(x,y,0)=\alpha=5$ days^{-1}, for both temperature and salinity, in all the domain, except for the Atlantic buffer zone where we kept of $\alpha_T(x,y,z)=\alpha_S(x,y,z)= \alpha=5$ days^{-1} in all the water column. The evaporation and precipitation events are then simulated by a fictitious surface salt flux done by the restoring terms in the equation. For this reason in the following we will speak of surface 'salt flux' or straits 'salt transport'. This kind of fluxes and transports obviously do not exist in nature, and when in the following we will use those expressions we will be implicitly refer to the (E-P) fluxes and to the corresponding salinity layer difference in the Strait.

The choice of adopting Haney relaxation avoid uncertainties in parameterization of the air-sea fluxes. One could force the model with more restrictive condition, as for example increasing the values of the α terms or prescribing specified fluxes in the sites of dense water formation (Haines and Wu 1995); by doing so the thermodynamics would be dominated by the external forcing everywhere, reducing the active role of the tracer advection in the THC.

Fig. 2 shows that after about 5-7 years of integration the model reaches the stationary equilibrium, with an slightly growing e-folding with K_v .

3 SENSITIVITY OF THE THC ON THE VERTICAL DIFFUSIVITY

The mass, heat and salt transport at the Gibraltar and Sicily Straits are computed in the 11th year of simulation. The averaged annual values are computed averaging over 10 days snapshot outputs. As a global checking of the way of operating and the of behaviour of the model we report in Fig. 3 the values of the salinity on the isopycnal surface 27.7. The tongue of outflowing salty Mediterranean water and its spreading along the Portugal coasts are evident. On the other hand we report in Fig. 4 the one year averaged mass heat and salt transport at the Sicily and Gibraltar outflow computed in the different runs. Fig. 4 displays a clear behaviour: the mass and salt transport are correlated and shows an inverse proportionality to the vertical diffusivity coefficient K_v. Even if there is a rather strong variation on the mass and salt transport with K_v, it is to stress that the averaged values ($0.97<Q_g<1.19$ Sv and $1.36<Q_s<1.63$ Sv for the Gibraltar and Sicily strait) fall within the observational uncertainties (see for instance Bethoux 1980, Candela 1991, Grancini & Michelato 1987). On the other hand the heat transport it is not adequately reproduced by the model; the numerical simulations overestimate the temperature of the out flow at straits of about 0.5 degree.

The Sicily mass transport shows a clear seasonality with a winter maximum, while the Gibraltar mass transport displays a more erratic behaviour during the year (Fig. 5). Note also that in the Sicily Strait the mass transport is sensitive to K_v only in the magnitude of the sub basins exchange. The qualitative behaviour during the year is the same for all the different runs, with a well definite winterly maximum. On the contrary the Gibraltar mass transport is sensitive to the vertical diffusivity also in its qualitative behaviour during the year. There is not a clear seasonality and the behaviour becomes more irregular as K_v decreases.

The strengthening of the THC for decreasing K_v is also evident in for the meridionally integrated transport, or 'zonal overturning' (see Hu 1996) (not shown here).

From Fig. 6, representing a meridional density section in the Rhode Gyre region, it is possible to see how in the most active THC case (K01) the 28.5 isopycnal surface intersects the surface and any homogenisation of the water column is reached during the winter. It is finally noteworthy that the convection events (i.e. number of time in which the convective adjustment is switched on in order to eliminate the hydrostatic instabilities) grows going from the K00 to K10 experiments.

Moreover, it is to note that the Salt content in the basin is constant during long integration i.e. the salinity averaged in all the basin is constant even in long integration.

All this is indicative of a well definite scenario: decreasing the vertical diffusivity the mass and the salt transport at the straits increase: as the salt content is constant in the basin, this more efficient THC has to be necessarily sustained by a stronger 'surface salt' flux somewhere in the basin.

In the next section we will return on this concept in a more quantitative way.

