Barotropic and baroclinic oscillations in strongly stratified ocean basins

Numerical study of the Black Sea

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Abstract

We investigate barotropic and baroclinic oscillations in strongly stratified basins using the Black Sea as a test case. The GHER 3D-model, which has active free surface and temperature and salinity fields as scalar state variables produces model results which are then analysed with a focus on barotropic and baroclinic waves at different scales. The model is forced by seasonally variable climatic data and river run-off. High frequency oscillations of the sea level simulated by the model are compared against observations. It is found that phases and amplitudes are simulated realistically. For the density field, long internal gravity waves dominate the solution in the sub-inertial range. The amplitudes of these oscillations increase over the continental slope, which provides an efficient mechanism for mixing in the western Black Sea. It is found that the sea surface oscillations interact with the oscillations in the pycnocline. This interaction could contribute to a modification of the vertical stratification in a long run. The vertical stratification, on its side, jointly with the bottom relief causes different appearances of oscillations over the continental slope and in the basin interior. Changes in the stability of stratification, caused by the seasonal cycle, are thus an important factor modifying wave processes and the resulting internal mixing.

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1. Introduction

When studying a given ocean system, scientists often decide to focus on processes by simplifying the real situation or to use all available data and forcings to produce the most coherent picture of the sea. These two approaches (named as diagnostic and metagnostic by Nihoul and Beckers (1992)) have their advantages and disadvantages. So is a metagnostic modelling approach of interest for interdisciplinary studies but leads to difficulties in interpretations of the simulation results. On the other hand, process-oriented studies allow a clear identification of the fundamental mechanisms, but lack several interactions with processes not taken into account. In the case of the physical oceanography of shelf seas...
Numerical simulations could be used to address issues, typical for continuously stratified ocean (with an infinite number of internal Kelvin waves, corresponding to the infinite number of the vertical modes), which are not easily addressed in the simplistic models. Some of these issues which are focal points of the present study are: (i) regional properties of the shelf oscillations; (ii) variability and the different appearances of the shelf waves throughout the year, depending on the seasonal changes in the forcing and stratification.

We will use the GHER 3D numerical ocean circulation model as a research tool in simulating shelf dynamics and some related oceanographic problems. The test area which we choose is the Black Sea. There are a number of motivations for this choice. The Black Sea is a relatively small sea, which makes it possible to use fine resolution without requiring large computer resources. Its closed boundaries facilitate imposing correct boundary conditions (zero heat flux at the bottom and at the coast, for instance), compared to the open sea boundary conditions, which are practically unknown. The closed boundaries provide also natural physical conditions for the propagation of coastal and shelf waves, which makes the Black Sea an appropriate area for studying processes on the shelf. This deep basin has a large variety of topography, from almost flat 200 km wide shelf in the northwestern part, to almost vertical continental slope along the Turkish coast (Fig. 1). The shelf and continental slope can be found below more than 60–70% of its surface, which makes the Black Sea a very good candidate for test studies on the coastal, shelf and continental slope processes.

One important external forcing in the ocean affecting Sea Surface Height (SSH) is the river discharge. Fresh water input due to rivers (300 km$^3$ per year, Unluata et al., 1989) amounts to about 0.1% of the Black Sea volume, which proves that this forcing is of utmost importance for the circulation. This sea is one of the most representative oceanographic examples for a fresh water controlled basin. Due to the very large river discharge and to the reduced exchange with the neighbouring Mediterranean Sea, the vertical stratification is extremely stable (salinity increases from 17 psu at sea surface to 22.3 psu at about 500 m). We will demonstrate in the present paper that the set-up of the model takes this forcing into consideration in an adequate way.

The interaction between barotropic and baroclinic processes is of constant oceanographic interest. The model used in the present study has an active sea surface and temperature and salinity as thermodynamic variables. It thus provides adequate model physics to address the interactions between barotropic and baroclinic waves, specially when ventilation processes can affect waves (winter ventilation, for example, seems to damp waves). As shown recently by the numerical simulations of Chen and Beardsley (1995), vertical stratification in the ocean contributes substantially to changing interaction between waves and bottom topography, which might drastically affect dynamics on the shelf/continental slope. The same issue, addressed for basins with extremely strong stratification (e.g., the Black Sea), could present an interesting test case.

Free oscillations of SSH, including shelf and topographic waves, have not yet been addressed in the
published literature for the Black Sea using simulations with continuously stratified models (neither with a theoretical, nor with a numerical model). Tidal oscillations, along with a rather wide spectrum of barotropic oscillations, have been studied for the first time by Engel (1974). However, most of the existing models for the shelf processes in this sea are barotropic, or two layer models (Archipkin et al., 1989; Blatov and Ivanov, 1992; hereafter BI; Demirov, 1994). This gives us the motivation to use a more adequate model (3D primitive equations model) capable to simulate barotropic, as well as thermodynamic processes governing a large class of waves in the Black Sea and to study their impact on the circulation. Experimental evidence that this impact might be important already exists. As shown by BI, who analysed data series for SSH and temperature, oscillations of the halocline in the coastal area are accompanied with energy shedding towards the open sea by generating long internal waves (with a wavelength of the order of the Rossby deformation radius). Whether this could present an efficient mechanism for internal wave generation in the Black Sea is not very clear, therefore, we address this issue as well, and base our analyses on model data and comparisons with observations.

After the introduction, we describe the numerical model and its implementation to the Black Sea. Then, we analyse model simulations of the horizontal circulation patterns, short periodic oscillations, the appearances of barotropic and baroclinic oscillations at frequencies of about one to several days, and finally the impact of the short periodic oscillations on the seasonal variability simulated in the model. The paper ends with a short summary of the results.

2. The implementation of the GHER 3D numerical model for the Black Sea

2.1. Description of the model

The GHER 3D numerical model (Nihoul et al., 1989) with the numerical implementation of Beckers
The model solves for the free surface, the three components of the current field, temperature, salinity and turbulent kinetic energy. This quantity is used for the computation of vertical diffusion coefficient through the classical $k-l$ model. The main equations of the model read:

$$\nabla \cdot \mathbf{u} + \frac{\partial u_3}{\partial x_3} = 0$$  \hspace{1cm} (1)

$$\frac{\partial \mathbf{u}}{\partial t} + \mathbf{u} \cdot \nabla \mathbf{u} + u_3 \frac{\partial \mathbf{u}}{\partial x_3} + f e_3 \otimes \mathbf{u} = - \nabla q + F_u + \frac{\partial}{\partial x_3} \left( \frac{\partial \mathbf{u}}{\partial x_3} - \frac{\partial \mathbf{u}}{\partial x_3} \right)$$  \hspace{1cm} (2)

$$\frac{\partial q}{\partial x_3} = b(T, S, p)$$  \hspace{1cm} (3)

$$\frac{\partial T}{\partial t} + \mathbf{u} \cdot \nabla T + u_3 \frac{\partial T}{\partial x_3} = F_T + \frac{\partial}{\partial x_3} \left( \frac{\partial T}{\lambda} \right)$$  \hspace{1cm} (4)

$$\frac{\partial S}{\partial t} + \mathbf{u} \cdot \nabla S + u_3 \frac{\partial S}{\partial x_3} = F_S + \frac{\partial}{\partial x_3} \left( \frac{\partial S}{\lambda} \right)$$  \hspace{1cm} (5)

$$\frac{\partial k}{\partial t} + \mathbf{u} \cdot \nabla k + u_3 \frac{\partial k}{\partial x_3} = \bar{v} M^2 \left( 1 - R_l \right) - \frac{k^2}{16 \nu} + F_k + \frac{\partial}{\partial x_3} \left( \bar{v} \frac{\partial k}{\partial x_3} \right).$$  \hspace{1cm} (6)

