Ventilation of the Black Sea pycnocline. Parameterization of convection, numerical simulations and validations against observed chlorofluorocarbon data

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Abstract

Data from field observations and numerical model simulations are used to understand and quantify the pathways by which passive tracers penetrate into the Black Sea intermediate and deep layers. Chlorofluorocarbon (CFC) concentrations measured during the 1988 R.V. Knorr cruise show strong decrease with increasing density in the Black Sea and illustrate the very slow rate of ventilation of deep water in this basin. We develop a 3D numerical model based on the Modular Ocean Model (MOM), and calibrate it in a way to produce consistent simulations of observed temperature, salinity and CFCs. One important feature is the implementation of a special parameterization for convection, which is an alternative of the convective adjustment in MOM and handles the penetration of the Bosporus plume into the halocline. The model forcing includes interannually variable wind, heat and water fluxes constructed from Comprehensive Ocean–Atmosphere Data Set and ECMWF atmospheric analysis data and river runoff data. The analysis of observations and simulated data are focused on correlations between thermohaline and tracer fields, dynamic control of ventilation, and the relative contributions of sources at the sea surface and outflow from the Bosporus Strait in the formation of intermediate and deep waters. A simple theory is developed which incorporates the outflow from the strait along with the vertical circulation (vertical turbulent mixing and Ekman upwelling) and reveals their mutual adjustment. The analyses of simulated and observed CFCs demonstrate that most of the CFC penetrating the deep layers has its source at the sea surface within the Black Sea rather than from the Marmara Sea via the Bosporus undercurrent. Under present-day conditions, the surface CFC signals have reached only the upper halocline. Intrusions below 600 m are not simulated. The major pathways of penetration of CFCs are associated with cold-water...
1. Introduction

Estimates of the ventilation rates and residence times of intermediate and deep layers of the Black Sea vary over a large range, and depend on the conceptual model and data sets used, as well as their associated errors (Boudreau and Leblond, 1989; Murray et al., 1991; Buesseler et al., 1991). Utilizing new and complimentary data and methods should increase the credibility of estimates, as demonstrated in the recent paper by Lee et al. (2002, hereafter LBMS), which presented chlorofluorocarbons (CFC) data from the 1988 Knorr cruise (Fig. 1). In the present paper we move one step further, addressing the above problems in a dynamically consistent way.

The vertical stratification in the Black Sea is extremely stable with stagnant (anoxic) conditions in the deep layers. Exchange of water between the Black Sea and Sea of Marmara is constrained by the two-layer exchange through the shallow Bosporus Strait. The stratification in the Black Sea is controlled by the fresh water fluxes from rivers, precipitation and evaporation, and influx of high-salinity water from the Sea of Marmara. The large net fresh water flux (~300 km³ yr⁻¹) to the Black Sea (which has a basin volume of ~540000 km³) and the narrow opening in the strait (0.7–3.5 km wide and ~35 m deep) results in an asymmetric exchange between the Black and Marmara Seas: the volume transport by the outflowing surface current is two times larger than the inflowing deep higher-salinity countercurrent. This asymmetric transport results in surface salinity values in the Black Sea (S = 18.5) that are about half those in the Mediterranean Sea (S = 37). The excess fresh water flux to the sea surface maintains a very strong stratification, which results in reduced vertical exchange. This is one of the major prerequisites leading to formation and maintenance of the Cold Intermediate Layer (CIL) and the suboxic and anoxic layers. A number of direct (e.g., Gregg and Özsoy, 1999) and indirect (e.g., Lewis and Landing, 1991) estimates have demonstrated that the vertical mixing coefficient in Bosporus inflow area, as well as in the basin interior is close to the background value in laminar flows (~10⁻⁶ m² s⁻¹). The reduced vertical exchange shields the CIL from mixing with warm surface water in summer (Friedrich and Stanov, 1988; Stanov, 1990) and makes the penetration rate of signals from the sea surface into the deep layers much slower than observed in the neighboring better-ventilated Marmara Sea (Besiktepe et al., 1993; LBMS).

The entrainment of CIL water by the inflowing Bosporus plume is the key mechanism that enables a mixture of surface and intermediate waters (in much larger quantities than the inflow from the strait) to reach the pycnocline (Oguz et al., 1990; Murray et al., 1991; Buesseler et al., 1991; Konovalov and Murray, 2001). The small magnitude of the inflow (annual mean value of ~300 km³), as well as the large amount of mixing on the shelf, result in a rapid decrease of the salinity of the inflowing plume, which makes the
effluent almost indistinguishable from the ambient water on the continental slope (Gregg et al., 1999; Di Iorio and Yuce, 1999). Thus most of the entrainment occurs before the Bosporus plume reaches the shelf break. Experimental data elucidating the characteristics of the plume are very limited and the indirectly estimated entrainment ratios vary over a large range (~1–10), depending mostly on the models and data used, as well as on the depth range studied (Boudreau and Leblond, 1989; Murray et al., 1991; Buesseler et al., 1991).

Our goal in this paper is to develop a 3D numerical model capable of adequately simulating physical processes, as well as the evolution of transient tracers in the Black Sea. To do this we use recent CFC measurements (Bullister and Lee, 1995; LBMS) to calibrate the numerical model in a way to produce consistent simulations. We then use this model to derive knowledge about mixing pathways and processes and to quantify their relative contribution to the ventilation of Black Sea subsurface waters. For example, an estimate of the rate of replenishment of the CIL is difficult to obtain from observations, and is therefore not well known. Another example is the ultimate penetration depth of Bosporus intrusions, for which there is no clear consensus in previous studies (e.g., Ostlund, 1974; Murray et al., 1991; Top et al., 1991; Özsoy et al., 1993; Bullister and Lee, 1995).

In this study we will address the following topics:

1. Parameterization of the outflow from the Bosporus Strait.
2. Vertical circulation.
3. Water mass formation (pathways and rates of penetration of heat, salt and tracers into the intermediate and deep levels) identified from the correlation between CFC data and CIL/halocline characteristics.
4. Do the model simulations give clear and resolvable predictions of interannual variability in the water mass structure of the Black Sea, as recently observed in the neighboring Mediterranean Sea (Roether et al., 1996)?
5. Have there been substantial changes in the vertical stratification of the Black Sea over the past 50 years triggered by interannual variability in atmospheric forcing?
6. Are the subsurface maxima of CFC, found by LBMS perennial features, or are they seasonally dependent?
7. What is a reasonable amount of data required from different depth (density) levels to efficiently develop a more informative database?
8. To derive estimates on the precision of measurements required to trace the mixing paths of CFC in the halocline.

The paper is structured as follows. We first present the data used, then the model development and model scenarios. The major part of the paper is devoted to analysis of the simulations and their comparison with observations.

2. Data

In this paper, we use both surface and subsurface data from the Black Sea. The surface data consist of observed and reanalyzed meteorological fields, as well as reconstructions of the atmospheric concentrations of CCl₃F (CFC-11) and CCl₂F₂ (CFC-12) for the period 1950–1988. The beginning of this period coincides with the time when the CFCs began to accumulate in the atmosphere in significant quantities. The end of the period coincides with the 1988 Knorr cruise when deep-water CFC data were first collected in the Black Sea (Bullister and Lee, 1995; LBMS). The subsurface data include mean temperature and salinity profiles needed to initialize the circulation model (Staneva and Stanev, 1998).

2.1. Meteorological data

The data used to force the circulation model described in Section 3 include monthly mean heat and water fluxes as well as the wind stress for the period 1950–1990. The heat flux data are based on the regional climatic data. After 1964, we use the Comprehensive Ocean–Atmosphere Data Set (COADS data, Woodruff et al., 1987), from which we compute the fluxes. The corresponding data series give us the low-frequency time variability for
the period 1950–1990. It is well known that using only climatological data to calculate fluxes can give large inconsistencies due to lack of the synoptic signal. This high-frequency signal can be obtained from the ECMWF reanalysis data. We extracted from this data set the mean high-frequency signal which characterizes the synoptic variability in the atmosphere for 1979–1993 and added it to the COADS data series. We then calculated the heat flux using data for air temperature and relative humidity, SST and surface winds, as well as the bulk aerodynamic formulae described by Stanev et al. (1997). To check the consistency of heat fluxes based on the hybrid COADS-ECMWF data, we compared the monthly fluxes calculated from the ECMWF 6-hourly data, with those calculated from the hybrid data, as well as the estimates based only on the COADS monthly climatology (Fig. 2) for the period 1979–1993. All data in this figure are presented as basin-wide monthly means. Obviously, the fluxes based on the hybrid data do not perfectly match the values calculated using ECMWF data only, but are closer to them than the fluxes calculated using only COADS data. The meteorological fields over the Black Sea exhibit large spatial variability; however, the available COADS data are too coarse to resolve this. To solve this problem we use the horizontal heat flux patterns from the monthly mean forcing data (Staneva and Stanev, 1998). Because it is not a priori clear whether these patterns match climatic patterns from the ECMWF data we analyzed the two data types, as well as individual monthly mean patterns from the ECMWF data. It appeared that the changes in the patterns from one year to another (e.g., January 1989, January 1990) are not too large, and both patterns are close to the climatology, thus repeating the same forcing patterns (from climatic data) every year does not result in large errors in the model forcing. The lack of monthly sea surface data with sufficient spatial resolution for the years before the reanalysis of ECMWF was initiated, and the need to have the same type of forcing throughout the integration resulted in our decision to repeat the same patterns every year. It is noteworthy that the basin mean heat exchange for the whole period obtained from the spatially and temporally variable surface fluxes matches the climatological basin mean heat flux estimates. There is interannual variability in the forcing functions (Fig. 3a); however, this signal is not accounted for before 1964 because the COADS data are too sparse for these earlier times and do not correctly resolve this variability. The spatial patterns of the annual mean heat flux and wind stress curl are shown in Fig. 4.

