ELSEVIER

Journal of Marine Systems 31 (2001) 65–76

JOURNAL OF MARINE SYSTEMS

www.elsevier.com/locate/jmarsys

Convection in the Black Sea during cold winters

L.I. Ivanov [a,*], J.O. Backhaus [b], E. Özsoy [c], H. Wehde [b]

[a] Marine Hydrophysical Institute, Ukrainian NAS, 2, Kapitanskaya St., Sevastopol 99011, Ukraine
[b] Institut fuer Meereskunde Universitaet Hamburg Troplowitzstr, 7, D-22529 Hamburg, Germany
[c] Institute of Marine Sciences, Middle East Technical University, P.K. 28 Erdemli-İçel 33731, Turkey

Received 17 January 2000; accepted 20 March 2001

Abstract

We investigate the effects of initial conditions and atmospheric forcing on the penetration depth of winter convection in the strongly stratified environment of the Black Sea. The amount of Cold Intermediate Water (CIW) formed annually in the Black Sea, and the amount entrained into the euphotic zone from below, largely depend on the convective activity in the cold season. Seasonal thermohaline evolution of the mixed layer has been simulated using a one-dimensional mixed layer model driven by the ECMWF meteorological data, and initial conditions based on hydrographic data collected in the northwestern Black Sea in autumn. Simulations have been carried out for three different regions with typical hydrological conditions. The results were compared with data obtained in the spring seasons of successive years. Convection events, primarily responsible for the replenishment of the CIW, are shown to last for several days and coinciding with severe storms. The model results suggest that inhomogeneities in the CIW density and thickness could be a consequence of initial stratification in different parts of the Black Sea. In all of our experiments, convection had limited effect in modifying thermal and density structure of the main pycnocline.

Keywords: Black Sea; Air–sea interaction; Mixing processes; Oceanic convection; Pycnocline erosion

1. Introduction

The upper layers of the Black Sea are strongly stratified. The salinity stratification is maintained due to the excess of the fresh water inputs over evaporation (Özsoy and Ünlüata, 1997). Ventilation of the upper pycnocline (halocline) occurs episodically through the process of winter convection that results in the formation of the so-called Cold Intermediate Water (CIW) characterising the Cold Intermediate Layer (CIL) of the Black Sea, typically characterised by waters with temperature below 8 °C. The CIL core, i.e. the layer of minimum temperature, occurs within a depth range of 30–80 m (Fig. 1). It overlies the main pycnocline, where a hydrostatically stable stratification is maintained by the stable salinity gradient despite a destabilising influence of temperature increasing with depth. The core of the CIL occurs at sigma-t = 14.2–14.8. Typically, the minimum temperature varies from about 6.5 °C in March to 6.8–6.9 °C in May–November (Fig. 2). The lower boundary of the CIL occurs at sigma-t = 15.3–15.9, corresponding to mid-pycnocline.

The main pycnocline in the Black Sea is dome-shaped. The CIL core is shallower in the interior of the basin (30–50 m) and deepens towards the conti-

* Corresponding author. Fax: +7-380-692-413-099.
E-mail address: max777@ukrcom.sebastopol.ua (L.I. Ivanov).

0924-7963/01/$ - see front matter
PII: S0924-7963(01)00047-1

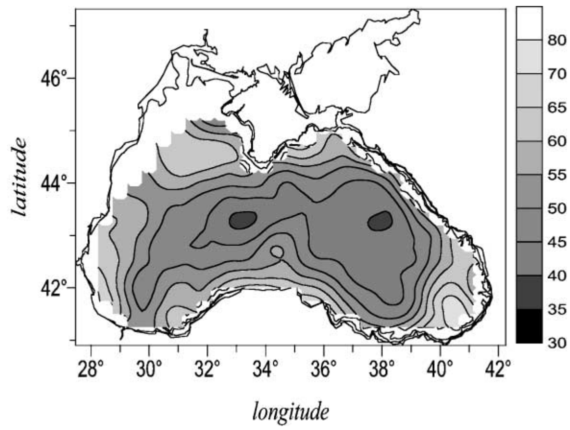

Fig. 1. Mean position of the CIL core for May–February derived from hydrographic data of the 1980s and 1990s.

nental slope to reach a depth of 80 m within the ‘convergence zone’ (i.e. the frontal region of the ‘Rim Current’, separating the cyclonic interior from the coastal domain typically containing a series of anticyclonic eddies). Some of the anticyclonic eddies are quasi-stationary structures confined to certain features of coastal and bottom topography (Blatov et al., 1984; Oğuz et al., 1993; Ivanov et al., 1998).

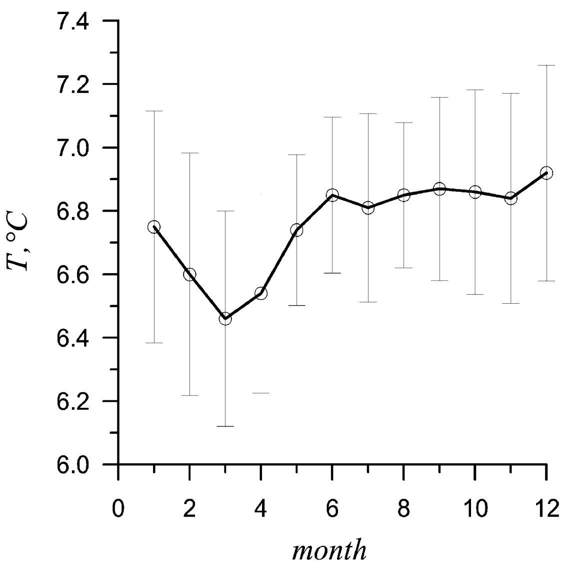

Fig. 2. Seasonal course of minimum temperature for the open part of the Black Sea derived from hydrographic data of the 1980s and 1990s.

Although observations are limited, convection driven by atmospheric cooling in winter is expected to erode the pre-existing structure and establish an isothermal layer reaching from the surface to the core of the CIL, as required by the initial explanations of the origin of CIL made in the 1930s (Knipovich, 1932; Zubov, 1938), which ascribed it to local deepening of the mixed layer by winter convection. The layer survived under a warm mixed layer in summer.