4 SURFACE CROSS ISOPYCNAL MASS FLUX FUNCTION

This section in entirely inspirated by the seminal paper of Tzipermann and Speer (1994, hereafter (TS). They shown as calculating the total air-sea buoyancy flux as a function of a surface water density, it is possible to obtain some estimation about water mass formed in any density range. We report here some of their speculation but we will refer to their article for a more exhaustive discussion on the interpretation of the cross isopycnal mass flux function. We simply computed this function using the model numerical output from the different experiments with different vertical tracer diffusivity, using a 10 days snapshot output in the 11^{th} year of integration.

The surface cross isopycnal mass function F(ρ) is defined as:

$$F(\rho') = \frac{1}{T_0 \cdot \rho_0} \int_T dt \int_A dxdy \left[\frac{\alpha}{C_p} H(x,y,t) - \rho\beta S(x,y,t) Q(x,y,t) \right] \cdot \delta\left[\rho(x,y,t) - \rho'\right] \tag{1}$$

Where the heat and the freshwater surface fluxes H(x,y,t) and Q(x,y,t) are expressed in [W l^{-2}] and [lt^{-1}], $\alpha = -\frac{\partial\rho}{\partial T}, \beta = \frac{\partial\rho}{\partial S}$ (where C_p is the specific heat coefficient at constant pressure).

The brackets term in (1) represents the buoyancy flux due to the air-sea interaction as a function of space and time. Calculating the integral (1), it is then possible to obtain the contribution of the surface air sea interaction to the cross isopycnal mass flux on the entire spatial domain in the time interval T. Note that F(ρ) is expressed in Sv. The delta function appearing in the integral make possible to express this mass flux as a function of any density. If T=1 year, then F(ρ') represents the contribution of the air-sea interactions to the mass flux across the isopycnal surface ρ=ρ'. Note that in this function there is not any information about isopycnal mixing or about any possible vertical deplacement of a water body. F(ρ) represents only the atmosphere contribution in changing the surface water hydrological properties; when a water body of density ρ' is made less or more dense by surface interactions, $F(\rho')$ is different from zero also if this water body remains in surface. The cross isopycnal flux in this case is not across any physical surface.

If $F(\rho)$ is greater (less than) 0 then the air-sea interactions contribute to making lighter (heavier) the water at the density ρ. Moreover it is possible to see (TS) that the mass of water formed in the density range (ρ_2, ρ_1) is equal to $F(\rho_2) - F(\rho_1)$, i.e., taking the limits $\rho_2 \rightarrow \rho_1$ we have that

$$M = dF/d\rho, \qquad (2)$$

where M represents the amount of water formed at the density ρ.

In Fig. 7 we compare the numerical (a) K03 experiment) and experimental (b) from TS) averaged cross isopycnal mass flux both for the year and for each season separately. We computed $F(\rho)$ approximating the integral with a sum as in TS. Both the seasonal cycle and the magnitude of the cross isopycnal mass flux are well reproduced (K03 case) in the numerical experiment. It is to note that also the haline and thermal contribution to $F(\rho)$ (not shown here) from the numerical experiment are separately in a good agreement with the experimental one (from TS). For a detailed description of the significance of the mass flux function we refer to TS; here we want only to stress that computing $F(\rho)$ shows the great importance of the cross isopycnal mixing in the water mass formation process. In fact, following TS and using (2), it is possible to see that the air sea fluxes transform an average 5-6 Sv of water from the density range ρ=(26.4, 28.4) to smaller and greater densities. Comparing this mass flux with the present Gibraltar exchange mass magnitude (~1 Sv) shows the huge consequence of the cross isopycnal mixing in the Mediterranean dynamics.