The temporal variability in the Black Sea fresh water balance (river runoff, precipitation and evaporation) for 1950–1985 was taken from the monthly mean data of Simonov and Altman (1991) and Altman and Kumish (1986). For the later period (1986–1990), these data were kindly provided by V. Belokopitov. The anomalies of the fresh water flux for 1950–1990 are shown in Fig. 3b. The horizontal patterns of fresh water flux at the sea surface (Fig. 4c) were taken from the climatic data of Staneva and Stanev (1998).

2.2. Atmospheric CFC data

The production of CFC-11 and CFC-12 and release of these gases to the atmosphere began in the 1930s. Atmospheric concentrations (as mole fraction in air) of CFC-11 and CFC-12 have been monitored worldwide since 1978. Prior to this
time, atmospheric concentrations can be reconstructed based on estimates of production and release of CFCs to the atmosphere (McCarthy et al., 1977), with corrections due to losses by stratospheric photolysis processes. Models of CFC-11 and CFC-12 concentrations (in parts-per-trillion-ppt) in the troposphere of the Northern Hemisphere from 1950–1988, based on the compilation of Walker et al. (2000) are show in Fig. 3c. Atmospheric concentrations prior to 1950
were extremely low, and dissolved CFC concentrations at equilibrium with the atmosphere during this period would be at or below the analytical detection limits for these compounds (Bullister and Weiss, 1988). In our model simulations, the atmospheric CFC concentrations prior to 1950 (the spin-up phase) are considered to be zero.

Air measurements of CFC-11 and CFC-12 were made during the 1988 Black Sea expedition, and found to be several percent higher than those predicted by Walker et al. (2000) for the mid-latitude Northern Hemisphere for this period. These small offsets may be due to calibration differences and/or to the proximity of this region to sources of CFC release in Europe. In this study, to create models of the atmospheric history of CFC-11 and CFC-12 over the Black Sea, the mid-latitude Northern Hemisphere trends of atmospheric CFC-11 and CFC-12 shown in Fig. 3c are normalized to atmospheric concentrations measured during the Black Sea expedition in 1988.

2.3. Water Column CFC data

Input of CFCs into the Black Sea occurs primarily by transport across the air–sea interface and by transport through the Bosporus Strait. At equilibrium, dissolved CFC concentrations in near-surface water are simple functions of the temperature and salinity of the water (Warner and Weiss, 1985) and the CFC concentrations in the overlying atmosphere (Fig. 3c). However, dissolved CFC concentrations as a function of time in subsurface waters entering the Black Sea through Bosporus Strait are highly unknown. This motivated us to analyze two typical profiles taken in 1988: one in the Black Sea (Knorr station 14) and one in the Marmara Sea (Knorr station 1). There are strong salinity stratifications in both seas (Fig. 5a). The salinity in the deep layers of the Marmara Sea (~38.5) and the Black Sea (22.3) are quite different demonstrating completely different water masses. Since the density stratification in both the Black Sea and Sea of Marmara is mainly determined by the variations in salinity (rather than temperature), the profiles in Fig. 5a are thus indicative that the stratification is very stable in both seas.

Differences in the ratio of inflow to basin volume can explain why the Black Sea subsurface waters are anoxic, while those in the Sea of Marmara contain relatively high levels of oxygen. Surface and deep layer volumes of the Marmara Sea are 230 and 3148 km³ (Besiktepe et al., 1993), respectively, with corresponding residence times of 6 months and 6–7 years. Thus the entire Marmara Sea is filled with relatively young water, with relatively high dissolved oxygen levels. The rapid flushing of the deep water explains why the deep profile of salinity is so homogenous with values typical of the Mediterranean Sea. The CFC concentration profiles (all dissolved CFC concentrations are expressed as pmol kg⁻¹, where 1 pmol = 1 picomoles = 10⁻¹² mole) in Fig. 5b illustrate the relatively vigorous deep-water ventilation in the Sea of Marmara. The values at 400m depth are about 40% those in near-surface waters (dashed line in Fig. 5b). In contrast, the ratio of inflow to basin volume in the Black Sea is very low (with an exchange time > 1000 years) leading to the development of stagnant conditions in the subsurface waters. The slow renewal processes prevent surface-derived CFC from effectively penetrating into the deep Black Sea. Thus the CFC-12 concentrations at 400m are less that 1% of surface values (— in Fig. 5b).

The unique distribution of the thermohaline fields in the two neighboring seas is shown as salinity versus density profiles in Fig. 5c. Below ~60 m, the linear relationship between salinity and density in the Black Sea reflects the negligible contribution of temperature to the formation of density anomalies. The two curves reflect the main mixing paths and the end-members. The slope of the Black Sea data tends toward the deep Marmara Sea value, which is the main mixing path between the waters of the two basins.

CFC-12, which is a conservative tracer, shows quite different behavior in the subsurface layers of the two basins (Fig. 5d). CFC-12 concentrations decrease sharply with increasing density in the Black Sea, demonstrating the poor ventilation conditions in the deep basin. In the Marmara Sea, the deep water is rich in CFC-12 reflecting the relatively rapid ventilation from the Dardanelles.
Note that the observations in the Marmara Sea were only to ~450 m, which is above the sill depths separating the three sub-basins of this sea (Besiktepe et al., 1993). The vertical profile of CFC-12 in Fig. 5d shows an increase in concentration in the two deepest levels sampled and supports the results of this work, which suggests that the deep oxygen maximum observed in the Sea of Marmara is a consequence of the rapid deep-water ventilation.

From this analysis of the differences in the water mass characteristics between the two seas we see that, unlike the Marmara Sea, the CFC signals penetrating into the deep Black Sea are extremely small. This makes their observation challenging and requires careful model parameterization in order to obtain correct simulations.

3. The model

3.1. General description, resolution and initialization

The circulation model for the Black Sea is derived from the model of Stanev et al. (1997), which is based on the Modular Ocean Model (MOM, Pacanowski et al., 1991). This model was used previously by Staneva et al. (1999) to study the penetration of $^{137}$Cs from bomb fallout and the
Chernobyl accident. The present work differs from this previous study, not only by the different tracers used, but also by three major improvements in the model: (1) The forcing here includes interannually variable wind, heat and water fluxes constructed from COADS data ECMWF atmospheric analysis data and river runoff data. (2) The parameterization of upper layer mixing and, in particular the mixing between Mediterranean and Black Sea waters, is investigated in more detail, and calibrated in a way to adequately match the CFC data. (3) Initial boundary conditions for the CFC tracers are utilized which are consistent with their penetration pathways into the Black Sea. The rest of the model setup remains the same as in Staneva et al. (1997) and Staneva et al. (1999) and therefore it is only briefly described. In Section 3.3 and Appendix A, we give more details for the treatment of diapycnal mixing.

The model was set up with a horizontal resolution of 1/4° in latitude and 1/3° in longitude and has 24 vertical levels, with model depths at 2.5, 7.5, 12.5, 17.5, 25, 35, 45, 55, 65, 75, 85, 105, 140, 185, 240, 310, 400, 515, 665, 870, 1145, 1470, 1820, 2125 m and realistic bottom topography. Solid side boundaries are non-slip and insulating for temperature and salinity. The bottom boundary is free slip and insulating.

The circulation model is initialized from a state of rest with horizontally homogeneous temperature and salinity fields taken from the basin mean vertical profiles. The spin-up phase of integration is carried out for 20 years with cyclic forcing until steady state is reached. The integration is then continued for the period 1950–1990 with variable in time forcing as described in Sections 2.1, 2.2 and 3.2 (see Fig. 3).

### 3.2. Model forcing

The physical part of the model is forced by atmospheric wind stress, heat flux, and water flux:

\[
\rho_0 A_v U_{h_0} = \tau, \quad \rho_0 C_p K_v T_z = q^T, \quad K_v S_z = q_w S^0, \tag{1}
\]

where \( A_v \) and \( K_v \) are the coefficients of vertical turbulent diffusion for momentum and heat/salt, respectively, \( \tau \) is the wind stress, \( q^T \) is the net surface heat flux, \( q_w \) is the water flux at sea surface, \( S^0 \) is a constant salinity value (taken as \( S = 18 \) in our experiments, which is the surface salinity value). A correction is introduced in the salinity forcing to weakly relax the surface values towards climatology. The corresponding flux resulting from the Newtonian relaxation is accounted for in the surface flux term, so that the water flux at the surface, when a mixed boundary condition is used, is equal to the water flux in the forcing data.

### 3.3. Subgrid parameterizations

#### 3.3.1. Mixing and diffusion

Mixing and diffusion in the horizontal are parameterized with biharmonic operators. The coefficients are: \( A_h = K_h = 0.8 \times 10^{19} \text{cm}^4\text{s}^{-1} \). The vertical mixing parameterizations include parameterizations of the surface mixed layer and vertical diffusion. The mixed layer model is based on the formulation of Gill and Turner (1976), relating the changes in the potential energy of water column to the rate of working of wind at the sea surface (Mitchel et al., 1985; Gordon and Bottomley, 1985). The application to the Black Sea is described in Staneva et al. (1998).

The coefficient of vertical exchange (for momentum) is \( A_v = 1.5 \text{cm}^2\text{s}^{-1} \). The vertical diffusion coefficient for heat and salt \( K_v \) is stability dependent

\[
K_v = \frac{a}{N^2}, \tag{2}
\]

where \( N \) is the Väisälä frequency, \( a = 0.004 \text{cm}^2\text{s}^{-2} \) (Gargett, 1984). This parameterization is tuned against the independent data of Lewis and Landing (1991) based on measurements of manganese and iron in the Black Sea. Details on model sensitivity to subgrid parameterizations are given in Stanev et al. (1997).