Fig. 3 shows the sea surface temperature climatology for February, corresponding to the period of severest weather in the Black Sea. If we were accepting a hypothesis of local formation of CIW in all areas, we would expect to have waters with freezing temperatures in the CIL core in the periphery of the western basin, and warm waters with core temperature exceeding 8 °C in the eastern basin. Such a tendency does not exist in the observations, which indicate relatively uniform temperature of the CIW core across the basin (except for a few places where reasonable temperature/salinity differences occur between core waters of coastal and central regions). It may thus be concluded that the process of CIW replenishment, around the entire basin, is mainly of an advective nature. This conclusion was first formulated by Kolesnikov (1953) and expounded upon by Filippov (1965, 1968). Accordingly, it was suggested that the CIW is formed at the northwestern shelf (NWS) and in the adjoining slope region of the Black Sea as well as in the proximity of the Kerch

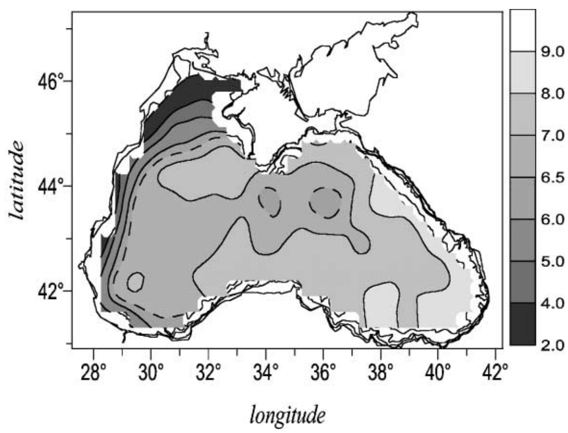

Fig. 3. Sea surface temperature climatology for February (3448 hydrographic stations for the whole Black Sea).

Strait (Filippov, 1965; Blatov et al., 1984; Tolmazin 1985). It was then suggested that the CIW should be advected to the south and the east. The severity of weather conditions in the NWS region (Altman and Simonov, 1991) attests to the possibility of enhanced cold water formation in the region.

The hypothesis of CIW production in the northwest part of the sea dominated until the 1980s. Measurements made with improved instruments (CTDs) revived the local generation hypothesis (Ovchinnikov and Popov, 1987) with a new emphasis on CIW formation confined to central part of cyclonic gyres, from where they would flow toward the coast, sliding along isopycnal surfaces, thereby replenishing the CIW in the convergence zone. This hypothesis excluded the possibility of cold water advection from the northwest region, as well as its effective formation in anticyclonic circulations. On the other hand, volumetric T,S analyses of water masses (Ivanov et al., 1997b) have revealed temperature/salinity indices of CIW within the convergence zone in distinct difference from the central cyclonic gyres, suggesting effective CIW formation within anticyclonic eddies under certain conditions. The bulk of CIW was often found trapped in anticyclonic eddies. It was thus concluded that the contribution of 'pure' shelf sources to CIW is negligible (Ivanov et al., 1997b). Modelling of CIW formation and spreading have supported these conclusions (Stanev and Staneva, 1997).

On the other hand, the hypothesis of Ovchinnikov and Popov (1987) on effective ventilation of the pycnocline in the cyclonic central part of the sea was supported by other studies investigating temporal variability within the pycnocline (Stanev and Staneva, 1997; Ivanov et al., 1997a; 1998).

Although the erosion of the main pycnocline largely depends on winter convection, it is very difficult to find a sufficient amount of winter-time hydrographic observations in the Black Sea to be able to confirm hypothesis relating to deep mixing in confined areas, and weigh the roles of isopycnal versus diapycnal mixing. It is thus important to investigate the role of extreme cooling events in pycnocline erosion, making use of hydrographic data analyses and mixed layer models in order to assess space-time variability of the process, and to determine the maximum penetration depth of convection. We follow this approach to investigate mechanisms that control the evolution of the mixed layer and CIW production under severe weather conditions. Section 2 of this paper describes the mixed layer model and the data. In Section 3, we focus on the description of pre- and post-cooling hydrographic conditions and results of numerical simulations. New results are discussed in the Summary.

2. Data and methods

2.1. The model

The Mixed Layer Model (MLM) solving the model equations for a one-dimensional water column is an independent part of the three-dimensional baroclinic ocean model HAMburg Shelf-Ocean Model (HAMSOM) (Backhaus, 1985; Schrum and Backhaus, 1999). It is comprised of the Ekman momentum equations, and the conservation equations for temperature and salinity:

$$\frac{\partial u}{\partial t} - fv = -\frac{1}{\rho} A_{\mathrm{v}} \frac{\partial^2 u}{\partial z^2}$$

$$\frac{\partial v}{\partial t} + fu = -\frac{1}{\rho} A_{\mathrm{v}} \frac{\partial^2 v}{\partial z^2},$$

$$\frac{\partial T}{\partial t} = \frac{\partial}{\partial z} K_{\mathrm{v}} \frac{\partial T}{\partial z}$$

$$\frac{\partial S}{\partial t} = \frac{\partial}{\partial z} K_{\mathrm{v}} \frac{\partial S}{\partial z},$$

complemented by non-linear UNESCO (IAPSO Working Group, 1981) equation of state, $\rho = \rho(S,T,P)$.· u and v are the horizontal velocities, A_{v} is the turbulent eddy viscosity, ρ is the density, and $f = 2\Omega \sin\varphi$ is the Coriolis parameter, φ is the geographic latitude. The turbulent eddy viscosity is derived from a zero-order turbulence closure scheme (Kochergin, 1987; Schrum and Backhaus, 1999) based on the instantaneous balance between shear production and hydrostatic stability. The eddy diffusivity (K_{v}) is thus related to eddy viscosity via a Schmidt number (set to unity in our experiments).

The heat flux between the ocean and atmosphere (Q_{NET}) is a function of air temperature, wind velocity, specific humidity and cloudiness, calculated from the following:

$$Q_{NET} = Q_L + Q_S + \Delta Q_{LO} + Q_{SH},$$

where Q_L is the latent and Q_S the sensible heat flux, ΔQ_{LO} is the difference between long-wave back radiation from the atmosphere and long-wave radiation from the sea surface, ΔQ_{SH} is the amount of the incoming short-wave solar radiation.