In Fig. 8 we plot $F(\rho)$ computed for 4 sensitivity experiment separately in the Eastern and Western sub basins. The magnitude of the cross isopycnal mass flux increases with K_v: for greater diffusivities the tracer values of the surface move away faster from the climatological values and the surface flux consequently increase. It is to note that in the western sub basin the surface mass flux function is practically equal to zero in the K00 experiment, while in the eastern sub basin, $F(\rho)$ also if decreases with K_v is still significatively different from zero also in the K00 experiment. All this shows that the dynamics in the western basin is strongly dominated by the vertical mixing (TS). On the other hand from Fig. 9, where are plotted the haline contributions to $F(\rho)$ computed only in the Levantine Basin (the site of most of convective events), it is possible to see how the haline contribution to the cross isopycnal mass flux *increases* as K_v decreases. It is to stress that we observe this behaviour only in the Levantine basin in the haline term. This behaviour, still not well understood, is coherent with the results reported in section 3 showing the inverse proportionality between K_v and the THC.

All this lead to us to perform an estimate of the annual rate of LIW formation. In fact computing the different contribution to (1) in the different sub basins we found evidence that the increasing of THC with decreasing K_v is sustained by a corresponding increase of 'numerical surface salt flux' in the Levantine basin. Also if the cross isopycnal mass function does not support any indication about vertical deplacement, we can exploit this information in order to individuate in the water mass made denser by (density ρ>28. 5, $dF/d\rho > 0$), the main part of the sinking water necessary to sustain the THC. It is to stress that all this and the relation (1) lead to us to have only an overestimate of the water mass budget in the THC. The resulting estimates for surface T and S for K01, yearly averaged in the Levantine basin on water parcels with $\rho > 28.5$, are 15.64°C and 38.78 psu. The values of the averaged T and S for K01 for the outflows at the straits are 14.01°C, 37.92 psu, 14.98° C, 38.58 psu for Gibraltar and Sicily, respectively.

Once again we have evidence of the greater relative importance of cross-isopycnal mixing in the western basin. In fact the averaged hydrological values of the Sicily outflow are similar to those of the Levantine surface source, showing as the eastern basin dynamics does not change much the hydrological characteristics of the sinked water and seems to be characterised by advection and convection more than isopycnal diffusion.

CONCLUSIONS

Performing several sensitivity experiments we showed as the intensity of the THC in the Mediterranean Sea, as well the magnitude of salt and mass transport at the straits, is a decreasing function of the vertical diffusivity. It was shown that tuning the vertical diffusivity it is possible to increase the convective activity in the Levantine Basin and the corresponding LIW rate formation using climatological restoring conditions. This result has to be compared with the previous work (Bryan 1987, Cummins et al. 1990, Hu 1996) in which was demonstrated both numerically and in a speculative way a direct correlation between K_v and the Atlantic thermohaline circulation cell. The understanding of this direct correlation was based on dimensional relation essentially relating the thermocline depth to the THC. Our present result has to be better rationalised introducing concepts as the strong topographic constraints and or the occurrence of frequent and strong convective events induced by the air sea interactions.

The Sicily and Gibraltar mass, heat and salt transport were compared with the experimental observation founding a good agreement for the mass and salt transport. On the other hand the numerical simulation underestimate the heat transports at the straits probably due to an overestimate of the temperature of the outflowing layer. This behaviour is probably due to an systematic shift in the Levitus surface temperature. Numerical simulation with different climatic data will be analysed in a forthcoming work.

Moreover it was shown that the cross isopycnal mass flux function computed from the numerical model is in good agreement with the experimental data. This function was shown to be an useful tool to rationalise the observed inverse proportionality between THC and vertical diffusivity.

More specifically, it was shown that the necessary buoyancy flux need to sustain the increasing THC with decreasing K_v is supported by the surface 'salt flux' (or realistically speaking by the excess of evaporation over precipitation) in the Levantine Basin. This shows as the haline component in spite of being numerically smaller than the thermal one in the buoyancy flux acts as the real engine of the THC.

Finally, a schematic mass budget from the 'Levantine Source' to the Gibraltar outflow was computed by the way of a synthetic budget. This as well the sensitivity of the surface cross isopycnal mass flux function on the vertical diffusivity, lead to us to differentiate the role of the vertical mixing in the two main subbasins. The huge relevance of the vertical mixing in the western sub basin has to faced with the eastern dynamics more characterised by advection and convection events that are much less affecting the hydrological properties of the water mass involved in the THC, from the Levantine source to the Strait of Sicily outflow.