#### 3.3.2. Convection

In case of convective instability, the convective adjustment procedure in MOM is activated. This is a standard procedure in this model, in particular when the instability originates from the sea surface (for more details about convective adjustment, see...
The final result is that the unstable layer is homogenized to depth \( h \), which can be calculated from the following equation:

\[
h = \frac{1}{\rho_h} \int_{z_1}^{h} \rho \, dz,
\]

where \( z_1 \) is the depth where dense water occurs (e.g., sea surface). Thus the new density in the unstable layer is homogenized to depth \( h \) can be calculated from the following equation:

\[
\frac{r}{h} Z \quad \text{is}
\]

The salt balance in the model is closed at the Bosporus Straits by continuously adding a positive salinity flux. This flux is calculated from the diagnosed salinity flux at the sea surface, using a simple physical model for water exchange in the Straits (Stanev et al., 1997) fitted to observations by Unluata et al. (1989) and model data of Oguz et al. (1990). The positive salinity flux results in an increase of salinity (density) below the sill depth, leading to vertical instability. In natural conditions this has to trigger down-slope convection, however, according to Winton et al. (1998), the resolution in our model is not sufficient to resolve this process because of the very steep slope (the depth changes from 100 to 1000 m in less than 50 km). For a vertical resolution of 20 m the desirable horizontal resolution would be \( \sim 1 \) km, which is \( \sim 20 \) times smaller than our model grid. This means that the model outflow “sees” the topography as an almost vertical wall which is deeper than the maximum depth of penetration of the Mediterranean plume. For such extreme slopes and insufficient horizontal resolution, the parameterization of Beckmann and Döescher (1997) could not be successfully applied because it works with the density within the bottom-most tracer cell; however, the actual outflow does not reach more than 500 m, which is shallower than the bottom (see Griffies et al., 2003). The alternative parameterization of Campin and Goosse (1999) does not fully address the question about entrainment and detrainment. Increasing the resolution locally (down to a scale of hundreds of meters) is also not feasible because it would preclude long-term simulations due to computational limitations (e.g., very small time steps).

The numerical simulations of Stanev et al. (1997) demonstrated that the Mediterranean Sea water is subject to vigorous mixing in the model and dilutes at a very small depth below the sill depth if no special treatment of the effluent is enabled. This is easy to be understand from the observations of Gregg and Özsoy (1999), who demonstrated that the major down-slope sinking might occur over distances smaller than our model grid interval. This means that convective adjustment, which is usually used when convection occurs over relatively large areas (more than one grid element), would be too strong for this tiny and localized outflow. Baines (2001) derives a non-dimensional buoyancy number \( B = QN^3/g f^2 \), where \( Q, N, \) and \( f \) are the two-dimensional volume flux, buoyancy frequency and reduced gravity, measuring the relative importance of forcing at the sill depth to stratification. For the conditions of the Bosporus Strait (transport \( \sim 10^8 \) m\(^3\) s\(^{-1}\), width of the strait \( \sim 1 \) km, \( \partial \rho / \partial z = 2 \times 10^{-2} \) kg m\(^{-1}\), \( \Delta \rho = 12.4 \) kg m\(^{-3}\) ) \( B \approx 2 \times 10^{-3} \), revealing the dominating role of stratification.

A basic question which we address in this paper is whether convection has to completely mix the water column (like in Eq. (3)). Suppose that the grid cell is much bigger than the convection area (or that we deal with 1D model where we have one water column with large area). Should then the convection homogenize the water column, or will the water column remain stable after the convection? If the latter is the case, then the mixing should be reduced (compared to the case of vertical homogenization), which will allow the deep penetration of the signal from the strait.

The above problems motivated Stanev et al. (1997) to formulate a simple parameterization, which is further developed and tested in the present study. We add a routine to the model code parameterizing the convection originating from the Bosporus Straits by controlling the exchange between Mediterranean and Black Sea waters. The concept describing the mixing between dense water parcels (DWP) and ambient water (see Appendix A) uses some ideas from the theory of
buoyant plumes (Batchelor, 1954; Morton et al., 1956, see also Turner, 1973, 1986). According to this theory the physics of vertical plumes are governed by three fundamental parameters which, in the case of conical plumes, are the vertical velocity \(w\), buoyancy \(g_0\) and radial length scale \(b\), with the corresponding equations:

\[
\frac{db^2 w}{dz} = 2\alpha bw, \quad \frac{db^2 w^2}{dz} = b^2 g', \quad \frac{db^2 wg'}{dz} = -b^2 wN^2, \tag{4a-c}
\]

where

\[
g' = g \frac{\rho^a - \rho^p}{\rho_0}, \quad N^2 = -\frac{g}{\rho_0} \frac{d\rho^a}{dz} \tag{5}
\]

and \(\rho^p\), \(\rho^a\), and \(\rho_0\), are the density of plume (index “p”), ambient water (index “a”), and standard density. The parameter \(\alpha\) controls the rate of entrainment of ambient fluid by the buoyant plume. If we take the simplest case of a uniform environment we obtain (from dimension analysis) \(w \sim z^{-1/3}\), \(g' \sim z^{-5/3}\) and \(b \sim z\). In this case, the mass flux \(wb^2\) increases with increasing depth while the buoyancy flux \(wb^2g'\) remains constant.

As atmospheric modeling (Gregory and Rowntree, 1990) and laboratory and theoretical studies (Baines, 2001, 2002) demonstrate, the two-way exchange (entrainment and detrainment) between DWP and ambient water better corresponds to the exchange in a stratified ocean than the entrainment-only case addressed in some classical theories. According to Gregory and Rowntree (1990) the entrainment is larger than the detrainment, thus the volume of the DWP increases with the depth (this situation is presented in Fig. 6a). This results in a decrease of the volume of the ambient water and an upward displacement of the water column. At the same time, the volume of DWP increases and its density decreases. The geometry of plume in Fig. 6a (the plume placed in the middle of grid element) better corresponds to the open ocean case (not limited by the presence of vertical boundaries). In the presence of a vertical boundary (Fig. 6b) we could consider only half of the model column of Fig. 6a, assuming thus that the model vertical wall passes through the middle of grid element.

We give in Fig. 6c an alternative (to the classical plume) representation of the sinking process.

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**Fig. 6.** The concept of the vertical plume. (a) Conical geometry, (b) the same as in (a) but in the presence of a vertical wall, (c) sinking prisms (analogues to thermals). Three grid elements are shown with a standard notation of grid intervals (\(\Delta x, \Delta z_i\)). The stratification is stable (the light-blue colors correspond to smaller density). The plume is always denser than the ambient fluid, however with increasing depth the difference between the densities of the plume and ambient water decreases. The red arrows represent the fluxes due to entrainment (inward with respect to the plume) and detrainment. The environment exhibits a compensating upward motion, and the thickness of the arrows indicates the change in vertical velocities with the changing depth. The concept presented in (c), although geometrically different from the one in (a), contains the same information: DWP sinks and its volume (thickness \(b^2\)) increases due to the imbalance between entrainment and detrainment. For simplicity, we present thermals at the moment when they have reached the bottom of the grid elements.
While in Fig. 6a the two fluids stay side by side in the grid box (as in the conventional plume theory), DWP-s in Fig. 6c are inside grid boxes (like thermals). In this representation the horizontal area surfaces of DWP-s are equal to the area of model grid interval and their changing thickness mimics the changing volume flux of buoyant plumes with increasing depth. Thus one of the most important characteristics of convection, which is the vertically variable volume flux, is accounted for. Of course, we could represent the same process using the geometry in Fig. 6a (vertical prisms with constant height, which is equal to the vertical grid interval \( \Delta z_k \)). In this case the radial length scale plays the role of the main variable, and it is related to the thickness of DWP \( h_k \) in Fig. 6c through \( b_k^2 = h_k \Delta x \Delta y / \Delta z_k \) (see Fig. 6 for the finite difference notation). This relationship reveals the identity of the two, seemingly different geometrical representations of the same process. It is also known from theory that the mathematical description of thermals and buoyant plumes is very similar (Turner, 1973). The advantage of working with \( h \) instead of \( b \) is the slightly simpler numerics. The other argument for our choice is that the parameterization of convection has already been briefly published using the concept in Fig. 6c (Stanev et al., 1997).

Some new experiments and theories (Baines, 2001) suggest that for slope currents spreading over slightly slopped topography the detrainment could be larger than the entrainment. Observations (e.g., Gregg and Özsoy, 1999) reveal that the Bosporus flow can be detected on the shelf, maintaining high-salinity values almost down to the continental slope. Although the funnelling effect might not correspond well to the two-dimensional analysis of Baines (2001), it seems possible that for this flow spreading over \textit{slightly slopped topography} (and even more when channeled by underwater canyons), the detrainment is dominating. We stress that the interest here is not on the shelf mixing (this model is too coarse to well resolve the shelf). Of major interest here is the penetration of surface signals into the deeper layers. This penetration occurs in the area of the abrupt continental slope, where the dilution of the effluent is quite strong. The large values of turbulent kinetic energy observed by Gregg and Özsoy (1999) might indicate that in this area the entrainment becomes dominant. The relevance of the two possible cases (entrainment or detrainment domination) to the processes addressed in this paper will be checked later. Here it is important to note that the developed parameterization is general enough and can be also applied when detrainment is dominant.

We can roughly understand the basic dynamics behind our convective parameterization by simplifying Eq. (4a–c). The equation which is inconsistent with the hydrostatic numerical model is the momentum equation in the vertical, thus we neglect it and suppose \( w = \text{const} \). Then, it follows from Eq. (4a) \( b = \alpha z \), which coincides with the solution in the non-stratified and non-hydrostatic case. From Eq. (4c) we obtain

\[
\frac{d \rho^p}{dz} = \frac{2}{\alpha z} (\rho^a - \rho^p),
\]

i.e., the density of DWP relaxes to the density of ambient water and the strength of this relaxation decreases with the depth. A similar result is obtained from Eq. (4) if we suppose that the radius of the DWP is constant. In this case

\[
w = w_0 \exp \left( \frac{2\alpha}{b} z \right)
\]

and

\[
\frac{d \rho^p}{dz} = \frac{2\alpha}{b} (\rho^a - \rho^p).
\]

Thus, the simplified Eqs. (6) and (8) demonstrate that in the stratified ocean, with increasing depth, the density of DWP tends to the density of ambient water (this is represented in Fig. 6 by the decreasing contrast in colors). The parameters controlling this trend (i.e., the depth of penetration of DWP) will be addressed in Appendix A.