The momentum flux, evaporation and precipitation are derived from six-hourly ECMWF re-analysis. The turbulent heat flux components are computed from bulk-formulae (Friehe and Schmidt, 1976; Gill, 1982):

$$Q_L = \rho_{air} L_{vap} \beta_{lat} |W| (q_{air} - q),$$

$$Q_L = \rho_{air} \beta_{sens} C_{p,air} |W| (T_{air} - T),$$

$$\Delta Q_{LO} = \in_{air} \sigma T^4_{air,abs} - \in \sigma T^4_{sea,abs}$$

where ρ_{air} is the density of the air, L_{vap} is the latent heat of vaporisation, β_{lat} and β_{sens} are the turbulent transfer coefficients for latent and sensible heat; $|W|$ is the wind velocity near the sea surface, q_{air} is the specific humidity of ambient air, q is the specific humidity near the sea surface, $C_{p,air}$ is the specific heat of the air, $T_{air,abs}$ and $T_{sea,abs}$ are the absolute air and sea surface temperatures (K); σ is the Stefan Boltzmann constant, $\in_{air}$ is the atmospheric emissivity depending on cloudiness, and $\in$ is the sea surface emissivity.

Incoming short-wave solar radiation, depending on time, latitude, and observed local cloudiness, is calculated after Dobson and Smith (1988):

$$Q_{SH} = (1 - a) Q_0 S (A_i + B_i S)$$

with $a = 0.06$, the albedo, the solar constant Q_0, the solar angle S, and the cloudiness parameters A_i and B_i.

The MLM was run with a vertical resolution of 2 m and a time-step of 10 min. An implicit scheme for the approximate numerical solution of the diffusion equations for both velocity vector and temperature and salinity is used. This effectively avoids any numerical instability in a scheme with high vertical and temporal resolution, and with a time-dependent turbulent diffusion coefficient. Instabilities in stratification within the water column (i.e. parameterising convection) are removed by locally increasing the turbulent eddy viscosity to a maximum value (800 cm^2/s). The decay of (fossil) turbulence, or of turbulence, which was locally increased to erode an instability, is simulated by a time constant (set to half an hour) entering a linear weighted interpolation between the actual turbulent viscosity predicted by the simple closure scheme and the viscosity from the previous time step. The background diffusivity in the pycnocline in most of the experiments was set equal to 0.02 cm^2/s. At the seabed, the MLM utilised a non-linear friction law. The MLM was typically initialised by an observed T,S profile, assuming an ocean at rest. The water column, thus containing a realistic stratification, is then exposed to atmospheric forcing which, in our experiments, comprised a time period of several months.

2.2. The data

The hydrographic data used in this study were collected on board of the research vessels of the Marine Hydrophysical Institute, Ukrainian National Academy of Sciences, in the northwestern sector of the Black Sea in the framework of the CoMSBlack and TU-Black Sea international programmes. Since the northwestern part of the sea is an area with the most severe weather conditions in winter (Altman and Simonov, 1991), we believe that studies of winter convection in this region adequately describe extreme cases of CIW formation in the whole Black Sea. The original hydrographic data used as initial conditions in model simulations constitute vertical temperature and salinity profiles with 1-m vertical resolution. The results related to Black Sea climatology and utilised in the paper have been attained from the available climatic data sets prepared in the database laboratory of the Marine Hydrophysical Institute. Only data with fine vertical resolution were used: about 3650 hydrographic stations for the whole Black Sea with the best coverage for the northwestern part (1770 stations) and minimum amount of stations in the southwestern part of the basin (310 stations).

In accordance with the winter severity index, traditionally calculated as the sum of negative air temperatures, $\sum t_a$, for meteorological stations of Odessa, Ochakov and Khorly (all in the northwestern part of the basin), the winter of 1992–1993 could be characterised as a winter with moderate weather conditions (Altman and Simonov, 1991; Belokopytov, 1998). However, the years 1992 and 1993 were years of cooling for the lower troposphere for the entire globe, and especially in an anomalous pattern centred near the eastern Mediterranean region, attributed to the radiation effect of dust from the Pinatubo volcanic eruption in 1991. The cooling continued to prevail in the winter-time air and sea surface temperatures of the two subsequent years and had a direct effect in the southern part of the Black Sea, where the cooling was greater in 1993 (Özsoy and Ünlüata, 1997; Özsoy, 1999). In addition, the response of the sea was anomalous in terms of CIL temperature and CIW formation (Belokopytov, 1998; Ivanov et al., 1997a, 1998). To understand the reason for that anomalous response of the sea during the winter of 1992–1993, we focused on simulations of the evolution of thermal conditions over the period from October 1992 to April 1993.

The survey conducted in April 1993 covered about 70% of the area of the sea, and hence, it had been possible to apply methods of statistical T,S analysis, similar to the analysis applied to the data of basin-wide survey of 1992 (Ivanov et al., 1997a) for the description of CIW thermohaline structure after the extreme cooling of the water column took place. Such analysis is useful for identification of the penetration depth of winter convection in different regions of the sea. These generalised results of observations, rather than data from individual hydrographic stations, were then compared with the results of numerical simulations. Twelve typical stations have been chosen for the cyclonic area and convergence zone to characterise post-cooling conditions. For the shelf, all stations shallower than 100 m have been chosen for the analysis.

The six-hourly ECMWF meteorological data, which were used to force the model, were obtained for the open part of the sea (43°30′N, 32°E) and for the northwest shelf area (45°30′N, 32°E). The calculated heat fluxes were compared with the monthly mean heat fluxes obtained from the data on seasonal evolution of the water column thermal structure for the same locations.

Since one of our objectives was to reveal the spatial inhomogeneity of mixing properties, we have calculated the change in the potential energy (PE), accounting for changes in the position of the centre of gravity of a water column:

$$\Delta \mathrm{PE} = g\left[\left(\int_{z_1}^{z_2} \rho_i(z)\, z\, \mathrm{d}z\right) - \left(\int_{z_1}^{z_2} \rho_f(z)\, z\, \mathrm{d}z\right)\right],$$

where $\rho_i(z)$ and $\rho_f(z)$ are the initial and ensuing vertical density distributions, respectively. Alterations in potential energy have been calculated from the surface down to a certain depth or down to a certain density level.