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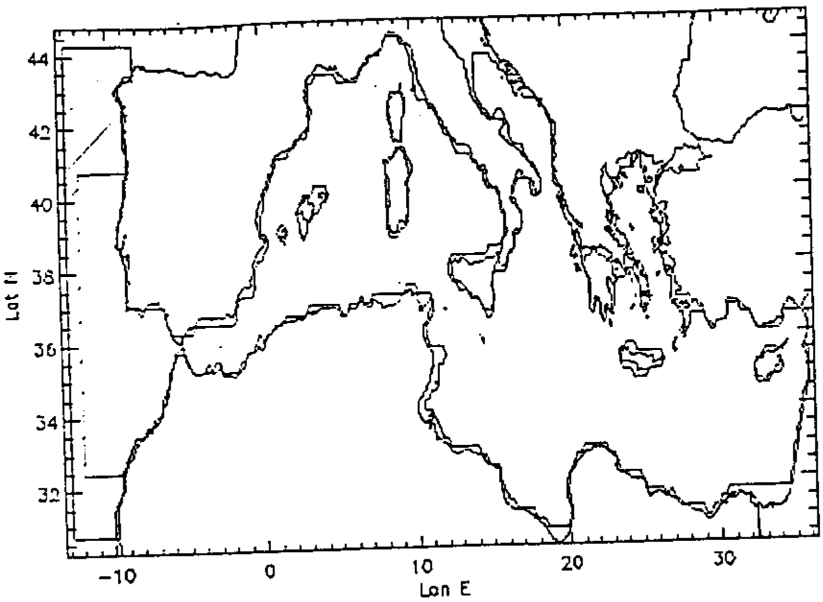
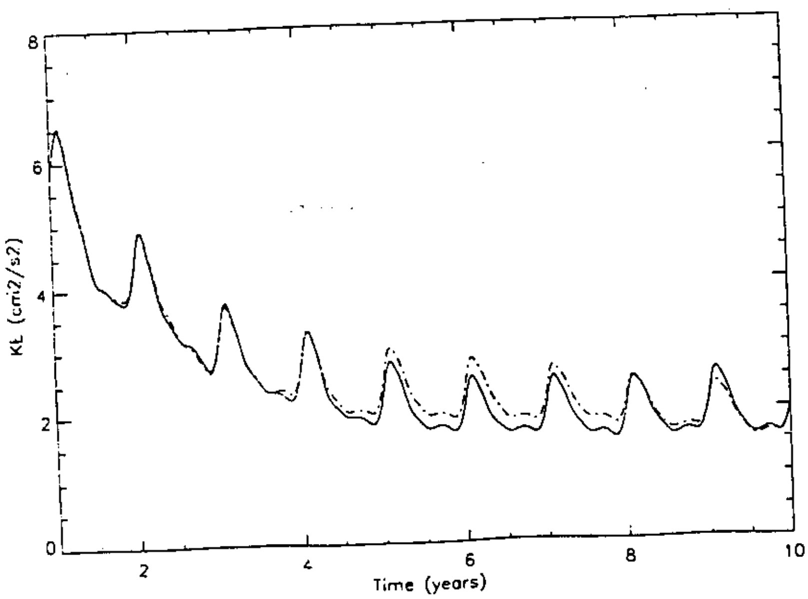
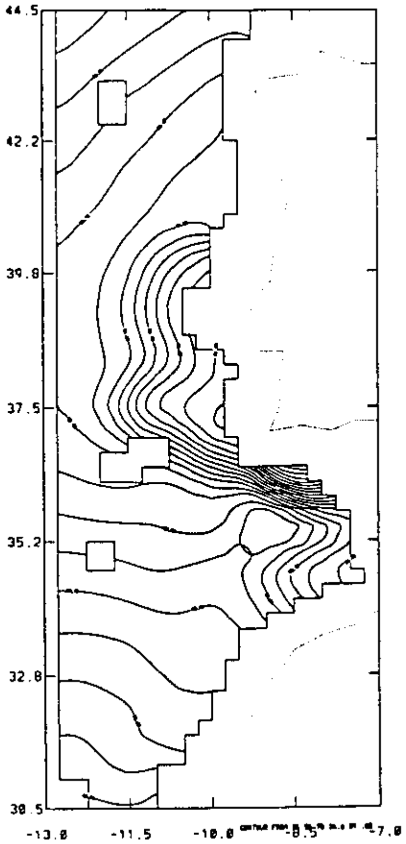
FIGURE CAPTIONS