### 3.4. Boundary conditions for tracers

Besides the main tracers, temperature and salinity, the model has two additional tracers, CFC-11 and CFC-12, which are integrated in the same way as the main tracers. As shown by Bullister and Lee (1995), CFC-11 is
non-conservative under anoxic conditions, thus we introduce a linear decay term in the right-hand side of this tracer equation with the rate constant taken as depth dependent from LBMS. The bottom is insulating for these two tracers.

The surface forcing for CFCs is formulated using Dirichlet-type boundary condition and the atmospheric CFC concentration history for the Black Sea region, derived from the Walker et al. (2000) atmospheric model (Fig. 3c) and normalized to the 1988 atmospheric CFC measurements. It is more difficult to formulate the boundary condition for CFCs in the Bosporus inflow entering the Black Sea. This forcing is crucially important for penetration of CFC into the deep layers; therefore, we carried out a number of sensitivity analyses, which led us to the following simple concept. As seen in Fig. 5b the vertical integral of the Black Sea curve in the deep layer is much smaller than the same integral in the Marmara Sea. The fact that the two curves intersect indicates that the surface signal has different rates of vertical penetration in the two basins (it does not easily penetrate deep in the Black Sea). The integrated amount of CFC in the Black Sea curve below the depth of intersection could be due: (1) the inflow from the Marmara Sea, (2) entrainment by this inflow of surface Black Sea water with high CFC content, and (3) vertical processes within the Black Sea. If (2) and (3) were negligible, the boundary condition could be formulated as follows:

\[
CFC_{\text{plume}}(t) = \frac{CFC_{\text{plume}}}{CFC_{\text{surf_forc}}} CFC_{\text{surf_forc}}(t),
\]

where

\[
CFC_{\text{plume}} = \frac{1}{H_c} \int_{\text{depth}} CFC_B \, dz,
\]

\[
CFC_{\text{surf_forc}} = \frac{1}{T} \int_{\text{time}} CFC_{\text{surf_forc}} \, dt.
\]

\[
H_c = \left( \frac{Q_2}{BS_{\text{surf}}} \right) T \text{ is the thickness of the water column corresponding to the total volume of Mediterranean Sea water transported into the Black Sea during the time of integration, } BS_{\text{surf}} = 45,000 \text{ km}^2 \text{ is the area of the Black Sea, } Q_2 \text{ is the mean transport of the undercurrent averaged for the whole period } T \text{ (if } Q_2 = 200 \text{ km}^3 \text{ yr}^{-1} \text{ and } T = 40 \text{ years, } H_c \sim 20 \text{ m}). \text{ The vertical integration in the above formulae starts from the depth where } CFC_{12B} = CFC_{12M}, \text{ (} \sim 80 \text{ m according to Fig. 5b). In the above equation, the indices } B \text{ and } M \text{ stand for Black and Marmara Sea. } CFC_B(z) \text{ and } CFC_{\text{surf_forc}}(t) \text{ are the CFC concentrations in the Black Sea water column (e. g. Fig. 5b) and for Black Sea surface water in equilibrium with the atmosphere (Fig. 3c). Obviously, Eq. (9) normalizes the surface concentration by the mean concentration of the tracer penetrated in the Black Sea during the period of model integration, and the time modulation is based on the time history of the forcing.}

We will show later that any reduction of CFC_{\text{plume}} that would be necessary in order to calibrate the model so that the simulations match the observations would reveal the relative contribution of entrainment versus strait flux (we remember that the above considerations assume the CFCs entering the Black Sea are carried by the DWPs and the entrainment of surface water is negligible). In the CFC-11 calculations we use the same value as for CFC-12, but normalized by the ratio of the concentrations of the two tracers at the sea surface.

4. 1D simulations

4.1. Description of the 1D model

The performance of parameterizations associated with unresolved vertical processes is illustrated below using a simple 1D model. The model consists of two parts: the open ocean column and the DWP, the latter carrying dense Mediterranean water into the deep layer. The physics of the ocean column includes parameterization of the surface mixed layer and vertical mixing in the deeper layers, the diffusion coefficient is depth dependent and is calculated in the same way as in the 3D model (Eq. (2)). The same vertical discretization and numerical schemes of the MOM are used when solving the finite difference equations in the 1D model. The forcing is based on the data described in Section 2.1.
The flux of much saltier Mediterranean Sea water at the sill depth justifies the assumption that each grid element is composed of two water bodies: ambient water and the DWP. In the absence of the negative buoyancy flux from the strait (e.g., closing the strait) the entire water column is identical with the ambient water. The parameterization of the outflow assumes that mixing between the two water types is not locally instantaneous (DWP changes its properties but still conserves its identity, which is manifested by its higher salinity/density compared to that of the ambient water). At any depth the model column gains (through detrainment) from the DWP some amount of Mediterranean characteristics (salt, heat, CFC) and exports (through entrainment) Black Sea water into the DWP. In the case shown in Fig. 17 the volume of the detrained water ($\sim \delta \Delta h_1$) is smaller than the volume of the entrained water ($\sim \epsilon \Delta h_1$), which leads to an increase of the volume of the DWP during each model time step. Our sensitivity experiments demonstrate that the specific choice of $\epsilon$ and $\delta$ controls the depth of convection. The response displayed in Fig. 7 (when large entrainment rates result in larger mixing and shallower convection) is known from atmospheric cloud modeling studies (Gregory and Rowntree, 1990). The ratio between the coefficients of entrainment ($\epsilon$) and detrainment ($\delta$) of 3 is consistent with the values used by these
authors and ensures best agreement with the observed salinity stratification. The evidence that this extremely dense inflow does not reach deep and bottom layers indicates large mixing (shallow convection) and support the choice \( \varepsilon > \delta \). As seen from the sensitivity experiments (Fig. 7) and also from the study of Gregory and Rowntree (1990), \( \varepsilon < \delta \) would result in much deeper penetration of the DWP. However, the exponential decrease of the volume flux of DWP (see Appendix A) would make it in realistic conditions subject to mixing with the dynamic environment and it would lose its identity, as this is clearly seen from the observations of Gregg and Özsoy (1999) in the area of the continental slope. Even more important is that for \( \varepsilon > \delta \), the simulated CFC signals do not reach the deep levels (see Fig. 5b). The latter presents an important constraint when specifying the model parameters (see Appendix A).

The excursion of the convective element from the sill depth down to its neutral density level (at this level the DWP is completely diluted and loses its identity) results in changing properties of the water column. These changes are treated formally in a similar way as those parameterized in MOM by the procedure of convective adjustment (Pacanowski et al., 1991). Thus the virtual convective element does not directly affect the momentum equations.

In the present study, we assume that the intensity of entrainment/detrainment is proportional to the difference between Black Sea and Marmara Sea salinity (\( S_B \) and \( S_M \))

\[
\varepsilon \sim m(S_M - S_B),
\]

where \( m \) is a constant taken as the inverse difference between present-day bottom (\( S_{Bot} \)) and surface (\( S_S \)) salinity. This parameterization is similar to that used to model entrainment–detrainment parameterization of air streams flowing through open doors/windows (Linden, 1999) and is consistent with the evidence that the present-day convection is very shallow (large entrainment), i.e., \( S_M - S_B \) is much larger than \( S_{Bot} - S_S \). This parameterization ensures more accurate specification of the exchange with the ambient water, thus the termination depth is free to adjust to the forcing parameters and model physics.

4.2. Sensitivity of CFC penetration on the external forcing

We next discuss the results of several sensitivity experiments. Our first task was to find the ratio between waters penetrating into the deep levels originating at the Black Sea surface, versus those from the Marmara Sea, as well as the most appropriate boundary conditions for the tracer equations. We carried out two extreme experiments. In the first the CFCs entered the Black Sea from the sea surface only, and the flux through the Strait of Bosporus was set to zero (dashed lines in Fig. 8). In the second, the flux of CFCs at the sea surface was set to zero, and the entire tracer flux in the model originated through the Bosporus (short dashed line in Fig. 8). In the latter experiment the boundary condition expressed as Eq. (9) was used. In each of these experiments the physical model was the same, and entrainment of Black Sea water by the DWP originating from Bosporus Strait provided the main mechanism of penetration of surface and intermediate water into the pycnocline. Triangles show a representative CFC profile from the Knorr station 14. The figures indicate that the integrals of the dashed and short-dashed curves are almost identical, which shows that both of these extreme boundary conditions provide the same total amount of tracer into the ocean. However, the shape of the curves demonstrates that in the “real” sea most of the CFC penetrating the deep layers has its source at the sea surface within the Black Sea rather than in the Marmara Sea! The slight underestimated of CFC-12 deep concentration in the “flux from the atmosphere only” model indicates that a small flux of this tracer may be due to inflowing Marmara Sea water. However, this flux is much smaller than that from sea surface. (—) in Fig. 8 (Control Run, CR) corresponds to the case where 17% of the flux given by the boundary condition of Eq. (9) is prescribed in the strait. The agreement of this simulation with observations (e.g., Murray et al., 1991) demonstrates that the inflow of CFC from the Marmara Sea is a small, but not totally negligible source for the Black Sea CFC concentrations.
Despite some small differences between the results presented here and those of LBMS, which are mainly due to the non-conservative nature of CFC-11, the behavior of the two models is quite similar. The low deep-water values demonstrate how isolated the deep water is from the surface signals, and that under the average present conditions the plume does not ventilate depths below 600 m in amounts that can be detected using CFC tracers. Rare ventilation events that penetrate deeper are not excluded, but cannot be resolved with the present model. The good agreement between the results of this direct 1D model and the inverse type of modeling of LBMS is quite encouraging. This demonstrates that the vertical parameterizations are well calibrated to the Black Sea conditions.