The mathematical model is a classical 3D primitive equation model. It describes the mass conservation (Eq. (1)) by using the Boussinesq approximation. The operator $\nabla$ is the horizontal version of the classical Nabla derivation operator. The horizontal component $u$ of the 3D velocity vector $u + u_3 e_3$ is computed by Eq. (2). This equation as well as the advection–diffusion Eqs. (4) and (5) for temperature $T$ and salinity $S$ take into account vertical diffusion which is computed by the means of a turbulent closure scheme based on a turbulent kinetic energy Eq. (6). A sub-grid scale parameterization is included in horizontal directions. We note the diffusion terms in Eqs. (2), (4) and (6) formally as $F_u, F_T, F_S, F_k$. Furthermore, the small aspect ratio of the vertical and horizontal scales allows the use of the hydrostatic approximation (Eq. (3)). In this case, the pressure field is readily computed once the sea surface elevation $\eta$ and the density field $\rho$ are known. Rather than to use pressure $p$ and density, which both include the strong signal of the uninteresting unperturbed reference ocean at rest, one prefers the definition of a reduced pressure $q = p/\rho_0 + g x_3$ and buoyancy $b = g(\rho - \rho_0)/\rho_0$. This buoyancy is computed by a standard equation of state for seawater in function of the model’s temperature and salinity. The parameterization of sub-grid processes in the model uses the following equations:

$$\bar{v} = \frac{1}{2} \sqrt{\bar{k} l_0 (1 - R_l)}$$  \hspace{1cm} (7)

$$\lambda = \gamma \bar{v} \left( 1 - R_l \right)$$  \hspace{1cm} (8)

$$R_l = \frac{N^2}{M^2}$$  \hspace{1cm} (9)

$$M^2 = \left\| \frac{\partial \mathbf{u}}{\partial x_3} \right\|$$  \hspace{1cm} (10)

$$N^2 = \frac{\partial b}{\partial x_3}$$  \hspace{1cm} (11)

The vertical turbulent viscosity in Eq. (7) and turbulent diffusion in Eq. (8) are computed by using the turbulent kinetic energy $k$ and taking into account the stratification effect through the definition of the flux Richardson number $R_f$ (Eq. (9)) and the Richardson number $R_i$ (Eq. (10)). Here, a neutral mixing length $l_0$ is to be specified by the modeller rather than by an additional turbulent closure transport equation for a combination of $k$ and $l$ as stated by Mellor and Yamada (1982).

The numerical resolution of the coupled set of non-linear partial differential equations is based on a vertical co-ordinate change (Kasahara, 1974; Beckers, 1991; Deleersnijder and Beckers, 1992). The advection scheme itself is a TVD method, discussed, for example, by James (1996), and known for its rather low numerical dispersion and diffusion. This small numerical diffusion is, of course, of prime importance for the simulation of oscillating systems.
otherwise damped out by inappropriate diffusion. Another particularity of the numerical resolution is the mode-splitting technique (Beckers, 1991; Killworth et al., 1991) which is necessary since the model has a free surface which needs a very small time-step for the computation of the barotropic mode. To decrease the computational burden due to this time-step restriction, the baroclinic 3D part is solved by a larger time-step than the barotropic 2D part. Similarly, the vertical grid spacing can be quite narrow due to the co-ordinate change. To avoid numerical instabilities or excessively small time-steps, an implicit scheme is used in the vertical direction.

The model has been used for modelling experiments in different ocean areas at different scales, ranging from local coastal simulations at the Belgian coast (Beckers and Van Ommeringhen, 1994) and on the shelf of Gulf of Lion (Beckers, 1995) to seas comparable to the Black Sea (i.e., the Mediterranean Sea, Beckers, 1991; Beckers et al., 1997). An extensive analysis of the model performance in simulating the Black Sea circulation and intercomparison with other model results for the Black Sea is given in our papers Stanev and Beckers (1998). Its application to a biological modelling of the Black Sea nitrate cycle can be found in the paper by Gregoire et al. (1997).

The model domain covers the whole sea (Fig. 1) with a horizontal resolution of 15 km. Here, the implementation of the vertical co-ordinate change is such that the Black Sea is cut into two regions superposed vertically. One region covers the deeper parts of the sea (called, hereafter, deep region) and a second region covers the shelf and the region above the deep region. In each of these regions a classical \( \sigma \)-co-ordinate change is introduced (Fig. 2a and b).

The model uses 15 non-uniformly distributed vertical levels in the first \( \sigma \)-region and 10 levels in the deep region. In the regions of minimum depth over the shelf, the vertical discretisation is done with an accuracy of 0.9 m. The distance between \( \sigma \)-layers increases in the basin interior (Fig. 2a), reaching at most a distance of 18 m in the upper \( \sigma \)-region (over the deepest part of the basin). The justification of using this double \( \sigma \)-co-ordinate system is illustrated in Fig. 2b. If we had only one \( \sigma \)-layer, then the gridlines would follow the slope. It is well-known, however, that the slope of the \( \sigma \)-levels has not to exceed some limits in the models (Deleersnijder and Beckers, 1992; Oguz et al., 1995). As seen from Fig. 2 with the particular choice of two \( \sigma \)-regions, we reduce drastically the slope of the \( \sigma \)-levels. A number of preliminary simulations were carried out, aiming to analyze and reduce the errors resulting from using a \( \sigma \)-co-ordinate system in the regions of strong slopes and stratification. In order to have a model topography consistent with the model resolution, we smooth the depth \( H \) of each region, so that the factor \( H^{-1}\delta H \) remains small enough in each layer. As shown by Oguz et al. (1995), more effective use of the advantages of the \( \sigma \)-co-ordinate system could be achieved by using a variable resolution in the horizontal direction. We partially compensate for the lower horizontal resolution, compared to the quoted work, by implementing the double \( \sigma \)-co-ordinate system in the vertical direction.

We have to mention one important consideration, related to the use of \( \sigma \)-co-ordinate system. The vertical density gradients in the Black Sea are larger than in the ocean. Therefore, special care has to be taken to avoid errors resulting from the pressure gradient formulation if both topography is very steep and the stratification is very strong. If the model is initialised with climatic data for temperature and salinity, which are normally not adjusted to the model topography, erroneous model trends are developing. With the same model applied to the Mediterranean Sea (a basin with much weaker vertical stratification), such trends have not been observed. To initialise the Black Sea model, we, thus, used horizontally homogeneous fields, reducing the magnitude of the spurious currents, which usually develop, below 1 cm\(^{-1}\).