Figure1 Model domain with in evidence the Atlantic buffer zone.
Figure 2 Mean kinetic energy vs time for the run K01 (continuos line) and K10 (dash).
Figure 3 Map of salinity on the 27.7 isopycnal surface.
Figure 4 a) Salt transport and b) mass transport at Gibraltar (cross) and Sicily (triangle) Straits vs K_v.
Figure 5 Mass transport for the 11[th] year at the Straits for K01 (Gib. continuous line, Sic. dot-dash) and K10 (Gib. dot, Sic. dash).
Figure 6 Meridional section at 29°E of sigma-t for March 15[th] 11[th] for K01.
Figure 7 Averaged cross isopycnal mass flux vs density for K03 (a) and from experimental data (sketched from TS) (b). For (a): annual mean (dash), Summer (cross), Fall (asterisk), Winter (circle), Spring (thin line). For (b): annual mean (thick line), Summer (dot-dash), Fall (thin line), Winter (dash), Spring (dot).
Figure 8 One year averaged cross isopycnal mass flux vs density for experiments K00 (continuous), K01 (cross), K03 (asterisk) and K10 (circle) for the Western basin (a) and Eastern basin (b).
Figure 9 Salt contribution to the one year averaged cross isopycnal mass flux vs density for experiments K00, K01, K03 and K10 for the Levantine basin (the notation is the same of figure 8).



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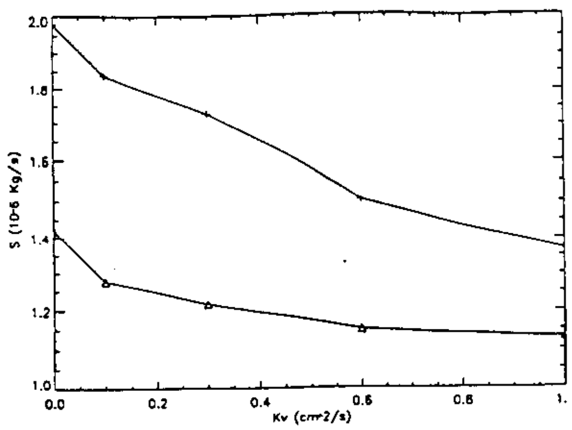
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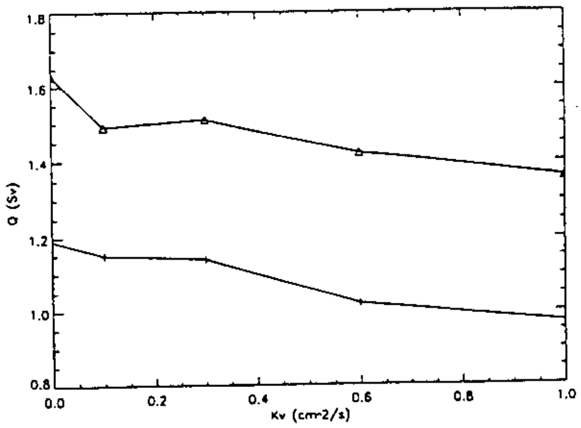
Kv (cm2/s)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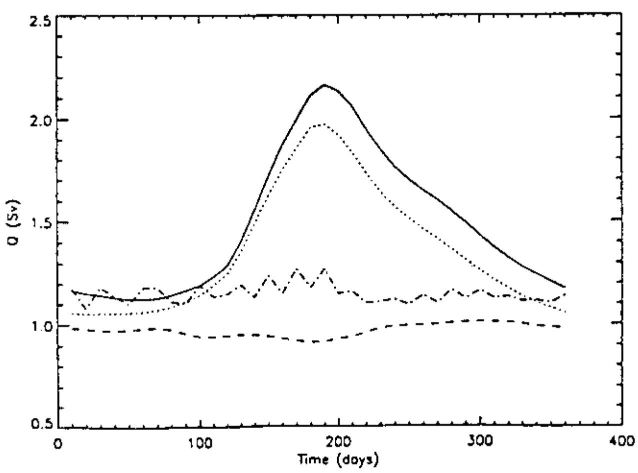
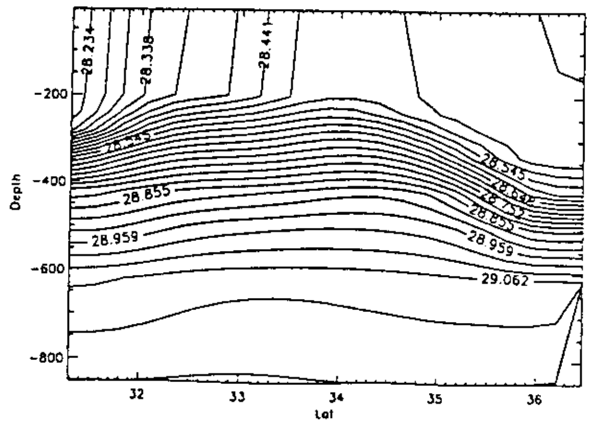
Kv (cm2/s)