5. 3D modeling: analysis of the simulated circulation

The combination of the major forcing mechanisms (negative buoyancy forcing along the coasts caused by fresh water from rivers, a compensating positive buoyancy flux through the Bosporus strait and the cyclonic wind stress field) tends to establish a cyclonic one-gyre circulation system, which is shaped by the simple coastline and topography. The extremely strong stratification and steep bottom slope tend to keep the mean position of the gyre attached to the continental slope. However, the baroclinic instability of the frontal interface creates mesoscale variability, which is characterized by meanders with large amplitudes (e.g., south of the Crimea peninsula) that intrude into the open sea and result in large variance in the magnitude of the velocity (Fig. 9). These eddies are partially responsible for mixing between coastal and open-sea waters.

The maintenance of vertical stratification in the Black Sea is not only a result of the balance between different buoyancy forces (thermal and haline), but is also associated with vertical circulation established as a result of Ekman drift of surface waters toward the coasts. This Ekman transport is compensated by inward motions in the deep layers. The cell is closed in the vertical by upward motion in the basin interior and descending motion in the coastal regions. This vertical circulation results in the formation of domed-shape isohaline surfaces (Fig. 10a). In the basin interior the pycnocline is very thin. The isopycnal
surface of $\sigma_0 = 14.5$ which marks the upper boundary of the CIL is at 40 m; however, in the coastal anticyclonic area the same surface is usually observed at 70–80 m. Thus, the coastal waters tend to occupy thicker layers, while the water masses in the interior are rather thin. This has a pronounced impact on the ventilation regimes, as shown in both data analysis (Ovchinnikov and Popov, 1987).
and numerical simulations (Stanev and Staneva, 2001).

The pattern of vertical circulation and the small vertical mixing under strongly stratified conditions control the formation of cold intermediate water. This water can be found between depths of 40–120 m (Fig. 10b). As can be seen from the vertical cross-section of temperature in isopycnal coordinates (Fig. 10c) it lies between the isopycnal surfaces of $\sigma_0 \sim 14.5$ (in coastal regions) and $\sigma_0 \sim 15.8$ (in the basin interior). The predicted differences in the depth of the core of the CIL indicate that pronounced diapycnal mixing dominates the Black Sea thermodynamics at the depth of the CIL. This is associated with upwelling in the basin interior, which brings saline deep waters closer to the sea surface. It is noteworthy that most waters penetrating the CIL have low density (salinity) and originate from the coastal regions rather than the central gyres (Fig. 10c). In summer, the CIL is overlain by a stable thermally stratified surface water mass, which is identified by the isopycnically thick layer of warm surface water in Fig. 10d.

The downward propagation of the surface signal with the Bosporus plume presents only one (though a very important) element of the vertical circulation and cannot be separated from the rest of vertical motions, e.g., the downwelling in the
region of coastal anticyclones and the interior basin upwelling. It will be shown in Appendix A that in the case of a very simple physics dominating the processes in the ambient (balance between vertical diffusion and convection of DWP), the stationary state is described by the following equation:

\[ K_p \frac{\partial S^p}{\partial z} = -K_v \frac{\partial S^a}{\partial z}, \]

where \( K_p = wH(\epsilon - \delta/\epsilon) \), \( w \) and \( H \) are vertical velocity of the DWP and the depth scale. Eq. (11) gives a simple example of coupling between physics governing DWP and physics governing macro-scale processes (e.g. the formation of the halocline). This is a two-way coupling. The DWP does not simply propagate into a fixed ambient water column; the water column in turn adjusts to the processes governing the propagation of the DWP. Adding wind-induced upwelling (i.e., extending the simple theory for the main oceanic thermocline to the Black Sea, Stanev, 1989) is just a technical problem. Thus by unifying downward motions in the coastal zone (caused by convective mixing and coastal anticyclones) and the upward motion in the basin interior (associated with the cyclonic wind forcing and resulting Ekman drift) the vertical circulation regulates the different dynamic controls. This leads us to a major hypothesis of this paper, which is that there can be different and complementary ways to estimate the vertical circulation: (1) estimates of Ekman pumping from wind data, (2) computing the intensity of vertical circulation cell (Stanev, 1990; Bulgakov et al., 1996), (3) end-member mixing analysis (Murray et al., 1991; Bueseler et al., 1991). Some of the approaches can be biased by the inherently large errors in the data. Others can be biased by the limitations of the assumptions. However, because the processes associated with ventilation of the pycnocline are hydrodynamically controlled, a combination of modeling and data should be able to provide more reliable estimates. This is justified by the results of the simple model developed by LBMS, which is a kind of inverse model. In our study we go one step further. Instead of minimizing the errors between observations and simple model results (as is done in inverse models), we develop a hydrodynamically consistent three-dimensional model capable of correctly replicating the observations. This serves as a reliable tool to make further analyses of the synthetic data set produced here much more complete.

6. CFC mixing patterns and estimation of Black Sea mixing mechanisms

6.1. Horizontal patterns of CFC

CFC-11 and CFC-12 have similar atmospheric trends and similar surface forcing trends in the model. This explains the similarity in the spatial patterns at different depths, thus here we will only analyze the simulated spatial patterns of CFC-12. It is clearly seen in Fig. 11 that the model simulated surface patterns (depth = 17.5 m) are dominated by the higher solubility (Warner and Weiss, 1985) in the cold coastal waters and by the dynamics along the southern and eastern coast, which result in tongue-like patterns. The horizontal gradients are about two times smaller in summer (Fig. 11b) than in winter (Fig. 11a), which agrees with the climatology of sea surface temperature (SST) in the Black Sea (Staneva and Stanev, 1998). In the deeper layers (\( z = 185 \) m) the distributions of CFC-12 are consistent with the general circulation maintaining upward motions (of low CFC-12 deep water) in the basin interior and downward motion (of CFC-12 rich coastal water) along the basin periphery. Unlike the surface patterns, which are strongly dominated by temperature (note the high concentrations in August in the region of cold-water formation west of the Crimea Peninsula in Fig. 11b) the deep patterns are shaped primarily by the vertical circulation (Fig. 11c,d).

A fundamental characteristic of the tracer fields in the Black Sea is their tendency to align along isopycnal surfaces, as shown by the horizontal patterns of CFC-12 in both winter and summer in the core of the CIL (\( \sigma_y = 15.0 \)) and in the upper halocline (\( \sigma_y = 16.0 \)) just above the depth of H2S onset (Fig. 12). The spatial inhomogeneity of CFC-12 (measured by the rms on \( \sigma_y = 15.0 \)) is
This is 5 times smaller than the corresponding inhomogeneity at 65 m (rms = 0.39), which is approximately the mean depth of $\sigma_0 = 15.0$. The corresponding numbers for the deeper level ($\sigma_0 = 16.0$, the mean depth of this surface is 145 m) are 0.0013 and 0.08. Note that the contrasts as seen on density surfaces are ~70 times smaller than the contrasts which one finds on z-surfaces. These estimates support the finding of Spencer and Brewer (1971), Shaffer (1986) and others that deep-water properties in the Black Sea are relatively constant with respect to the density (salinity) field. We also point out that Fig. 12 gives an indication of the mixing paths: the gradients are in the area of the main gyre where CFC-12 rich water intrudes from the margin into the open sea. Thus the cross gyre transport tends to increase the open sea concentration of CFC in deeper levels, where the processes tend to evolve along the isopycnals. This is just the opposite to what is observed for the salinity field, where the cross gyre transport tends to decrease the salinity in the open sea. The CFC-12 is more uniform in the basin interior than in the coastal area, which indicates that the Black Sea circulation supports the theory of homogenization of potential vorticity. Another region of pronounced mixing occurs near the Bosporus Straits where the DWPs entrain CFC-12 rich water from the CIL. The main gyre carries this signal along the coast.

Even though small errors are possible in these estimates, they are quite indicative of the precision of measurements needed to trace the mixing paths of CFC in the halocline. The maximum contrast at $\sigma_0 = 16.0$ is ~0.04 pmol kg$^{-1}$ (Fig. 12c, d). The estimated precision for CFC measurements on the
1988 Black Sea cruise was ~0.01 pmol kg\(^{-1}\) (Bullister and Lee, 1995) which restricts the ability to discriminate fine-scale features on these deeper density levels. On the other hand, the alignment of tracers along isopycnals tends to reduce the amount of information needed to specify the characteristics of vertical propagation of surface signals across the pycnocline.

The dependence of CFC-12 on the circulation is clearly illustrated in Fig. 13 where we show a model prediction of the meridional cross-section along 31.62°E in winter and summer plotted against depth (Fig. 13a, b) and density (Fig. 13c, d). The meridional cross sections of CFC-12 agree with the major characteristics of the vertical circulation (the model simulates upwelling in the basin interior which shapes the pycnocline, Fig. 10a). The low CFC-12 values simulated in the upwelling zone are due to the supply of low CFC-12 deep waters. The domed shape of the CFC-12 concentration patterns versus depth correlates well with the characteristics of the Black Sea isopycnal surfaces. The isolines of CFC-12 concentration deepen in the coastal areas (more pronounced when plotted against density). Their slope increases over the continental slope, which would enhance cross gyre exchange in the case of baroclinic instability. Thus, diapycnal mixing in the region of the main gyre provides an important mechanism for penetration of CFC-12 into the pycnocline and deeper layers.