The method of time integration (the mode splitting) consists of using many small time-steps for the barotropic part, during which the baroclinic part is assumed to be frozen. After \( \Delta t_{3D}/\Delta t_{2D} \) integrations of the equation for the SSH \( \eta \), where \( \Delta t_{3D} \) and \( \Delta t_{2D} \) are the time-steps for the baroclinic velocity and for the barotropic part, respectively, one baroclinic time-step is performed. In our particular case, \( \Delta t_{3D} = 0.5 \) h, \( \Delta t_{2D} = 45 \) s.

Splitting the solution into barotropic and a baroclinic parts, as described by Killworth et al. (1991), necessitates to include barotropic friction, which makes possible to have an explicit control on the
amplitude of SSH oscillations. Calculating friction for the barotropic part at each barotropic time-step provides some horizontal smoothing if the amplitudes of SSH increase due to barotropic dynamics. The damping of these processes is done by using a barotropic non-linear bottom friction. Horizontal sub-grid scale processes are parameterized by Laplacian term with coefficients for the heat...
and salt of 50 m$^2$ s$^{-1}$ and for momentum of 500 m$^2$ s$^{-1}$. These values have been chosen to be consistent with the coefficients used in previous studies (Stanev, 1990; Oguz et al., 1995) and after some sensitivity runs with different coefficients. Solving the equation of the turbulent kinetic energy provides model estimates which are used to calculate the vertical mixing. The background vertical mixing is kept small enough to prevent unrealistic erosion of the halocline and of the Cold Intermediate Layer (CIL).

2.2. The Black Sea topography and the initial data

The shelf area covers 25% of the whole Black Sea surface (Fig. 1) and presents a slightly tilted plane with depths between 0 and 100 m. It is very wide (about 200 km) in the northwestern part of the sea, and practically vanishes (width from several kilometres to 10 km) along the southern coast. The continental slope reaches maximum values of about 20–30% along Caucasian and Anatolian coasts. The areas with depths between 100 and 2000 m cover 40% of the Black Sea surface. The abyssal plane in the Black Sea covers only 35% of its surface. The high percentage of slope areas shows one very important difference between the Black Sea and the ocean. Processes controlled by topography as JE-BAR or topographic Rossby waves are very significant over at least half of the basin. Different numerical approximations of the shelf and continental slope in the model could result in simulating quite different regimes of wave transformations in these areas.

To make the requirements for the resolution of shelf and continental slope more clear, we refer to BI, who estimated the length of the shelf waves as a function of the shelf width for several typical sections. For the wavelength along the Crimean and Anatolian coasts, they give the following ranges: 116 km (shelf width about 46 km), and 18 km (shelf width 3 km). The same authors estimate the Rossby radius of deformation to be about 20 km. This proves that the specific continental slope and shelf in the Black Sea could impose stronger restrictions on the model resolution than the Rossby radius of deformation.

The impact of increasing resolution on the estimates of SSH oscillations is addressed by Demirov (1994), who used boundary fitted curvilinear coordinate system and different resolutions on the shelf and in the open sea. In the recent study of Oguz et al. (1995), another example of curvilinear system, well-fitted to the Black Sea boundary, is given (like in the present study, these authors use a $\sigma$-co-ordinate model, but they do not address the issue of the free oscillations). Even with better horizontal resolution, the model of Oguz et al. (1995) does not resolve accurately enough the waves along the southern coast, and smoothing of the slope or putting vertical walls in some areas was necessary to avoid inconsistent results. In our case, with the combination of double $\sigma$-co-ordinate system and 15 km horizontal resolution, we hope to sufficiently resolve the lowest wave modes, except perhaps for the southern coast, where the model does not sufficiently resolve the continental shelf waves.

Vertical profiles for temperature and salinity, used to initialise the model (Fig. 3), illustrate the main characteristics of the vertical stratification: a sharp halocline from the sea surface down to 200 m and the CIL (Cold Intermediate Layer), analogous to the 18°C water layer in the Atlantic Ocean. This CIL is caused by extreme cooling of the western Black Sea in winter and penetrates advectively throughout the whole sea. The deep layer (below 1000 m) is very homogeneous, with temperatures about 9°C and salinity about 22.3 psu. There are indications that these deep waters were formed a long time ago, remaining presently unaffected by the ventilation from above (Ozsoy et al., 1993).

2.3. Model forcing

All sea surface boundary conditions are taken from monthly mean data and interpolated linearly for each model time-step. The wind stress data are based on ship observations (see Staneva and Stanev, 1998). The annual mean wind stress curl shows a pronounced cyclonic circulation with basin mean maximum in winter and minimum in summer. Sea surface temperature and salinity in the model are relaxed to time dependent climatology. The motivations for using this type of boundary conditions and its compatibility with the other boundary conditions in the model are given in detail later in text.

Since the present paper addresses the free surface oscillations, we have to give some information on the formulation of boundary conditions for the barotropic part. Increased river discharges in spring
Fig. 3. Basin averaged temperature (a) and salinity (b) vertical profiles.
Concerning the water budget, the conservation law reads
\[ \frac{\partial \eta}{\partial t} + \nabla \cdot U = q \] (13)
where \( \eta \) is the SSH, \( U \) is the transport and \( q = P - E \) is the net water import due to local rain (\( P \)) and evaporation (\( E \)).

Rigid lid models neglect the first term and the right hand side, whereas classical free surface models neglect the right hand side only. Here, we have to keep the right hand side, if we want to retrieve a flux in the Bosphorus Strait coherent with the river run-off and the net \( P - E \), as they are all of the same order of magnitude. Indeed, the flux in the Bosphorus Strait \( Q \) is self-adapting to the interior evolution by the following rule derived from Oguz et al. (1990):
\[ Q = \alpha \eta_B + Q_B \] (14)
where \( \eta_B \) is the SSH at the Bosphorus Strait, and \( Q_B \) is the barotropic transport when \( \eta_B = 0 \), the constants \( \alpha \) and \( Q_B \), being estimated from the study of Oguz et al. (1990).

The precipitation and evaporation over the basin must, thus, be added or subtracted from the sea surface elevation. Otherwise, the Bosphorus flux would simply tend towards the value of the river discharges, which is, of course, not the case in reality.

If we denote by \( \bar{\eta} \) the basin average of the sea surface elevation, we have indeed
\[ \frac{\delta \bar{\eta}}{\delta t} = -\alpha \bar{\eta} - \alpha \delta \eta_B - Q_i + v \bar{q} \] (15)
where \( \dot{} \) stands for the basin average, \( Q_i \) for river inputs, \( s \) for the surface of the sea and \( \delta \eta_B \) for the anomaly of SSH at the Bosphorus Strait due to internal dynamics of the Black Sea. From this equation, it is clear that the system is self-adjusting towards a situation, where all the water fluxes are giving a zero budget. Furthermore, it is seen that the adjustment is done with a time scale of \( s \alpha^{-1} \), which is about 45 days. From there, it is also clear that a delay between changes in river discharges and fluxes through the Bosphorus Strait is to be expected.