Q (Sv)
Time (days)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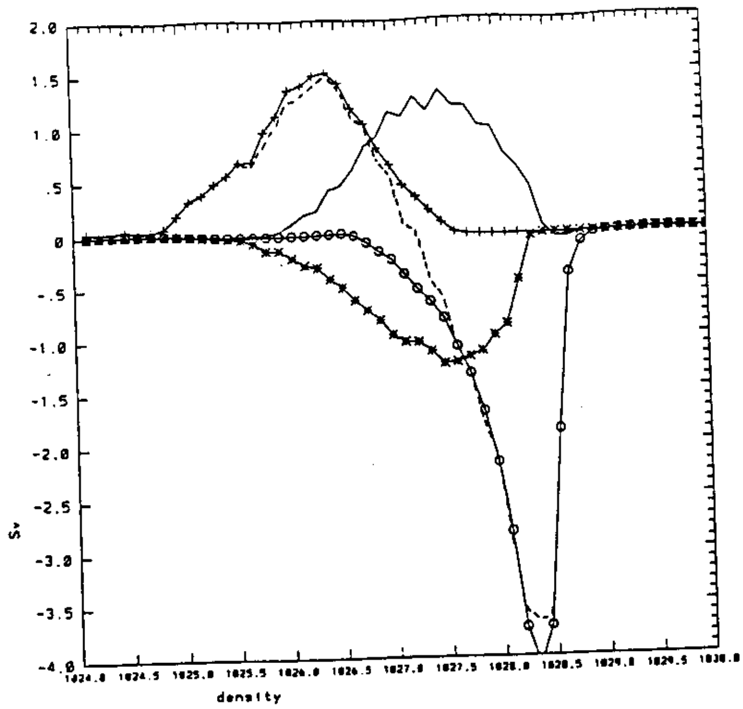
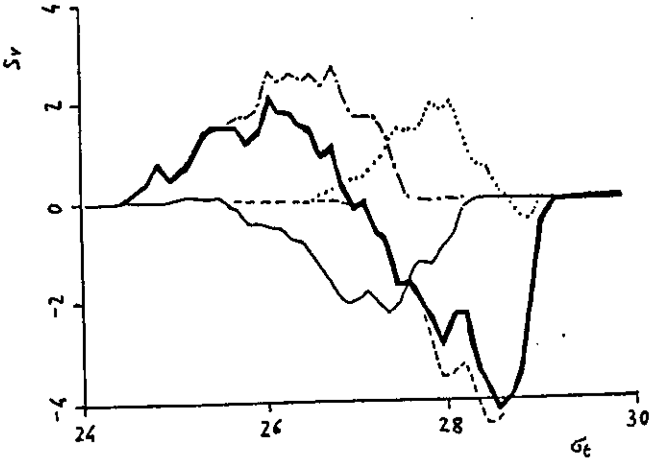
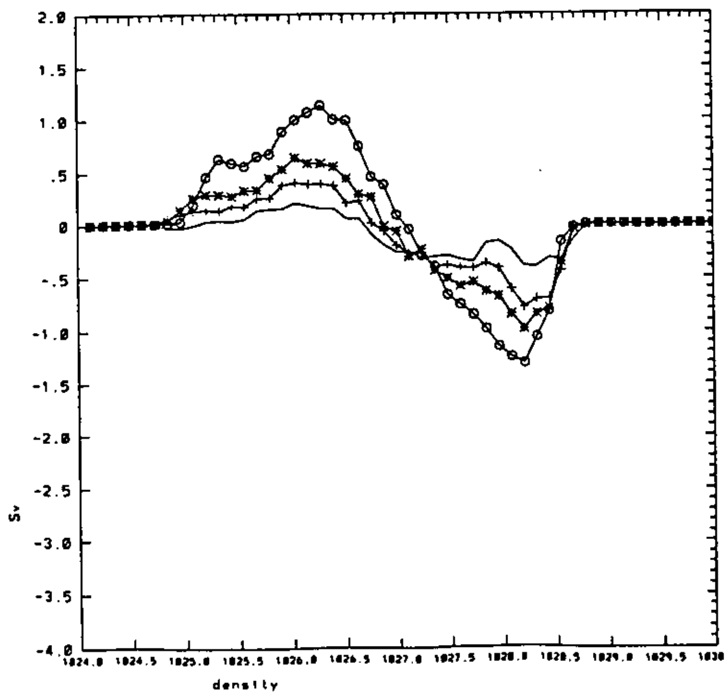
Depth
Lat