Two spatial characteristics also deserve attention, namely:

1. The source waters with high CFC-12 concentration are well pronounced in the northernmost regions of the cross section, correlating with the temperature and salinity at the sea surface.
2. Along the coast in August there is a local subsurface concentration maximum at 40–50 m (Fig. 13b).
The areas of penetration of CFC from the sea surface to deeper levels are best seen when concentrations are plotted versus density levels (Fig. 13c,d). The maximum concentrations of CFC-12 are observed along the coasts. These are also the areas where the slope of isohalines also contributes to deeper penetration of CFC-12. Along the southern coast (in the Bosporus area) the higher concentrations reveal the intrusions of Mediterranean Sea water. The comparison of the meridional sections of CFC (Fig. 13) with those of temperature and salinity (Fig. 10) demonstrates that CFC correlate with salinity rather than temperature (there is no intermediate layer like CIL seen in the CFC cross-sections plotted in density coordinates!).

6.2. Temporal variability

In order to illustrate the control of basin dynamics on the seasonal variability in the intermediate layers we analyze simulated CFC-12 and temperature data series (from January, 1985 to December, 1998) at two locations with coordinates 30.3°E, 42.5°N (Fig. 14a, c) and 33.0°E, 44.5°N (Fig. 14b, d). The first location is in the interior cyclonic part of the western sub-basin and the second one is in the coastal region west of the Crimea Peninsula. The main difference between the two locations is the deeper ventilation of intermediate layer water in the coastal location and much shallower penetration of surface signals in the open ocean location. This difference is

Fig. 13. Meridional cross-section of CFC-12 along 31.62°E plotted against depth (a,b) and density (c,d) in February and August 1988.
controlled by the depth of the pycnocline (the smaller depth of ventilation of the open sea is due to the upwelling there). In both locations the temporal evolution of temperature (Fig. 14c, d) is typical for the seasonal thermocline.

The interannual signal is also quite pronounced, and can be detected in the temperature field down to 100 m (see the large production of CIW in the cold 1986, 1987 and relatively small production in the warm 1985). The temporal evolution of the CFC-12 fields (Fig. 14a, b) is smoother and more regular compared to temperature. The deep layers show very weak interannual variations, in contrast to the Mediterranean Sea (Roether et al., 1996), because the stable stratification creates a two-layer vertical structure decoupling surface and deep layers. Theoretically, because of the high precision of computed values, the simulations of CFCs could serve to trace weak changes in the Black Sea circulation over the past 50 years triggered by atmospheric interannual variability. However, this period is too short, and the values of CFCs are too small, which makes it difficult to find deep penetrating signals in the observed data, and to carry out intercomparisons between observations and simulations for the deep layers.

The slope of the CFC contours in the warm part of the year associated with the decreasing surface values (this process starts in May, Fig. 14), and in particular the well pronounced negative anomalies in the warmest part of year (well seen during 1987 in the open sea station) support the subsurface

Fig. 14. Temporal evolution of vertical profiles of CFC-12 from January 1985 to December 1988 (a,b) and temperature (c,d). The coordinates of the locations are given in the figures.
maximum of CFC measured during the Knorr-1988 cruise (LBMS, see also Fig. 8). Obviously, this feature is due to the seasonal evolution of the upper ocean and is associated with the large SSTs in summer reducing the solubility of CFC (see also Haine and Richards, 1995).

The ventilation of the Black Sea pycnocline on density coordinates (Fig. 15) seems quite different when analyzed using the penetration of different surface signals (e.g. temperature, salinity, CFCs). The temperature plot (Fig. 15b) shows that during the time of maximum outcropping, the low-temperature water penetrates down to $\sigma_\theta = 14.7$ (in open sea locations this signal reaches higher-density surfaces) and persists throughout the year, deepening slightly with increasing time. At the time of outcropping, a veering of CFC-12 is observed with slight deepening of the isolines in the CIL, which correlates with deepening of the core of CIL. However, no traces of an intermediate minimum (or maximum) are observed in the core of CIL. Structures similar to those in temperature (Fig. 15b) are seen when plotting the difference between local values and the horizontally averaged CFC value. These CFC-12 anomalies originate from the outcropping area (Fig. 15c), demonstrating that the location chosen acts as a source of positive CFC anomaly (see also the horizontal patterns in Fig. 11). However, comparison between Fig. 15c and d indicates that the CFC anomaly patterns reveal characteristics of the salinity anomaly field. While the cold temperature
anomalies are strongly associated with the areas of continental slope in the northwestern Black Sea. The anomalies of CFC take the pathways of salinity.

The above results indicate that the difference between penetration patterns of passive tracers and temperature could also be due to the difference in the surface fluxes (in the case of temperature the fluxes are controlled by extremely strong seasonal variability). For example, the ratio between the amplitude of seasonal variations of SST and its annual mean value is 1.4. The same ratio for the surface concentrations of CFC is 0.6 (Fig. 14). This demonstrates that the seasonal variability, which results in a periodic replenishment of CIL with cold water, is not as effective mechanism for creation of intermediate layer with maximum CFC concentrations (remember that the solubility of CFC is a function of temperature and low temperatures correspond to high values of CFC). Nevertheless, layers with intermediate maximum of CFC are observed and simulated at 20–30 m. The above results are indicative of the possible evolution of the CIL if winters are warmer (weak forcing). The observations during extremely warm 2001–2002 do give support to above results demonstrating that the CIL in this period was much less pronounced.

The accuracy of the results of the 3D tracer model for studying the dynamics and mixing of Black Sea waters is supported by comparisons between field observations and model simulations (Fig. 16). The shaded area indicates the simulated 1σ deviation from the basin mean. Almost all data from observations (symbols) fall within these limits, except some of the deep values from station 3, which is close to the Bosporus Strait, and which exhibit a subsurface maximum of CFC-12 (see also LBMS). The absence of any other available CFC data makes the extension of intercomparison between simulated and observed data impossible for other periods and motivates further observational activities. However, even for just one
snapshot in time, the agreement between field and model data is very promising and suggests that with this approach it should be possible to derive useful information about the impact of circulation and mixing on intermediate and deep-water properties.

7. Conclusions

We present in this paper results of numerical simulations, which replicate recent CFC measurements in the Black Sea and quantify fundamental physical characteristics associated with mixing pathways and ventilation of intermediate and deep water. In order to realistically simulate the convection in the Bosporus Straits area, a specific parameterization is developed accounting for mixing between Mediterranean and Black Sea water. In this way, it is possible to include in the model all important branches of the vertical circulation: the downward motions in the coastal zone (caused by the sinking plume and anticyclonic eddies), the upward motion in the basin interior (associated with the cyclonic wind forcing and the resulting Ekman drift). As it was shown in the simplified theoretical example (Eq. (11)), the dominating dynamical controls are mutually adjusted. By realistically representing the vertical circulation and mixing it became possible to obtain consistent simulations of water mass formation (including pathways of different tracers).

Some of the more important estimates are the ultimate penetration depth of Bosporus intrusions (not deeper than 600 m), for which there was no clear consensus in previous studies, and the correlation between CFC data and CIL/halocline characteristics. The major role of the Black Sea pycnocline is that it blocks vertical exchange, thus the CFC-12 concentration decreases consistently with increasing density, reflecting the poor ventilation conditions. Low deep-water concentrations require careful measurements and parameterizations in numerical models, but also give very clear means of precisely detecting the penetration depth and the patterns associated with different dynamic controls.

The model simulated data give a clear indication that most of the CFC penetrating into the deep layers has its source at the sea surface rather than the Bosporus Straits. However, inflow of CFC from the Marmara Sea is not negligible. The results of the 3D tracer model support the observations and present a powerful tool for detecting the dominating mixing pathways. One major pathway is the cross gyre transport, which tends to increase the concentration of CFC in deeper levels, where the tracer fields align with the isopycnals. This is just the opposite of what is observed for salinity, where the cross gyre transport tends to decrease the salinity in the open sea. The homogeneous CFC-12 in the basin interior is consistent with the theory of homogenization of potential vorticity.

The second important mixing pathway is associated with the convection sites, which are controlled by upwelling in the basin interior and by the downward motion in coastal areas. The source waters with high CFC-12 concentration are well pronounced along the shelf edge in the northern region. The pycnocline acts as the major factor controlling the depth of winter cooling. The low CFC-12 values simulated in the upwelling zone are due to the supply of upper layers with low CFC-12 deep waters. During the time of maximum outcropping, the signals from the sea surface penetrate to shallower depths in the open sea locations than in the coastal areas. The CFC-12 anomalies originating from the outcropping area trace the source of surface signal and their time-depth patterns (Fig. 15) correlate with those of the salinity anomaly.

One important result from the simulations is that the deep layers of the Black Sea do not show pronounced sensitivity to interannual variations in forcing. This is due to the decoupling of the surface and deep layers because of density stratification and to the very low levels of CFC in the pycnocline (unlike the active response of the neighboring well ventilated Mediterranean Sea to surface forcing). We demonstrated that the simulations could provide estimates of the precision of measurements required to trace variability and mixing paths of CFC in the halocline. The maximum contrast at $\sigma_B = 16$ (the depth of
H$_2$S onset) of ~0.03 pmol kg$^{-1}$ is difficult to resolve with the precision of measurements (~0.01 pmol kg$^{-1}$), which restricts the ability to discriminate fine-scale features on these deeper density levels. On the other hand, the alignment of tracers along isopycnals tends to reduce the amount of information needed to specify the characteristics of vertical propagation of surface signals across the pycnocline. Thus simulations could provide first-order criteria for deciding what is the reasonable amount of data required from different depth (density) levels to efficiently develop more informative databases. The absence of additional CFC data makes the extension of intercomparisons between simulated and observed data impossible for other periods and motivates further observational activities.