The net local precipitation or evaporation could also be used for the salt budget, by using this information directly in the salt equation as proposed.
by Huang (1993), where the height of the grid box is modified without changing its total salt content. In the case of the generalised $\sigma$-co-ordinate change, this can be easily achieved by modifying the height of the top-most layer by adding $q$ locally to $\eta$. Then, the salt conservation would be satisfied. Another way is not to change the height of this layer for the salinity component, but to add an equivalent virtual salt flux at the surface. Mathematically, these formulations are exactly identical, but in practice, due to different time-discretisations, this could lead to problems mentioned by Huang (1993).

In our case, the estimates for $q$ can certainly be used to close the basin wide global budgets, but their use as forcing for the salinity field is much more disputable due to the lack of error estimates of the precipitation and evaporation fields.

Some available data sets (Sorkina, 1974; Golubeva, 1984) are produced using various assumptions and extrapolations in space, which could give arguments for speculations on the quality of simulated results. This is why the relaxation procedure towards climatological sea surface salinity fields has been used. In this case, the virtual flux formulation can be used without any problem, since there is always a stabilising feedback towards climatology. The method simply adds thus a surface flux counted positive upwards to the surface layer:

$$-\lambda \frac{\partial S}{\partial x_3} = c(S - S^*)$$

(16)

where $S$ stands for the model salinity at the surface, $S^*$ for the climatological field and $c$ gives the strength of the relaxation. In a model with horizontally constant vertical grid spacing, this is equivalent to using a relaxation in the surface box with a time scale of $\Delta z/c$, but the flux formulation has the advantage of introducing sources which do not depend on the vertical resolution. The relaxation is kept small in the present case by using a value of 1 m per day for $c$. Since we impose a salt flux, the change of the water level due to evaporation–precipitation $q$, used in the water mass conservation, must not be taken into account in the salt budget. The value of $q$ is, thus, taken into account only in the barotropic dynamics. One could argue whether the combination of the boundary conditions for salinity at the sea surface and for the river discharge (the last condition acts as a typical natural boundary condition for salinity) are compatible. The formal answer is negative, but we should keep in mind that model relaxation is very weak, providing only a general trend of the sea surface salinity toward lower values, rather than to the specific horizontal patterns, corresponding to the climatic data. As we will show later in the paper, the frontal area in the western Black Sea is practically a consequence of the river inflow and not a result of the model relaxation. The detailed study on this subject done by Stanev and Beckers (1998) proved that the combination of the two types of boundary conditions does not create physical inconsistencies in the system with the parameters specified above.

Observational data show that the Black Sea SSH oscillations consist of wide spectrum of time–space properties, related to different physical processes and to different forcing. Sometimes forced oscillations are strongly amplified and mask the free oscillations. In this study, we chose to force the model with relatively smooth in time forcing functions to facilitate the understanding of free oscillations. This is in order not to obscure the physical interpretations by the effects resulting from using external forces from different origins, or quality. Therefore, we do not consider tidal oscillations (these oscillations show much less amplitude in the Black Sea than in the Adriatic Sea, for instance), or oscillations forced by meteorological processes with short time scales (as synoptics). For the response of the Black Sea model to high frequency forcing, see the paper by Stanov et al. (1995). On the other hand, forcing the Black Sea model with a seasonally variable data is of utmost importance, since this provides the physical mechanism maintaining the stratification over many years of integration. This is important when studying the effects of short time variability on the processes with much longer time scales. Otherwise, as we show in the paper, simple perturbation experiments are enough to understand some specific processes.

2.4. Numerical experiments

In the rest of the paper, we analyse the results of three numerical experiments, aiming to elucidate the
impact of surface gravity waves on the model ocean. In the first experiment, named Central (C) experiment, we approximate the bottom relief as close as possible to the real one, as this could be done with the model resolution (and the present version of double \(\sigma\)-co-ordinate change). In the second experiment, named hereafter Smooth Bottom (SB) experiment, we filter the bottom by applying several times a Laplacian filter. With the intercomparison between the simulations in C and SB experiments, we aim to find how important the precise resolution of the bottom relief is for the simulated circulation (temporal variations over different bottom topography could produce different residual flows). In the third experiment, named further Free Free Surface (FFS) experiment, we do not put an explicit friction on the barotropic mode, hence, the free surface ‘feels’ completely free. We initialise experiments C and SB from the same initial data and run them for 10 years, which is enough to reach a quasi-periodic state. For more details on the determination of the spin-up time for the Black Sea see Stanev 1990 and Oguz et al. 1995.

The FFS experiment is initialised with the results of experiment C after 10 years of integration, and both models are integrated further for 5 years to provide data needed to intercompare the simulated physical processes. Additional perturbed runs, based on the FFS experiment, aim to more clearly illustrate the oscillations in the SSH and their impact on the circulation.

### 3. Analysis of the simulated circulation

The vertical stratification and the circulation in the sea provide the physical background for the formation and propagation of oscillations. On their side, waves can also contribute to changing circulation and stratification. Since the present study addresses the simulation of oscillations in the Black Sea, it is important to first give an illustration of the model performance in simulating the circulation in this basin. This issue has previously been addressed for different freely available GCM-s: Stanev (1988, 1990), giving details on the performance of Bryan and Cox model (Bryan, 1969; Cox, 1984), and Oguz et al. (1995) giving details on the application of Princeton Ocean Model (POM) to the Black Sea. More results from the simulations with the GHER model for the same area can be found in the paper by Stanev and Beckers (1998). We illustrate in Fig. 4 the simulated annual mean fields: (i) the barotropic part, exemplified by the pattern of SSH, Fig. 4a; (ii) the mass field, exemplified by the salinity patterns in the surface layers, in the main halocline and in the deep layers, Fig. 4b–d; (iii) the baroclinic part, exemplified by the change in the currents with increasing depth (vectors superimposed on scalar plots).

It is widely recognised in the Black Sea oceanography that the SSH has a maximum in the coastal zone and decreases with increasing distance from the coast. This is well-known from computations based on the dynamic method (Filippov, 1968) diagnostic models (Gamsakhurdiya and Sarkisyan, 1976) and from the model studies quoted above. Our model simulations correlate with the above concept, giving for the annual mean maximum difference of SSH in the coastal and open sea about 16 cm. SSH reaches highest values in the Danube inflow region, and lowest values in the eastern cyclonic gyre (Fig. 4a).

Comparison of the patterns of SSH and sea surface salinity clearly shows that the increased elevation in the western basin correlates with the decreased salinity, caused by the river discharge. Actually, the salinity pattern in Fig. 4 shows much stronger contrasts than the climatic sea surface salinity, proving that the simulations are not significantly affected by the horizontal salinity pattern imposed by the relaxation boundary condition. As it will be shown in Section 4.2.1 (see the paper of Stanev and Beckers (1998) for more details), the salinity front tends to continuously change its position, leading to a smeared appearance when integrating over the year. The trend of changing direction of the front
Fig. 4. Annual mean model simulated results with currents patterns superimposed, C-experiment. (a) SSH (cm), (b) salinity at 5 m (psu), (c) salinity at 150 m (psu), (d) salinity at 500 m (psu).
from south to southwest (Fig. 4b) persists over most of the year and correlates with the concept that river fronts usually follow the coasts. The decrease in the salinity gradient across the front with the increasing distance from the Danube river mouth agrees well with the existing observations. The front can be traced by the low salinity values (about 17.8 psu) down to the Strait of Bosphorus. It can also be identified on the SSH pattern, where the SSH in the Bosphorus inflow area is only about 2 cm lower than in the Danube discharge area (Fig. 4a).