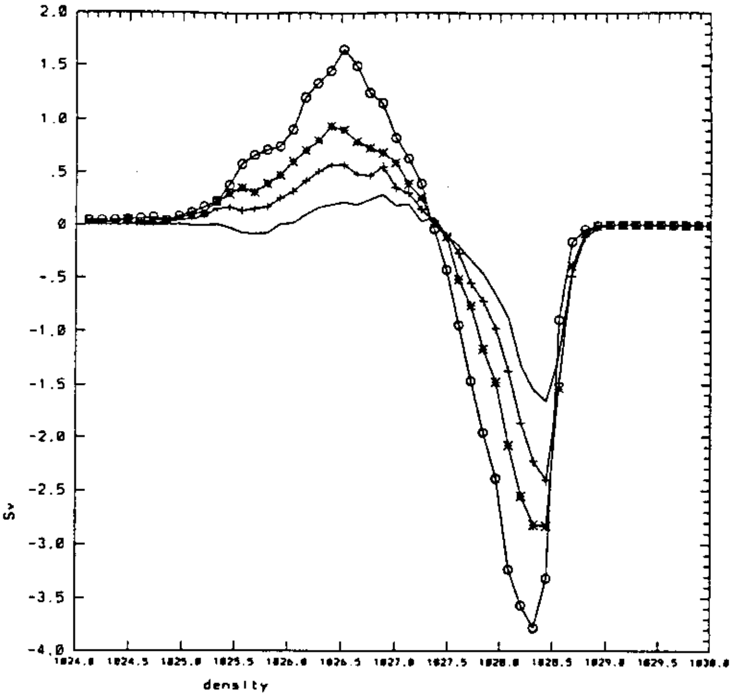




WEST LEV NS3 Kz 0.0-1.0



EST LEV NS3 Kz 0.0-1.0



LEVANTINE LEV-NS3 Kz 0.0-1.0 SALT