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Appendix A. The parameterization of convective mixing in the inflow area

The convection process can be described both in terms of changing volumes of DWPs and by changing vertical velocity (or more generally volume flux). The former concept has several advantages: (1) it can be easily accommodated within the grid concept of numerical models, and (2) it avoids introducing new velocities, thus it does not directly affect the momentum equations which could create difficulties for the models, (3) it allows DWPs to be considered as virtual elements (Figs. 6c and 17), which only redistribute buoyancy in the vertical, (4) the role of DWPs are to some extent analogous to bottom slabs in other parameterizations in that the physics is extremely simple. However, our DWPs (Fig. 17) do not need to be attached to the bottom, which makes them also a useful concept for parameterizing convection originating from the sea surface, in particular under condition of small mixing between ambient

![Fig. 17. The concept of the entrainment of ambient water by DWPs. The initial volume $V_0 = b^2 h_0$ (where $b^2$ is the horizontal area and $h_0$ is the height), which equals the inflow over one time step, exchanges water with the ambient fluid. Every grid column (with thickness $\Delta z_k$, where $k$ is a vertical grid level in the model) can be represented as being composed of ambient water (thickness $h^a_k$) and DWP (thickness $h^p_k$). The entrained ambient water ($\delta \Delta h_1$) exceeds the detrained into the ambient water ($\Delta h_1$), thus the thickness of the convective element increases and the thickness of the ambient column decreases (b). The increase of the thickness of the convective element and the decrease of the thickness of the water column induces a displacement of the water column from below. The stratification in the water column changes as a result of vertical displacement and horizontal mixing. The final state after the mixing between two neighbor layers and the DWP has been completed is given in (c).](image-url)
water and DWPs. With respect to arguments (3) and (4) above, it should be noted that we do not explicitly account for convective dynamics, but only redistribute buoyancy and mass in the vertical using simple geometrical and mixing considerations. Thus the task in this Appendix is to demonstrate that the considerations behind the numerical algorithm based on the concept of mixing controlled by entrainment and detrainment (Fig. 17) are physically consistent.

When analyzing the dynamical consistency of the proposed parameterization, two approaches are possible: (1) formulate the basic laws and derive from them a numerical analogue, (2) present the concept and the corresponding numerical algorithm and then study their consistency. Although the first alternative is more straightforward, we decided to start with the second one because the theory of buoyant plumes could lead to a number of simplified formulations in finite differences, and our concept (Figs. 6c and 17) is only one of these possibilities. Again, in choosing finite volumes as the basic elements of our parameterization, we chose here to start with the numerics because the concept of water types and the mixing between different waters, which can easily be described in terms of finite volumes and their properties, can be better revealed using finite differences. Finally, from the analysis below, it is simple to derive the finite difference equations from the differential ones.

We assume that the model grid elements are filled by two water types: ambient water and DWPs. At the level of the inflow from the strait the water flux per model time step is $V_0$. Knowing that the horizontal area of the model grid element is $\Delta x \Delta y$, where $\Delta x$ and $\Delta y$ are the horizontal grid intervals (in a one-dimensional model $\Delta x \Delta y$ is the basin area), we obtain for the initial thickness of the convective element entering the model grid element $h_0 = V_0/\Delta x \Delta y$. In the following discussion we derive the equations for each of the water types and test their consistency.

The first task is to decide upon the geometry of the convective element. Ideas about this can be borrowed from theory, experience in atmospheric and ocean modeling, and laboratory experiments. The latter are particularly useful and we refer first to the recent work of Baines (2001) who extended the classical (non-stratified) model of Ellison and Turner (1959) to stratified fluids. The observations of Baines (2001) showed that the turbulent exchange across the upper boundary of down-slope gravity currents results in continuous loss of dense fluid. If the slopes are gentle, detrainment tends to dominate. In a later study, Baines (2002) showed that there is a fundamental difference between down-slope currents and two-dimensional plumes in stratified environments (in the latter case, entrainment seems to dominate). Unfortunately, we cannot directly transfer results from laboratory experiments to the Black Sea, and Table 1 lists significant differences between laboratory experiments and actual conditions in the Black Sea. One of the major problems is formulated in the last entry of Table 1 illustrating the difference between the approach here and in laboratory experiments: in our approach we address the adjustment of effluent to the environmental conditions, along with the corresponding feedbacks. Due to these differences, as well as to the problems with observation and simulation of the Mediterranean water inflow (see Section 3.3.2) it seems premature here to develop sophisticated parameterizations accounting for a large number of subgrid controls. Rather, we develop a simple parameterization (but superior to the convective adjustment for the case which we study), and by carrying out sensitivity experiments and comparing results with observations we find the optimal values of parameters which replicate the observed profiles. We reiterate that with the insufficient resolution of the model, the Mediterranean water “sees” the continental slope as a vertical wall and therefore we will use for the DWP an analogy with thermals.

The change of the volume of DWP with depth reflects the basic physics governing the convection in the model column. From Eq. (4), we find that the change of the normalized volume flux $Q = b^2 w$ obeys the equation:

$$\frac{1}{Q} \frac{dQ}{dz} = 2 \frac{a}{b},$$

which is valid irrespective of the form of stratification (Turner, 1973). This is a fundamental
difference from the case of simple convective adjustment used in MOM. The end results of the sinking of DWP are: (1) mixing of the two water types, and (2) vertical displacement of the ambient water (see Fig. 6). In the following we refer to the water column, or model column, assuming that this is the ambient fluid through which the DWP sinks.

We represent the DWPs as prisms with horizontal area $b^2 = \Delta x \Delta y$. Their height $h$ increases with increasing depth, which is caused by the difference between entrainment and detrainment. The resulting increase of the volume is $dV = b^2 dh$, where $dh$ is the increase in thickness. Because $b = \text{const}$ and $Q = b^2 w = b^2 h/\Delta t$ the analogue of Eq. (A.1) is

$$dh = \frac{h}{H} \, dz,$$

(A.1')

where $H$ is the penetration depth, at which the initial volume $V_0$ changes $e$-fold. As will be shown, the exponential change of $h$ with depth allows simple solutions to be obtained and analyzed.

Because $w = h/\Delta t$

$$dw = \frac{w}{H} \, dz,$$

(A.1'')

which says that the vertical velocity obeys the same law as the thickness of the DWP. This exponential dependence corresponds to the geometry of the model ($b = \text{const}$), leading to exponential change of the volume flux $Q = b^2 w$ with the depth (in neutrally stratified fluids $Q \sim z^{5/3}$, see Section 3.4). Obviously, the above two relationships allow analysis in terms of thickness rather than velocity, which makes

<table>
<thead>
<tr>
<th>Laboratory experiment</th>
<th>Black Sea case</th>
</tr>
</thead>
<tbody>
<tr>
<td>Simple topography with constant slope</td>
<td>Variable slope, underwater canyons, hills and other topography roughness</td>
</tr>
<tr>
<td>Uniform (linear) stratification in the initial state. In the final state the density gradients in upper layers are smaller than in deep layers</td>
<td>Non-uniform stratification (Fig. 5). The density gradients are large in upper layer and smaller in deeper layers</td>
</tr>
<tr>
<td>The density of inflowing fluid is lower than the density in deepest most levels</td>
<td>The density of inflowing fluid is much larger than the density of deep levels (see Fig. 5c). Nevertheless, the convection does not reach deepest most levels indicating that mixing (due to wind or breaking waves) in surface layers is much larger than in laboratory experiments</td>
</tr>
<tr>
<td>Two-dimensional downslope currents and buoyant plumes are studied. These currents originate from a line source</td>
<td>The strait is very narrow resembling a point source. The observations on the shelf (salinity, thickness of the inflow, pathways) reveal a strong veering indicating that the real case is far from two-dimensional</td>
</tr>
<tr>
<td>The outflow is constant with time</td>
<td>There is a clear evidence of pronounced variability of the transports in the strait causing variability in gravity currents</td>
</tr>
<tr>
<td>Almost motionless ambient fluid</td>
<td>The real ambient water is dominated by strong time-varying currents and eddies</td>
</tr>
<tr>
<td>There is no upper layer mixing caused by dynamics (wind and baroclinic instabilities)</td>
<td>The wind mixing and diapycnal exchange between coastal and open sea waters control the vertical stratification</td>
</tr>
<tr>
<td>Non-rotating fluid</td>
<td>Rotating fluid</td>
</tr>
<tr>
<td>The time needed to fill the laboratory tank with dense water is relatively short</td>
<td>The replenishment time scale is on the order of thousands of years</td>
</tr>
<tr>
<td>Relatively short experiments are carried out in order not to allow the initial isopycnals to change much</td>
<td>The density is free to adjust to the gravity currents. This adjustment is the key issue in this paper and the sensitivity experiments have been carried out for thousands of years</td>
</tr>
</tbody>
</table>
numerical formulations easier. For understanding the results in terms of transports, we just have to keep in mind the proportionality between velocity and thickness of DWPs.

We postulate that heat, salt, and mass are conservative in the DWPs and ambient water. Suppose that the grid element at level “k”, which has a thickness $\Delta z_k$, can be considered as filled with ambient and dense water, and their volumes are $b^2 h^p_k$ and $b^2 h^a_k$ correspondingly, where $h^p_k + h^a_k = \Delta z_k$ (see Fig. 17). For simplicity in the following presentation, the properties of the DWPs and ambient water are described using salinity as the only thermodynamic variable. However, in the model, the calculations for temperature are carried out in the same way as for salinity.

The thickness of the convective element $h^p$, its salinity $S^p$ and temperature $T^p$ are initialized at the depth where the current exits the strait by $h_0$, $S_0$ and $T_0$, where $S_0$ and $T_0$ are salinity and temperature of the Mediterranean water. The DWP gains water from the water column by entrainment at a rate, which is proportional to $\epsilon \Delta h_1$ and exports water into the water column by detrainment at a rate which is proportional to $\delta \Delta h_1$ (see Fig. 17), where $\Delta h_1$ is a thickness scale, which we will determine below. The change of volume of the DWP is proportional to $\epsilon \Delta h_1 - \delta \Delta h_1 = (\epsilon - \delta) \Delta h_1 = \Delta h^p_1$, thus $\Delta h_1 = \Delta h^p_1 / (\epsilon - \delta)$, where $\Delta h^p_1$ is the increase of the thickness of convective element after passing from level “k” into the level “k + 1” (see Eq. (A.1’)). If $\epsilon \approx \delta$ the thickness of DWP remains constant ($\Delta h^p_1 = 0$), and the parameterization tends to the standard convective adjustment.