Isohalines tend to follow the continental slope at the depths of the main halocline (compare Fig. 4c with Fig. 1). Currents decrease significantly with depth, and are of about 2 cm s\(^{-1}\) at 150 m. The general transport pattern reveals most of the features observed in the surface layers. However, one important difference is worth noting: the southern part of the western basin is dominated by an anticyclonic circulation. The northern part, including the continental slope area, is dominated by pronounced cyclonic transports. Both circulation elements merge, forming an intense jet which crosses the sub-basin. As it will be shown later, this direction coincides with the direction of propagation of the model long internal gravity waves. More extended analyses on the temporal variability of simulated patterns and on their correlation with other estimates are given in the paper by Stanev and Beckers (1998).

4. Free oscillations

4.1. Phase and amplitude characteristics of short periodic oscillations of the free surface

Analyses of data from observations in the eastern Black Sea indicate that short periodic oscillations contribute to less than 8% of the potential energy of oscillations (BI). Oscillations in the synoptic range (forced by the synoptic processes in the atmosphere) give another 40% of the total energy. The remaining energy, that means more than 50%, is contained in the seasonal variability. We analyse in this section the simulations in the FFS experiment, focusing on the oscillations of SSH with periods from several hours to about 1/f day. As it will be shown later in the paper, the most energetic high frequency oscillations in the model have periods below 12 h. These oscillations are usually dominated by barotropic processes, including Kelvin waves. We will refer to the oscillations considered in this section as to high frequency oscillations.

Looking for solutions of the spectral problem in the Black Sea, Engel (1974) focused on the gravitational modes ($\omega > 10^{-1}$ cph). The lowest gravitational mode (when $f = 0$) is a standing wave with a nodal line passing along the shelf break in the northwestern part of the sea. Free oscillations in the deep part of the sea have been studied later by Demirov (1987), who took the continental slope as a model coast. He estimated for the one nodal wave, running along the long axis of the sea, a period of 4.63 h, which finds strong support in the observations. Oscillations with higher frequencies (period of 2.3 h) simulated in barotropic models, and found also in the observations, appear to be very stable. These oscillations are characterised by two nodes with amphidromic centres in the western and eastern Black Sea sub-basins. We aim to validate our simulated wave appearances against observations and to provide intercomparisons with the results of such earlier model studies.

In order to analyse the wave spectrum produced by the model, we use an initial state in which the SSH is perturbed over the whole sea. The perturbed SSH is created either by taking a SSH elevation field from a snapshot at another moment or by adding to the adjusted SSH a random perturbation with a rms of a few centimeters. Both ways of perturbing the system gave similar results. In this way, the adjustment between the new initial SSH and the rest of variables taken from the restart file in the moment of the perturbation initiates oscillations in the SSH since it is not anymore adjusted to the velocity fields. The model starts, thus, from this perturbed state and produces outputs of the SSH over the entire Black Sea. To more accurately look in the spectral and phase/amplitude properties of the simulated data, and in order not to process unreasonably large data sets, we average the model output for different time intervals. The spectral curve corresponding to 1 h averaged data (Fig. 5a) shows maxima at 2.1, 4.36, and 6.4 h, which are typical periods of free surface oscillations in this sea (see Table 1). As known from observations (BI), the amplitude of barotropic seishes
and of the barotropic Kelvin waves in the Black Sea are of about several centimeters. This indicates that our perturbation and the amplitudes of the resulting oscillations are in the range close to observations. With 15 min time averaging of the simulations, we find additional spectral maxima in the higher frequency range (at 1.2 and 0.62 h, not shown here). The magnitudes of the very short periodic oscillations are about 3–4 times smaller than of the 2.1 h oscillations. Observational evidence that oscillations with such periodicity exist along the Bulgarian coast is given by Belberov and Kostichkova (1979) and Mungov (1984).

We calculate amplitudes and phases for each frequency using a fast Fourier transformation. The corresponding amplitude and phase characteristics of the 2.1 h oscillations (Fig. 6a and b) show a very good correlation with the results of Baklanovskaya et al. (1986) and Demirov (1987). This wave runs along the long axis of the sea, having two nodal lines. The first nodal line is between the Turkish coast and the western coast of Crimea Peninsula. The second one is in the eastern Black Sea, and has a significant slope to the southwest. Both nodal lines separate the sea in three large sub-areas.

Highest amplitudes, usually observed along the Bulgarian coast (between Varna and Burgas, see BI) are supported by the model simulations (solid lines in Fig. 6a). Oscillations in this area have the same phase as the oscillations in the eastern most Black Sea (Fig. 6b). The third area with maximum amplitude is simulated in the central Black Sea, but the amplitudes there are lower than in the coastal regions (Fig. 6a).

The patterns of the amplitude-phase characteristics with 2.1 h period, simulated in the northwestern Black Sea, have smaller length scales than in the basin interior. We find very high amplitudes to the west of Crimea Peninsula, along the Dnepr river mouth and south of the Danube delta. This correlates well with the observations documented by BI. The amphidromic area located south of the Danube delta (Fig. 6b) is characterized by an anticyclonic rotation.

<table>
<thead>
<tr>
<th>Period</th>
<th>0.62</th>
<th>1.2</th>
<th>2.1</th>
<th>4.36</th>
<th>6.3</th>
<th>11.4</th>
<th>16</th>
<th>24</th>
<th>28.6</th>
<th>53</th>
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<tr>
<td>η T150</td>
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*($: indicates oscillations for which this period is pronounced. (-): indicates oscillations for which this period is not pronounced.*
This is also the sign of rotation of the smallest amphidromic area (west of Crimea Peninsula). However, the rotation of the main amphidromic area, located on the shelf, is cyclonic. These results, indicating different types of rotation, support experimental evidence given by BI.

The 4.36 h maximum in the SSH spectrum corresponds to amphidromic structures with cyclonic rota-
tion, which could be found on the northwestern shelf, as well as in the central basin (Fig. 6c and d). The increase in the amplitude towards the amphidromic periphery, illustrated by the solid lines in Fig. 6c, and the phase rotation (Fig. 6d) indicate that a Kelvin wave-type structure dominates the oscillations at this frequency. We admit that for the Black Sea, where the barotropic radius of deformation is comparable to the long axis of the basin, Kelvin waves are modified by basin scales. For brevity, we will, however, refer further in text to the oscillations at 1/4.36 cph as to Kelvin waves.

The wave running along the longer axis of the sea is characterised by a narrow area of zero amplitudes. It extends in the meridional direction from the Turkish coast to the eastern coast of Crimea Peninsula. Amplitudes in the easternmost Black Sea are much smaller than along the western coast. The oscillations in the eastern and western basins have opposite phases (Fig. 6d), which corresponds to the well-known properties of the first gravitational mode for the deep part of the sea. Our patterns are similar to the ones based on the solution of the eigenvalue problem for the deep part of the Black Sea only (Demirov, 1987).