3. Results

3.1. Pre- and post-cooling hydrography

The initial temperature profiles for the fall of 1992 were typical for the areas of cyclonic circulation, anticyclonic eddies and shelf area (Fig. 4). The thickness of the upper mixed layer was about 20–30 m. The depth of the CIL core varied from 30–40 m in the centre of the western cyclonic gyre to about 65

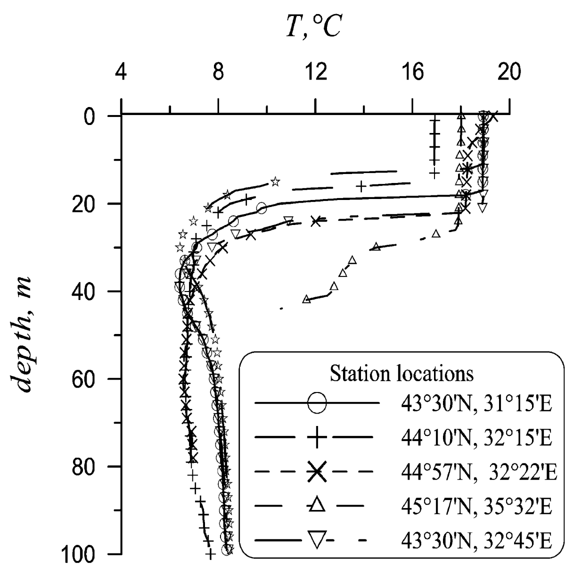

Fig. 4. Initial temperature profiles used for numerical simulations.

m in the Sevastopol anticyclonic eddy. The minimum temperature varied from 6.4 °C in the centre to 6.6 °C at the shelf break. The salinity of the CIL core was about 18.2–18.3 in the convergence zone and about 18.6 in the centre of the western basin (Ivanov, 1993). Over the shallower part of the shelf, the CIL was absent and temperature reached 10 °C at the bottom. Thus, initial hydrological conditions were typical for that part of the year (see Figs. 1 and 2). The Rim Current, which is at the same time a transition zone between the cyclonic domain and the convergence zone, was confined to the continental slope, approximately, along 44°10′N.

In the spring of 1993, the most noticeable circulation features for the northwest part of the basin were two anticyclonic eddies (the Sevastopol and Kaliakra eddy) where the main pycnocline, characterised here by the maximum vertical salinity gradient (Fig. 5), deepened to 140–150 m. These eddies were separated by a well-noticeable onshore meander of the Rim Current. In the centre of the western sub-basin, the pycnocline shoaled to about 35 m. On the average, the CIL core temperature, salinity and density were 5.91 °C (compared to $\sim$6.5 °C in the autumn of 1992), 18.57 and 14.61 kg/m^3. However, CIW characteristics essentially differed for the cyclonic and anticyclonic regions, similar to the situation in 1992 (Ivanov et al., 1997b).

The major portion of the newly formed CIW was stored in the convergence zone, where the CIL core showed features that were similar to those observed at the shelf. Statistical T,S diagrams for the anticyclonic and cyclonic circulation regions and the northwest shelf region in Fig. 6a,b indicate the origins of water masses. The volume of extremely cold (less than 5 °C) waters formed near the coast is shown to be small, when compared to the main body of water with temperature about 6.1 °C and salinity about 18.25. The volume of these rather homogeneous waters in the shelf region was about 25% of the total amount. The homogeneity of the shelf water mass indicates that the waters replenishing CIL in this region were formed under similar external conditions.

The volume of the newly formed CIW in the cyclonic region (salinity 18.6–18.8) was several

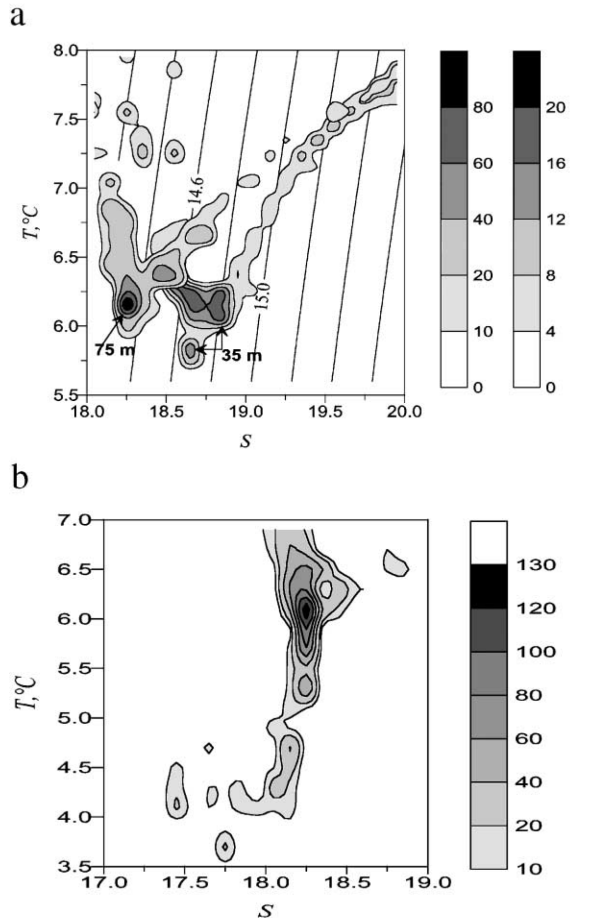

Fig. 6. Statistical T,S diagrams for the cyclonic and anticyclonic regions (a) and for the northwest shelf area (b).