The increase of the volume of the DWP is accompanied by a decrease of its density, but the latter can still remain larger than the density of the grid element at level “k + 1”. If this is the case, the DWP sinks down to the bottom of the next grid element “k + 1”. An equal volume of water from level “k + 1” is displaced upwards into the level “k”. Only after reaching its neutral density level does the DWP lose its identity and the convection terminates.

Assuming conservation of salinity, we can quantify the above description as
\[
S^p_k (h^p_k + \Delta h^p_k) = h^p_k S^a_k + \frac{\epsilon}{\epsilon - \delta} \Delta h^p_k S^a_k - \frac{\delta}{\epsilon - \delta} \Delta h^p_k S^p_k,
\] (A.2)
where $S^p_k$ is the new salinity of the DWP at level “k”. Recalling that with this salinity the DWP enters the underlying grid element ($S^a_k = S^p_{k+1}$) we can write from Eq. (A.2) the following equation:
\[
S^p_{k+1} h^p_{k+1} = S^p_k h^p_k + \frac{1}{\epsilon - \delta} \Delta h^p_k (\epsilon S^a_k - \delta S^p_k),
\] (A.3)
where $h^p_{k+1} = h^p_k + \Delta h^p_k$ (Fig. 17). We can write in a similar way the finite difference equation governing the conservation of salt of the ambient water:
\[
S^a_k \Delta z_k = \left( h^a_k - \frac{\epsilon}{\epsilon - \delta} \Delta h^p_k \right) S^a_k + h^p_{k+1} S^a_{k+1}
+ \frac{\delta}{\epsilon - \delta} \Delta h^p_k S^p_k,
\] (A.4)
where $S^a_k$ is the new ambient salinity at the kth level after the convective element has moved to the (k + 1)th level. After substituting in the above equation $h^a_k$ by $\Delta z_k h^p_k$ (recall that the vertical column is composed of two water masses with thickness $h^a_k$ and $h^p_k$, Fig. 6) we obtain:
\[
S^a_k \Delta z_k = \left[ \left( \Delta z_k - h^p_k - \frac{\epsilon}{\epsilon - \delta} \Delta h^p_k \right) S^a_k + h^p_{k+1} S^a_{k+1}
+ \frac{\delta}{\epsilon - \delta} \Delta h^p_k S^p_k \right],
\] (A.5)
which leads to the following equation for the evolution of ambient salinity:
\[
\frac{S^a_k - S^a_{k-1}}{\Delta t} = \frac{1}{\Delta t \Delta z_k} \left[ h^p_{k+1} S^a_{k+1} - h^p_k S^a_k + \frac{\delta}{\epsilon - \delta} \Delta h^p_k S^p_k
- \frac{\epsilon}{\epsilon - \delta} \Delta h^p_k S^a_k \right].
\] (A.6)

The time-indices in the right-hand side of Eq. (A.6) have been omitted.

Below we will analyze the major balances associated with the above finite-difference equations, i.e., the governing physics behind the proposed parameterization. Using Eq. (A.1’)) and the relationships between $h$ and $w$ ($h/\Delta t = h/\Delta z \Delta z/\Delta t = w$) we can write Eq. (A.6)
in differential form as
\[
\left( \frac{\partial S^a}{\partial t} \right)^c = \frac{\partial w S^a}{\partial z} + \frac{1}{\varepsilon - \delta \frac{H}{c}} \left( \delta S^p - \kappa S^a \right),
\]
(A.7)
where \((\partial S^p/\partial t)^d\) is the evolution of salinity during the convective step. This evolution equation, which is simply the alternative of the convective adjustment in MOM, states that time change of ambient water properties is due to imbalance between convergence (divergence) of vertical flux, and detrainment minus entrainment fluxes. It is the analogue to the finite difference Eq. (A.4). If we formally assume \(S=\text{const}\), we obtain Eq. (A.1”).

The right-hand side of Eq. (A.7) gives the balance between entrainment and detrainment, from one side and the vertical gradient of salinity flux, from another. For ease of interpretation, this term can be written as
\[
w \frac{\partial S^a}{\partial z} + \frac{\delta}{\varepsilon - \delta} \frac{\partial w}{\partial z} (S^p - S^a),
\]
(A.8)
which includes vertical transport and horizontal exchange of water between the DWP and environment. Thus the parameterization ensures: (1) vertical displacement of the ambient water, and (2) mixing of the two water types.

Eq. (A.3) can be rewritten in differential form as
\[
\frac{\partial w S^p}{\partial z} = \frac{1}{\varepsilon - \delta \frac{H}{c}} w \left[ \varepsilon S^a - \kappa S^p \right],
\]
(A.9)
which is a simple diagnostic relationship corresponding to the finite difference Eq. (A.2). Using Eq. (A.1”) we can rewrite Eq. (A.9) as
\[
\frac{\partial S^p}{\partial z} = \frac{\varepsilon}{H \varepsilon - \delta} [S^a - S^p],
\]
(A.9’)
where \(Z = H(\varepsilon - \delta)/\varepsilon\) is the scale of adjustment of \(S^p\) to \(S^a\). The comparison with the theoretical case (Eq. (8)) reveals the consistency of this parameterization. From this and from the numerical simulations (Fig. 7) it follows that the case \(\varepsilon > \delta\) leads to shallow penetration of the signal, which actually corresponds to the standard convective adjustment in MOM. The shallower convection in the case of stronger entrainment supports the results of Gregory and Rowntree (1990). The specifications of \(\varepsilon\) and \(\delta\) obviously requires a deeper physical analysis in the case where \(\varepsilon < \delta\), and this issue will be addressed in future work. For the purposes of the present study (which focuses on the deep ventilation and not on the shelf mixing), we simply search for the “best” combination of parameters, which ensures that the simulations correlate with the observations. The examples given in Fig. 7 demonstrate that a large variety of combinations would be possible with quite different possible solutions (including substantial change of salt content in the basin). It is interesting to note that the case \(\varepsilon < \delta\) can also give a stratification which is close to the observed one. In this case, however, the simulated plume penetrates very deeply. Since the volume flux decreases exponentially with depth in this case, there are not large intrusions into the deep layers and the stratification is not severely affected. It is noteworthy that, although the intrusions into the deep layers are small in the case \(\varepsilon = 1\), \(\delta = 3\), the DWP results in a slight increase of deep-water salinity compared to the case when \(\varepsilon = 3\), \(\delta = 1\), as well as compared to observations. Because there are no tracer and other data supporting the current penetration of signals deeper than 500–600 m, we carry out the simulation in this paper with \(\varepsilon = 3\), \(\delta = 1\).

From Eqs. (A.7) and (A.9), we obtain for the ambient salinity the following prognostic equation:
\[
\left( \frac{\partial S^a}{\partial t} \right)^c = \frac{\partial w (S^a - S^p)}{\partial z}.
\]
(A.10)

The right-hand side of Eq. (A.10) gives the divergence of salinity (or in more general terms buoyancy) flux. The full evolution of the ambient salinity is obtained after integrating in time the remaining flux terms. If for simplicity we consider the diffusion as the only remaining term, we have
\[
\left( \frac{\partial S^a}{\partial t} \right)^d = \frac{\partial \left( A(\partial S^a/\partial z) \right)}{\partial z},
\]
(A.11)
where \((\partial S^a/\partial t)^d\) is the evolution of ambient salinity due to diffusion (or remaining terms). If we assume that we have in the stationary case a balance between diffusion and convection, \((\partial S^a/\partial t)^c + (\partial S^a/\partial t)^d = 0\), i.e., the ambient water is free to adjust to the dynamics governing the
DWP, we obtain from Eqs. (10) and (11):

$$w (S^p - S^a) = K_v \frac{\partial S^a}{\partial z}.$$  \hspace{1cm} (A.12)

Eq. (A.12) is written under the assumption that the velocity at the outflow depth is small, leading to zero integration constant. This important equation states that the ambient fluid is not passive. Obviously, the largest intrusions are observed (in the stationary case) in the layers where the vertical flux due to diffusion is strongest (in the case $K_v = \text{const}$ where the stratification is strongest). This conclusion supports the dominating physical balances, which are known from theory and observations.

Using Eq. (A.9) we can rewrite Eq. (A.10) in the form:

$$\frac{\partial S^a}{\partial t} = \frac{\partial}{\partial z} \left( w H \frac{\varepsilon - \delta}{\varepsilon} \frac{\partial S^p}{\partial z} \right).$$  \hspace{1cm} (A.13)

We obtain then in the stationary case

$$K_p \frac{\partial S^p}{\partial z} = -K_v \frac{\partial S^a}{\partial z},$$  \hspace{1cm} (A.14)

where $K_p = w H (\varepsilon - \delta / \varepsilon)$. Eq. (A.14) displays the most important balance in our parameterization, which is between the salinity of DWP and of the ambient fluid. The ratio between the two gradients is proportional to the ratio between $K_p$ and $K_v$, the latter giving the dynamics. Because this ratio is a function of $w$, and because $w$ increases/decreases exponentially in the case of dominating entrainment/detrainment (in the case $\varepsilon < \delta$, $H < 0$) the slope of the two salinity profiles could differ substantially. This is well pronounced in the entrainment dominating case where $K_p \to 0$ in the deep layers. We again recall that, although this is a simple parameterization, it is very different from the standard convective adjustment where the unstable fluid can only homogenize in the vertical (Eq. (3)).

References