The oscillations at 1/4.36 cph show much higher amplitudes on the shelf than in the deep sea (Fig. 5c). Oscillations with amplitudes of 2–3 cm have been documented from observations along the Romanian coast (Marinsek and Sclarin, 1968; quoted from Demirov, 1987). Similar peculiarities are documented by BI, as well.

Model simulations give a clear spectral maximum at frequency 1/6.4 cph (Fig. 5b and c), which agrees with some earlier estimates (compare Tables 1 and 2). The amphidromic line, corresponding to these oscillations, is close to the continental slope. Deep basin oscillations have phases which are opposite to the phases in the northernmost Black Sea. Amplitudes in the interior Black Sea are small, and the pattern does not show significant gradients (Fig. 6f).

Unlike the deep sea, oscillations on the shelf show pronounced cyclonic rotation and very high amplitudes (Fig. 6e and f). The main difference between the oscillations at 1/6.4 and 1/4.36 cph (compare Fig. 6e and f with Fig. 6c and d) is that no energetic oscillations are simulated along the long axis at 1/6.4 cph.

Despite the similarity of our results with the earlier ones (Engel, 1974; Archipkin et al., 1989; Demirov, 1994) we will identify below some differences. Basin averaged model spectra show a lower number of spectral maxima than the previous analyses on the basin eigen modes. We remind here that we do not discuss the phase-amplitude properties of all frequencies existing in the model data, but only the most energetic ones. To illustrate the spectral variations in space and the local domination of some frequencies, we refer again to the spectral curves in some particular locations (Fig. 5b and c). In locations A and B on the shelf (see Fig. 1), the energy at 1/6.4 cph increases strongly when approaching the northernmost part of the shelf. This is not so strongly pronounced for the oscillations at 1/4.36 cph. Differently from the basin averaged spectrum (Fig. 5a), the ones on the shelf show very strong oscillations with periods between 2 and 4 h, which is not typical in the basin averaged spectra.

The diversity of appearances of spectral maxima basin-wide supports the same conclusion derived by BI from observations. This indicates that estimates based on localised observations do not give a full description of the dominant barotropic oscillations. On the other side, we do not find spectral maxima at 1/9.7 cph, which have been previously reported in other studies (see Table 2). The phase-amplitude properties at this frequency (not shown in the figure) do not give indications that these oscillations are simulated in the model. Such model indications could give motivation for future improvement of model characteristics (e.g., topography, coastal line), which might affect specific type of wave processes. Another possible source of differences between our estimates and some previous ones is that baroclinic and barotropic processes are not separated in our model. The stratification could have a significant impact on the residual effects of SSH oscillations (see also Chen and Beardsley (1995)). Model dissim-
One major difference between model data and observations is due to the fact that high frequency oscillations are not forced in the same way in the model and in the real sea. As shown by the observations, the oscillations with periods of 12 and 24 h carry substantial part of the energy in the high frequency range (BI). This type of oscillations is not effectively excited in the model due to the lack of semi-diurnal tide in the forcing (these oscillations do exist in the model, but their appearance is rather a response of free surface to internal waves, see Section 4.2). Thus, the idealised forcing gives a further explanation of the reduced number of oscillations in the model. This is coherent with the conclusion of Demirov (1994), that the smaller number of pronounced modes simulated in models in comparison to the abundance of spectral peaks in experimental data is due to the stronger excitation of the forced oscillations in the real sea. This basin response could have the character of a resonance at the forcing frequency. The issue of tidal oscillations in the Black Sea can be subject to further studies. However, the Black Sea is not a tidal basin, thus, our main interest in the present paper is focused on free oscillations, which are more important for the dynamics of this sea.

### 4.2. Barotropic and baroclinic oscillations

In the present section, we address some processes in the frequency range where the baroclinic oscillations start to dominate the solution, i.e., to show spectral peaks. Unlike the oscillations discussed in the previous section, the ones dominated by baroclinicity are characterised by time and length scales, corresponding to the periods of internal waves and to the internal Rossby radius. We remind here that with the horizontal resolution of the model we resolve the low wave numbers of baroclinic oscillations. The shape of the bottom normal to the coast controls the oscillations in the coastal zone. The factor $H^{-1} \partial H/\partial x$, measuring the possible contribution of the topographic waves is much larger than $\beta/f$ in a rather large area along the continental slope. This makes these regions a natural wave guide for coastal trapped waves. These waves dominate the experimental and simulation data stronger than the barotropic Rossby waves do (the barotropic radius of deformation for the Black Sea is comparable with its longer axis). Here, we have to mention that there are some experimental studies for the Black Sea on the oscillations with periods ranging from 0.5 day to a couple of days (Blatov et al., 1984; BI) and some theoretical estimates (Demirov, 1994), but no analyses for this sea exist on the performance of ‘layer’ baroclinic models in this frequency range. Therefore, we first give some short information about these processes.

#### 4.2.1. General analysis of the oscillations in the low frequency range

Here, our main interest will be focused on the coexistence of barotropic and baroclinic oscillations, exemplified by SSH, temperature at 150 m ($T_{150}$) and salinity at 15 m ($S_{15}$). In order to illustrate more clearly the oscillations, we do not impose additional friction on the barotropic part (FFS experiment). We introduce a small perturbation in the model in the same way as in the perturbation experiments on short periodic oscillations. We first run the model for 10 days only and produce an output for the time filtered SSH, $T_{150}$ and $S_{15}$ (averaging is done for 1 h to provide precise resolution in the high frequency range). Afterwards, we redo the integration for 4 months, producing model averaged output for the same variables with an averaging of 8 h.

The temporal evolution of SSH, $T_{150}$ and $S_{15}$ along $43.13^\circ N$ is shown in Fig. 7. To visualise more clearly the results, we subtract at each time-step the initial value. Several conclusions can be drawn from Fig. 7a–c. SSH oscillations have amplitudes of several centimeters and temperature oscillations are about 0.04°C (we have to keep in mind that temperature stratification at this depth is very weak). $S_{15}$ evolution is not dominated by similar oscillations. The amplitude of waves is much larger in the western Black Sea for both SSH and $T_{150}$. Wavelengths are much shorter than in the case of pure barotropic waves (compare with the results discussed in the previous section). Periods range from 8 h to several days.

The differences between the temporal variability of SSH and of $S_{15}$ are due to the different character-
Fig. 7. Temporal evolution along the section 43°13'N of: (a) SSH (b) T150 (c) S15. The numbers on the horizontal axis (I) are model grid points, and the ones on the ordinate (K), the time. To convert the axes on length and time, units are to be multiplied by $\Delta x = 15$ km (see also Fig. 1) and $\Delta t = 8$ h, which is the discretisation of this plot.
istic time scales of both fields. We have also to keep in mind that we show in Fig. 7 the oscillations in the cold part of the year, when the homogenisation of the vertical stratification in the surface layer tends to eliminate the internal gravity waves. They start to emerge before the end of the integration, that is, in spring (Fig. 7c). The existence of these oscillations proves that the relaxation boundary condition does not significantly damp oscillations of salinity at the sea surface.