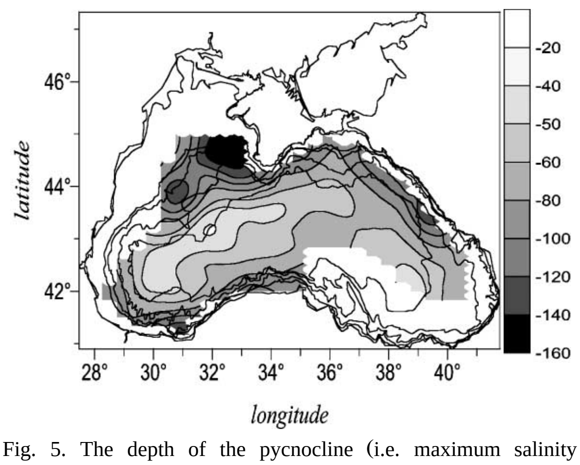

Fig. 5. The depth of the pycnocline (i.e. maximum salinity gradient) in April of 1993.

times less than the volume of water in the convergence zone (Fig. 6a). Waters with T,S indices typical of the cyclonic interior or the anticyclonic region do not appear to mix with each other. The density of the CIL core across the Rim Current (about 30–40 km wide) varied from 14.8 kg/m^3 to the left of the current to about 14.3 kg/m^3 in the convergence zone. There is some overlap of the T,S diagrams in Fig. 6a, but this overlap occurs only for the water with temperature of about 6.4 °C and salinity of 18.4–18.6, which lie above the CIL core in the cyclonic region, and below it in the anticyclonic region. A comparison of the depth of the CIL core with the halocline depth (see Fig. 5) reveals that the CIL core lies immediately above the pycnocline in the cyclonic regions, and several tens of meters above in the anticyclonic regions. This suggests that direct ventilation of the upper layers is limited by the pycnocline within the cyclonic gyres, but not in the convergence zone where salinity stratification and hence ventilation appear to be controlled by sub-basin scale dynamical processes including advection effects.

The CIL negative heat storage, defined as $Q = \int_l^u (8° - T(z))\mathrm{d}z$, reached a value of 220 °C m in the convergence zone in the spring of 1993, and was thus, approximately, 30% higher than in March 1992 (Ivanov, 1993). This comparison once again emphasises that the winter of 1993 was anomalous in terms of CIW formation. The response of the pycnocline to cooling at the surfaces was notably different for the cyclonic and anticyclonic regions. Cooling within the convergence zone was confined to a layer above a sigma-t value of 14.4 kg/m^3 (at 80 m depth) while it penetrated down to a depth of sigma-t 14.8 kg/m^3 (at 40 m depth) in the central part of the basin (Fig. 6a).

From the comparative analysis of pre- and post-cooling hydrographic conditions in the region, it appears that the winter of 1993 was anomalous in terms of cooling, leading to CIW formation. As a result, the minimum temperature in the sea dropped by 0.6 °C below the average for this part of the year, convection reached below the earlier core of the CIW from the previous winter, and replenished it in the central part of the sea. In the convergence zone, the replenishment was more complex, with traces of CIL core from the earlier period partly surviving until April. Values of minimum temperature observed in April were slightly lower in the central part of the western basin than over the northwest slope region due to advection of warmer waters from the east with the Rim Current.

3.2. Results of numerical simulations

The daily mean heat fluxes for the selected locations during the period from October 1992 to May 1993 varied in the range from −500 to 100 W/m^2 (Fig. 7a). The average heat flux at 45°30′N was −90 W/m^2, and it was −69 W/m^2 at 43°30′N. Days with the greatest heat losses occurred during storms in late October, beginning of January (two spells of stormy weather) and in February. In most of these days, heat losses at the northern location exceeded heat losses in the central part of the basin by about 40 W/m^2. On the other hand, during several cold days in November, heat losses in the central part of the western basin exceeded those over the shelf. The calculated heat losses in October–January of 1992–1993 significantly exceeded (by 50–60 W/m^2) monthly mean heat fluxes derived from the data on

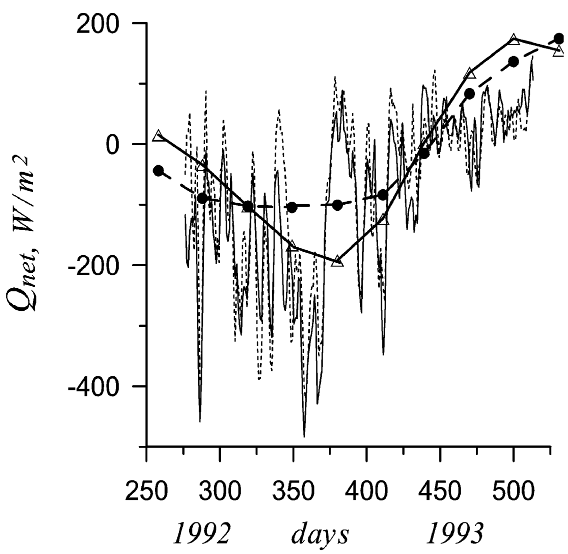

Fig. 7. (a) Time series of daily mean heat fluxes calculated for the October 1992–May 1993 period (3 days running average) at 43°30′N, 32°E (dashed line, anticyclonic region) and 45°30′N, 32°E (solid line, cyclonic region) and (b) reference monthly mean heat fluxes versus sigma-t density, calculated from historical seasonal temperature profile data for the same locations.

seasonal evolution of the water column at 43°30′N and at 43°30′N (Fig. 7b). After the second half of January 1993, the heat losses were smaller than those for an average year.

The prevailing CIL in the central part of the western basin was essentially eroded during two spells of cold air outbreaks, accompanied by storms (with wind speed of about 15 m/s) in the beginning of January and in February 1993 (Fig. 8a). The time scale of vertical mixing is estimated on the order of 2–3 days. The evolution of the thermal structure was predicted fairly well. For example, the resultant CIL core temperature in the central part of the sea dropped to less than 6 °C, i.e. the predicted minimum temperature was very close to the minimum temperature observed in April. Convective mixing reached a depth of 40 m (Fig. 8a), in very good agreement with the results of the *T*,*S* analysis for spring data (Fig. 6a). Remarkably, applying artificial changes introduced into the air temperature and wind speed did not notably increase the convection depth over the dome of the pycnocline in the cyclonic region, thus proving a robust prediction of penetration depth by the model in this region. For stations with deep initial level of pycnocline in the convergence zone, convective mixing did not penetrate much below a depth of about 50 m, and the newly produced CIW overlay the remnants of the CIW from the previous year (Fig. 8b). The multiple episodes of mixing immediately homogenised the upper ocean, but the temperatures produced in early winter (December) were considerably warmer than those in the cyclonic region. Only after repeated pulses of cooling in February were the minimum temperatures attained in the convergence zone (Fig. 8a,b).