The simulated Danube river front, traced by the difference between less saline water near the coast and water from the sea interior, shows a number of similarities with the observations and theory. Since the initial salinity distribution is subtracted from each individual value in Fig. 7c, this figure represents rather the evolution of the front. It is clearly seen that negative anomaly is formed in the coastal area, which is due to the increasing river run-off in the first part of the year. At the latitude of the cross-section (43°13′N, which is close to the latitude of Cape Kaliakra), a negative anomaly is formed in the first 3 months of the year. After this period, the plume turns to the right (shoreward), carrying less salty water further down the coast (see also Fig. 4b). Similar behaviour of river plumes has been previously simulated by Oey et al. (1985), Chao and Boicourt (1986), and Oey and Mellor (1993). The comparison of Fig. 7b and c proves that the wave-length in the oscillations of temperature and the thickness of the front are comparable. This might serve as an indication that the Rossby radius of deformation gives a characteristic length scale controlling the plume dynamics, which agrees with the results of the above modelling studies. Increase in the river discharge in spring tends to sharpen the front. This triggers non-linear processes, which tend to shift the front closer to the coast (see the above studies). This is another proof that the model simulates realistically this well-known characteristic of river plumes. Short periodic oscillations in the sea surface salinity, emerging in spring indicate that variations in the SSH start to slightly dominate the variability of the front in the warm part of the year.

Observational evidence indicate a pronounced correlation between SSH and temperature in the pycnocline. Measurements along the Crimean coast give for the period of oscillations of the isotherms the value of one day (BI). The spectral maxima in the kinetic energy at 11 h and at 42–51 h, estimated from observations, are explained by BI as being related to periodic strong displacements of the pycnocline, which usually appears once every 15–16 days. These appearances are induced by the atmospheric forcing. Thus, forced oscillations are excited at frequencies, which are very close to what the model estimates give. However, the variability simulated in the model is a consequence of the model dynamics, rather than of the variability in the external forcing. This gives some confidence that with real synoptic winds (Stanev et al., 1995) the model will accurately simulate some important features related to the internal wave dynamics.

The rms of SSH and T150 and their correlation (Fig. 8) give more insight in the model simulated oscillations. Variability in the SSH in the eastern part of the sea (Fig. 8a) is induced by the seasonal variability of the wind (see also Section 5). The triple centre area of large rms in the SSH is localised approximately in the area of the strongest variability in the wind stress curl (Staneva and Stanev, 1998). Another large area of increased variability of SSH is localised in the western Black Sea and is due to the variability of the western part of the gyre with long time scales (longer than what is studied in this section). The pattern of the SSH variance in the western-most area is rather patchy, but clearly shows a maximum of the Cape Kaliakra. Such local features are repeated several times further south. This area is known as one of the regions in the Black Sea with strongest oscillations of the halocline, resulting periodically in strong instabilities, breaking waves and upwelling phenomena (Trukhchev et al., 1985; Demirov, 1994; Sur et al., 1994). The strong variance of the SSH in these areas might serve as an indication that barotropic and baroclinic processes in this region are strongly coupled.

Temporal variability of T150 (Fig. 8b) shows at least three important features: (i) oscillations are much weaker in the eastern Black Sea than in the western coastal area, this is also clearly seen from Fig. 7; (ii) the rms pattern of T150 close to the inflow area gives some arguments to speculate that the increased variability is amplified by the inflowing Marmara Sea water; (iii) not only the Marmara effluent could be the reason of the increased variabil-
Fig. 8. Horizontal distribution of (a) rms of SSH (m), (b) rms of T150 (°C), (c) correlation between SSH and T150.
High rms values are observed in Fig. 8b, starting from the Crimea Peninsula (note the extension of the high variability area to the south of the Peninsula). This band follows the continental slope to the southwest and further to the south, along the Bulgarian coast.

The correlation between SSH and T150 are strongest in the western Black Sea, and particularly along the continental slope and in the Bosphorus inflow area. They have a negative sign, which proves that when sea surface rises the pycnocline deepens. When we make this explanation, we have in mind that the temperature above 150 m is lower than at 150 m (see Fig. 3). Over the rest of the sea, the correlation has very small values. This result motivates us to look in more detail in the interaction between barotropic and baroclinic oscillations.

4.2.2. Interaction between oscillations in the SSH and internal waves

To understand the interaction of barotropic and baroclinic oscillations, we analyse local spectral curves. Local analyses on the propagation of the waves in the vertical and along the normal to the coast can help in better understanding the coupling between barotropic and baroclinic oscillations. The subsampling of 0.5 h for a perturbation experiment of 40 days, starting at the 1st of January, ensures correct estimates of both high frequency, and sub-inertial oscillations.

We chose a cross-section along the parallel 43°27′N (shown in Fig. 1) extending from the western coast to about 450 km in the open sea. The relief along the cross-section is characterised by a narrow shelf and a steep slope (Fig. 9). We introduce as usual a perturbation in the model sea and show the oscillations of temperature during 5 days in two points along the cross-section (grid points 11 and 15, Fig. 10a and b). As seen from the bathymetric curve, point 11 is just at the foot of the continental slope, and point 15 is on the abyssal plane. It is well-known from the observations that the amplitude of the oscillations of the pycnocline increases when approaching the coast. This is supported by our model results as
Fig. 10. Oscillations of temperature (time vs. depth) at different locations along the section shown on Fig. 1. The horizontal axis gives the time (days from the perturbation), the vertical is the depth (m). Upper panel: location 11, see Fig. 9. Lower panel: location 15, see Fig. 9.
Fig. 11. Time vs. x plots of the SSH (upper panel) and T150 (lower panel) during the integration. Horizontal axis gives the location (numbers correspond to model grid points of Fig. 1). Vertical axis gives the time in days. The slope of the thick line gives the velocity of the waves.
Fig. 12. SSH (a and c) and T150 (b and d) amplitudes along the cross-section in Fig. 1 in January (a and b) and in June (c and d). Horizontal axes gives the period of waves (hours), vertical axes gives the amplitude. The length of this axis corresponds to: 0.3 cm (a), 0.003°C (b), 1 cm (c) and 0.01°C (d).
Fig. 12 (continued).
well (compare the amplitudes of gravity waves in the two points). The vertical propagation of the signal is very rapid and the oscillations in the whole water column have almost the same phase. The cooling in the coastal region is strong, which tends to significantly damp wave processes in the surface layer in point 11.

The interaction of internal and surface gravity waves in the coastal area is well illustrated by the parallel examination of the oscillations of SSH and T150 on the zonal cross-section (Fig. 11). From the comparison between the two figures it becomes clear that the spectral composition of SSH is much more diverse than that of T150. Some of the frequencies dominating the oscillations in the high frequency range were examined in Section 4.1. This signal is coherent along the cross-section indicating that different modes of barotropic oscillations reach this cross-section simultaneously. Obviously these are the barotropic Kelvin waves. For these waves, the SSH oscillations have the same phase along the whole section line, which is directed approximately from the coast towards the amphidromic point in Fig. 6d.