In the coastal region further north, at the shelf break and over the shelf, the mixed layer penetrated deeper, to about 70 m, as a result of the more severe weather conditions in this region (Fig. 8c). In this region, the upper part of the remnant CIL from the

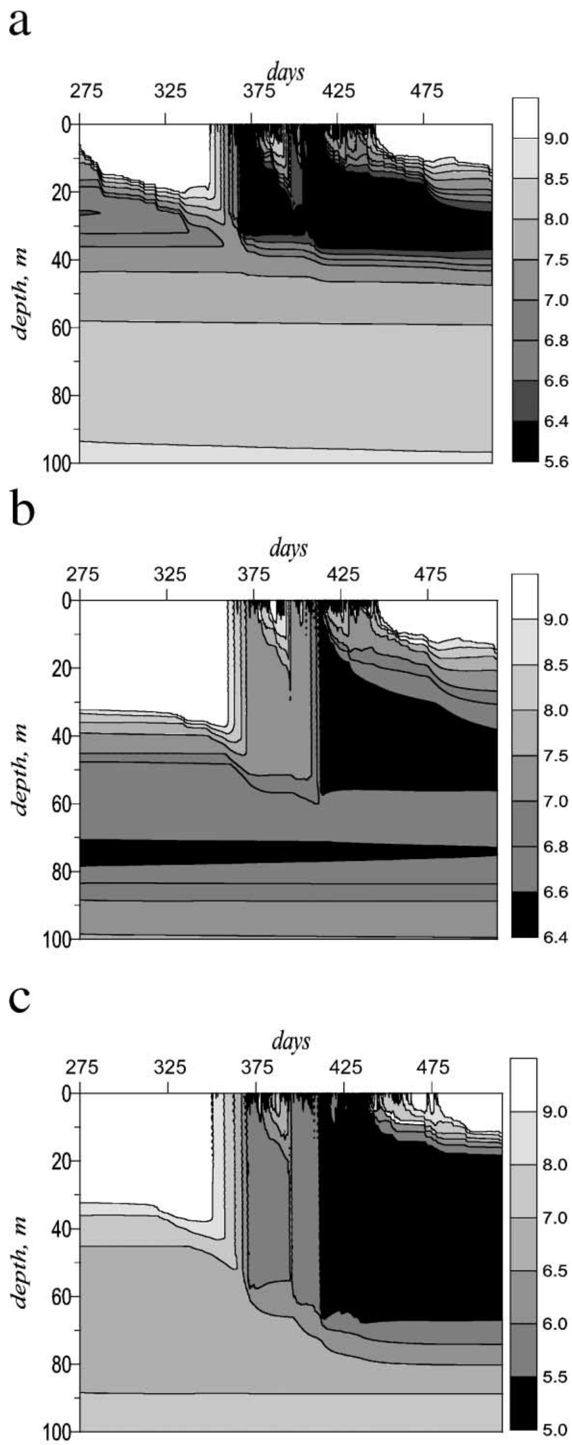

Fig. 8. Predicted evolution of the upper layer thermal structure for stations in the centre of the western basin (a), in the convergence zone (b), at the shelf (c) over the period from October 1992 to May 1993.

previous year was completely eroded, and the minimum temperature dropped to less than 5.5 °C. This last result is in discord with the observations, which indicate minimum temperature in this region in April to be slightly higher than that in the centre of the basin. On the other hand, the penetration depth of convection is in agreement with the results of spring observations (Fig. 6a). We believe that the mismatch between the results of simulations and observations can be explained by the importance of horizontal advection, as inferred from the climatological temperature distribution (see Fig. 3). The Rim Current advects warmer water from the east onto the northwest shelf. This diminishes the role of horizontal inhomogeneity in the meteorological fields in controlling CIL temperature. In all of our experiments, convection had limited effect in modifying the thermal and density structure of the pycnocline, as illustrated by Fig. 9 comparing initial and resultant temperature profiles from observations with those resulting from the numerical simulation at a station within the western cyclonic gyre. As convection is the only mechanism of vertical mixing in the model, it does not affect the properties at the pycnocline. It should be noted that we were able to reproduce the observed changes in the pycnocline thermal structure

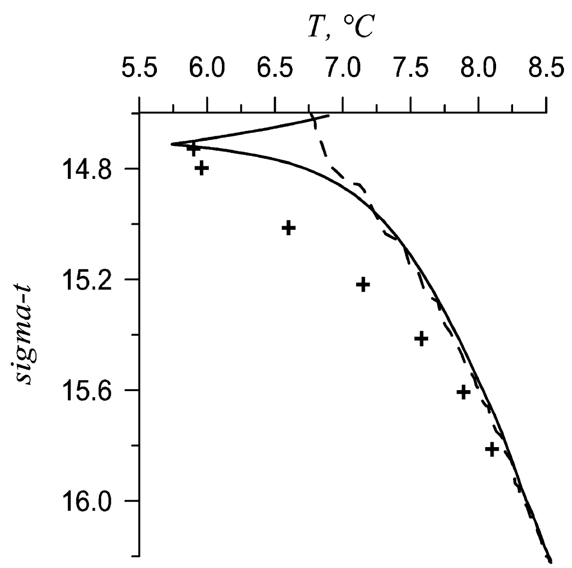

Fig. 9. Initial (dashed), predicted (solid) and observed in the post-cooling period (crosses) temperature distribution versus sigma-t for the central part of the western basin.

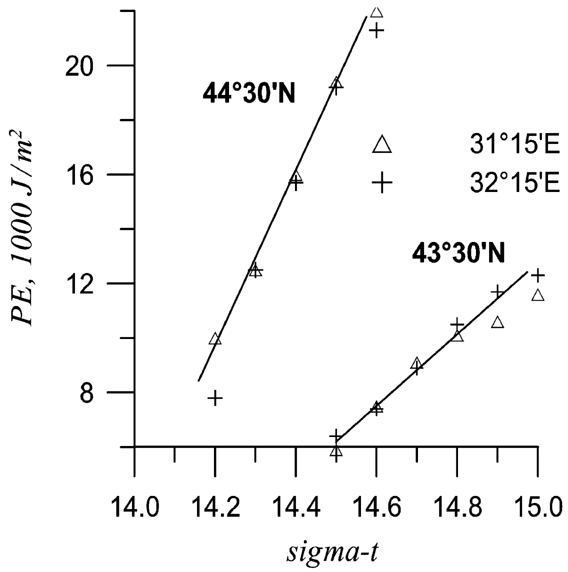

Fig. 10. Changes in the potential energy of the water column based on the initial stratification of October 1992 as a result of convective mixing (calculated between selected sigma-t surfaces) in the western cyclonic gyre (43°30′N) and in the convergence zone (44°30′N). Data points from stations located at two different longitudes show similar trend at the same latitudes.

only by artificially increasing background diffusivity in the pycnocline to 0.5 cm^2/s. One might thus conclude that processes other than convection are responsible for the observed diapycnal mixing at the pycnocline level.

3.3. Estimates of ΔPE caused by convective mixing

As revealed in Section 3.1, the CIL core in April 1993 was found at 30–40 m in the central part of the basin and at 70–80 m depth in the area of anticyclonic eddies (Fig. 6a). Numerical simulations in Section 3.2 have confirmed that winter convection could reach down to these depths. Similar results could be predicted by estimating the amount of energy spent for mixing, expressed as the change in the potential energy of the water column. Convection penetrates down to 14.8 kg/m^3 sigma-t level in the centre of the western cyclonic gyre and down to 14.2 kg/m^3 sigma-t within the convergence zone if 10×10^3 J/m^2 were to be spent, based on an initial density stratification of the autumn of 1992 (Fig. 10). Therefore, inhomogeneities in CIL density and thickness could result from spatial differences in initial stratification, even when a spatially uniform forcing was applied.

In reality, weather conditions were, of course, more severe in the northwest shelf region, and therefore convective mixing in this area penetrated down to about 14.3 kg/m^3 sigma-t surface, or, approximately, 10 m deeper than 14.2 kg/m^3 sigma-t surface. It should be also mentioned that the position of a particular sigma-t surface found in spring could notably differ from its position in February, after the mixing takes place. This is especially true for the area of Sevastopol anticyclonic eddy where the pycnocline deepens as cooling at the surface yields further CIW replenishment (Stanev and Staneva, 1997). Summarising, one may conclude that the increase in potential energy of the water column, as a result of convective mixing in the winter of 1992–1993, should constitute about 10×10^3 J/m^2 for the central part of the western cyclonic gyre and up to 13×10^3 J/m^2 for the area of Sevastopol anticyclonic eddy.

4. Summary

In comparison with earlier studies (Ivanov et al., 1997b), the results of statistical T,S analysis applied here for the data collected in spring of 1993 confirmed basic conclusions derived from the 1992 data. Indeed, temperature/salinity indices of CIW within the convergence zone were in distinct difference from the central cyclonic gyres, suggesting effective CIW formation within anticyclonic eddies under certain conditions. On the other hand, the analysis of the post-cooling-conditions data alone could not provide understanding of the time scales of convection events and describe evolution of the thermal structure of the water column in winter when observations are scares.

The analyses of hydrographic data and the results of numerical simulations with a mixed layer model revealed various space-time scales on which the pycnocline erosion takes place. The existing CIL was essentially eroded during two spells of cold air outbreaks when daily mean heat losses from the sea surface reached 400–500 W/m^2 under extreme cooling conditions of the studied period. The time scales of the mixing events are estimated to be on the order of 2–3 days. Convective mixing reached down to the main pycnocline only in the central part of the sea over the dome of the pycnocline. Within the convergence zone, the lower boundary of the mixed layer did not reach the main pycnocline. This fact indicates that the general feature of salinity stratification, the main halocline, limits direct ventilation of the upper layers within the cyclonic gyres but not in the convergence zone. In this area, the effects of direct ventilation are limited to a layer in which salinity stratification is controlled by sub-basin scale processes. The results reveal that, even with a spatially uniform forcing, a typical inhomogeneity in the CIL density and thickness could be formed due to spatial differences in the existing stratification of the upper layer of the Black Sea.

In the Black Sea, due to perennial existence of the CIL, the upward heat flux out of the pycnocline is limited to short winter periods when the layer is exposed to atmospheric forcing. In other words, the largest fluxes occur when the CIL core is completely eroded and convection reaches the upper boundary of the pycnocline in the centres of the large-scale cyclonic gyres. Hydrographic data of recent basin-wide monitoring surveys allow us to make estimates of the magnitude of heat losses from below the level of the temperature minimum. An assessment of the basin averaged changes of heat content below the CIL core during the winter of 1992–1993 (Ivanov et al., 1997a) gives an estimate of 6×10^7 J/m^2. At the same time, negative heat fluxes at the sea surface during winter storms were 300–500 W/m^2. In that manner, the time scale of atmospheric impact on the main pycnocline in the winter of 1992–1993 is estimated to be about 4 days if we were to assume that the convective mixing reached the pycnocline over 50% of the area of the basin. It is worthy of note that the latter estimate is in full agreement with the results of numerical experiments showing that the upper pycnocline was essentially eroded during two cold spells.

The results of numerical experiments show that convection, even in confined areas, cannot erode the pycnocline indefinitely, as the salinity gradient is too strong. In all our numerical experiments, convection itself had limited effect in modifying thermal and density structure of the pycnocline layer. This suggests that the observed evolution of the thermal structure of the pycnocline in winter was basically a result of diapycnal mixing, and that other processes were responsible for diapycnal mixing in the pycnocline.

One of the results of numerical experiments conducted in the framework of this study provides clear evidence of overlying cold cores that may take place within the convergence zone when convection does not erode the underlying remnant CIL core. It is worth mentioning that such evidence was found during conditions favourable for CIW formation. For a mild winter, such overlaying could be even more notable. Another source of interleaving could be isopycnal intrusions formed within the transition zone between the cyclonic gyres and convergence zone.

Acknowledgements

The paper has been prepared in the framework of the INCO Copernicus Project 'Ventilation of the Black Sea Anoxic Layer' (IC15CT96-113). Partial support from NATO through the SfP ODBMS-Black Sea Project is also acknowledged.

References

Altman, E.N., Simonov, A.I. (Eds.), Hydrometeorology and hydrochemistry of USSR seas. The Black Sea, vol. IV Gidrometeoizdat, St. Petersburg, 428 pp. (in Russian).

Backhaus, J.O., 1985. A three-dimensional model for the simulation of Shelf Sea dynamics. Dtsch. Hydrogr. Z. 38, 165–187.

Belokopytov, V., 1998. On hydrometeorological conditions of Cold Intermediate Water renewal in northern part of the Black Sea. In: Ivanov, L., Oğuz, T. (Eds.), Ecosystem Modelling as a Management Tool for the Black Sea. NATO ASI Ser., vol. 2. Kluwer Academic Publishers, Dordrecht, pp. 47–52.

Blatov, A.S., Bulgakov, N.P., Ivanov, V.A., Kosarev, A.N., Tuzhilkin, V.S., 1984. Variability of Hydrophysical Fields of the Black Sea. Hydrometeoizdat, Leningrad, 240 pp. (in Russian).

Dobson, F.W., Smith, S.D., 1988. Bulk models of solar radiation at sea. Q. J. R. Metereol. Soc. 114, 165–182.

Filippov, D.M., 1965. The cold intermediate layer in the Black Sea. Okeanologiya 5, 47–52.

Filippov, D.M., 1968. Circulation and Structure of Black Sea Waters. Nauka, Moscow, 136 pp. (in Russian).

Friehe, C.A., Schmitt, K.F., 1976. Parameterisation of air–sea interface fluxes of sensible heat and moisture by the bulk aerodynamic formulas. J. Phys. Oceanogr. 6, 801–809.

Gill, A.E., 1982. Atmosphere Ocean Dynamics. Academic Press, San Diego, CA.

IAPSO Working Group, 1981. The practical salinity scale 1978 and the international equation of state of seawater 1980, UNESCO Technical Papers in Marine Sci.: 36, 124, UN. Educ. Sci. and Cult. Organ., Paris.

Ivanov, L.I., 1993. Cold Intermediate Water formation in the western Black Sea as derived from 1992 data. In: Eremeev, V.N. (Ed.), Problems of the Black Sea. Marine Hydrophysical Institute NASU, Sevastopol, pp. 78–89.

Ivanov, L.I., Besiktepe, S., Özsoy, E., 1997a. Physical oceanography variability in the Black Sea pycnocline. In: Mikaelyan, A., Özsoy, E. (Eds.), Sensitivity to Change: Black Sea, Baltic Sea and North Sea. NATO ASI Ser., Kluwer Academic Publishers, Dordrecht, pp. 265–274.

Ivanov, L.I., Besiktepe, S., Özsoy, E., 1997b. Black Sea cold intermediate layer. In: Özsoy, E., Mikaelyan, A. (Eds.), Sensitivity to Change: Black Sea, Baltic Sea and North Sea. NATO ASI Ser., Kluwer Academic Publishers, Dordrecht, pp. 253–264.

Ivanov, L.I., Konovalov, S.K., Belokopytov, V., Özsoy, E., 1998. Regional peculiarities of physical and chemical responses to changes in external conditions within the Black Sea pycnocline: cooling phase. In: Ivanov, L., Oğuz, T. (Eds.), NATO TU-Black Sea Project: Ecosystem Modelling as a Management Tool for the Black Sea, Symposium on Scientific Results. NATO ASI Ser., vol. 2. Kluwer Academic Publishers, Dordrecht, pp. 53–68.

Knipovich, N.M., 1932. The hydrological investigations in the Black Sea. Tr.-Azovo-Chernomorsk. Nauchnopromyslovoy Eksped. 10, 274 pp. (in Russian).

Kolesnikov, A.D., 1953. Seasonal course of temperature, stability and vertical turbulent heat exchange in the open part of the Black Sea. Tr. Morsk. Gydrophys Inst., USSR AS 3, 3–13 (in Russian).

Kochergin, V.P., 1987. Three-dimensional prognostic models. In: Heaps, N. (Ed.), Three-dimensional coastal ocean models. Coastal and Estuarine Sci., vol. 4. American Geophysical Union, Washington, DC, pp. 201–208.

Ovchinnikov, I.M., Popov, Yu.I., 1987. Evolution of the cold intermediate layer in the Black Sea. Okeanologiya 27, 555–560.

Oğuz, T., Latun, V.S., Latif, M.A., Vladimirov, V.V., Sur, H.I., Markov, A.A., Özsoy, E., Kotovshchikov, B.B., Eremeev, V.V., Ünlüata, U., 1993. Circulation in the surface and intermediate layers of the Black Sea. Deep-Sea Res. 40, 1597–1612.

Özsoy, E., 1999. Sensitivity to global change in temperate Euro-Asian Seas (the Mediterranean, Black Sea and Caspian Sea): a review. In: Malanotte-Rizzoli, P., Eremeev, V.N. (Eds.), The Eastern Mediterranean as a Laboratory Basin for the Assessment of Contrasting Ecosystems. NATO Science Ser. 2, Environmental Security, vol. 51. Kluwer Academic Publishers, Dordrecht, pp. 281–300.

Özsoy, E., Ünlüata, U., 1997. Oceanography of the Black Sea: a review of some recent results. Earth-Sci. Rev. 42, 231–272.

Schrum, C., Backhaus, J.O., 1999. Sensitivity of atmosphere–ocean heat exchange and heat content in North Sea and Baltic Sea. Tellus 51A (4), 526–549.

Stanev, E.V., Staneva, J.V., 1997. Cold Intermediate Water formation in the Black Sea. Analysis on numerical model simulations. In: Özsoy, E., Mikaelyan, A. (Eds.), Sensitivity to Change: Black Sea, Baltic Sea and North Sea. NATO ASI, 2/27, Kluwer Academic Publishers, Dordrecht, pp. 375–394.

Tolmazin, D., 1985. Changing coastal oceanography of the Black Sea: Part I. Northwestern shelf. Prog. Oceanogr 15, 217–276.

Zubov, N.N., 1938. Marine Waters and Ice. Hydrometeoizdat, Moscow, 451 pp. (in Russian).