It is noteworthy that the same regime of oscillations at the sea surface dominates the solution from the amphidromic centre up to the longitude of the 17th point that is about 100 km from the continental slope. The free surface oscillations in the area of the continental slope are dominated by a standing wave with nodes between point 9 and point 13, which are 60 km apart from each other. Maxima and minima of the standing wave follow in regular intervals of about 9 h, which is comparable with the largest period of the SSH, seen from Fig. 11a. An approximately 90 km large area separates the zone dominated by Kelvin wave-type oscillations in the basin interior and the trapped waves at the shelf break/continental slope.

The comparison of the oscillations of the pycnocline, exemplified by the T150 and the SSH, helps in understanding the wave interaction in this region. The coastal trapped wave, seen in the temperature (Fig. 11b), clearly proves that the amplitudes of oscillations on the slope tend to decrease on both sides of the shelf break. They practically vanish on the shelf. At the point 12 in Fig. 11b, which almost coincides with the open sea nodal point, oscillations change phases. Their magnitude rapidly increases seawards of this nodal point. Seawards of location 21, the amplitudes of internal waves start to decrease again.

The instability in point 14 resulted after the day 22 in a detachment of a small ‘warm ring’. Afterwards, the amplitude of the oscillations decreases. This illustrates the possible impact of the increasing nonlinearities for the formation of mixing in the model.

It is important to note here that this area between points 13 and 20 (about 100 km wide) is dominated by seaward propagation of internal waves. Their speed, measured by the slope of the thick line in Fig. 11, is 160 km per day (1.5 m/s). These waves (shedded from the coast) have time and space characteristics which are comparable to what has been earlier documented by BI. Our model results (see also further in the text) support the mechanism of wave shedding in the Black Sea coastal zone as an important mechanism for inducing baroclinic oscillations in the interior sea.

The evolution of the amplitudes of oscillations in the western Black Sea along the section line shown in Fig. 1. (a and c) SSH, (b and d) T150. (a and b) Correspond to winter, (c and d) to summer. Abscissa axis gives the number of the model grid points on the section line, ordinate axis gives amplitude of oscillations. Lines marked by a, b, c correspond to oscillations with periods 10.5 h, 8 h and 4.5 h. In the summer plot, 1/12.5 cph oscillations are chosen as the lowest frequency ones (the full line).
Fig. 13 (continued).
band. The first indication for oscillation is at 1/4.3 cph. Obviously, this small peak indicates negligible response of the pycnocline to sea surface oscillations. This proves that no strong direct coupling exists between the barotropic Kelvin wave and the internal waves. An important increase of the spectral energy of T150 is first observed at 1/8 cph, but the most energetic oscillations are at 1/10.5 cph.

With the following analysis, we will illustrate how stable the above results are during different periods of the year. We perturb the model in the middle of the year at this time the stratification in the seasonal thermocline is very stable, and produce the same output as for the winter. The corresponding spectra are shown in Fig. 12c and d. Several important differences are to be noted: (i) The oscillations of SSH in the high frequency range are strongly suppressed in summer. Only the oscillations with periods of about 1.2 h are well presented in this frequency band. (ii) The first important maximum in the spectra for the SSH occurs at 1/4.3 cph, but unlike the winter case, these are not the most energetic oscillations. The response of T150 to these oscillations is negligible, as in winter. (iii) The oscillations with periods of 8 h give the highest peaks in the spectra, both for the SSH and for T150. This frequency was the second important in the winter oscillations of T150, and not as important in the oscillations of SSH in this season. (iv) The next pronounced maximum in the spectra is at 1/12.5 cph for the both fields. This corresponds qualitatively to the winter case, but the oscillations are now displaced slightly towards the lower frequencies. (v) The summer oscillations of T150 show pronounced low energy part of the curve at frequencies of about 1/24 cph, which was not the case in winter. The general conclusion from the above illustration on the seasonal dependency of the wave appearances is that for frequencies lower than the one of the basin Kelvin wave, the model simulates strong correlation between SSH and pycnocline oscillations in both seasons.

As it can be expected from the complicated character of oscillations in Fig. 11a and b, the spectral energy is a function of the distance from the coast. In order to understand the interaction between barotropic and baroclinic oscillations, which show a clear space dependence, we analyse below the spectral energy on the same zonal cross-section, starting from the coast and going to about 450 km in the open sea (Fig. 13). The short dashed line gives the increase of the amplitude of barotropic Kelvin wave oscillations when approaching the coast. This curve correlates well with the basin wide distribution of the amplitudes of this wave, shown in Fig. 6. The lack of a segment with an exponential decrease of amplitude in the coastal zone can be explained by the fact that in the small Black Sea the Kelvin waves are modified by the basin scales.

At the frequency of the coastal trapped wave (1/8 cph, long-dash curve in Fig. 13a), we observe very low amplitudes along the section, except in the narrow coastal region. In the same location, baroclinic oscillations reach absolute maximum (for this section), and though confined in a very narrow coastal region give an important contribution in the spectral curve (Fig. 12b). On the opposite, though well pronounced locally, the oscillations of the coastal waves, seen in the amplitude of SSH, have a relatively small contribution in the averaged spectrum for the whole cross-section. This shows the possible mechanism of excitation of baroclinic and barotropic oscillations in this area. The main factor is the presence of a steep bottom topography, which creates coastal trapped waves. They excite strong baroclinic oscillations of two types. The first type of baroclinic oscillations is trapped in the coastal region. These excite lower frequency oscillations, which propagate towards the open sea. These oscillations have as a response large variability in the SSH in the transition zone (about 75 km). Then several areas with minimum and maximum amplitudes follow. These areas are also seen by the changing amplitudes of oscillations in Fig. 11b. More detailed examination of these ‘strips’ of increased spectral energy will be done in Section 4.2.3 where we examine the basin wide amplitude-phase characteristics of oscillations.

Finishing this part we will illustrate the east–west variation of the amplitudes of oscillations in summer (Fig. 13c and d). In this season, the pure barotropic oscillations of SSH (the short dash line-c) are lower than the ones correlating with the oscillations of the pycnocline. Another important difference between the two sets of experiments is that the oscillations of the pycnocline in the coastal zone increase significantly, compared with the winter case. At the lowest
most energetic maximum (which in the summer data is shifted towards the lower frequencies) we see again sequences of zones with increased amplitudes of oscillations (the full line in Fig. 13d). Their extensions are about 100 km, which clearly indicates dependency on the Rossby radius of deformation. If we plot the amplitude of the lowest frequency oscillations in Fig. 13c and d not for the most energetic frequency, which is \(1/12.5\) cph, but for the frequency \(1/10.5\) cph, which dominates the spectrum in winter (this plot is not shown here), no localisation of zones with high amplitudes is observed. Obviously, these structures are coherent with the major types of oscillations in the western Black Sea.

### 4.2.3. Basin wide phase-amplitude characteristics of model internal waves

There is observational evidence (B1) that the Black Sea internal waves have periods which are equal to the periods of some energetic oscillations of SSH. We will examine in this part the basin wide spectra and the appearances of phase-amplitude properties in

![Fig. 14. Basin averaged spectrum: (a) SSH, 4-month data series, (b) T150, 4-month data series, (c) T150, 10 days data series, but with finer resolution in time. Horizontal axes gives the period of waves (hours), vertical axes gives the amplitude. The length of this axis corresponds to 0.5 cm for SSH and 0.005°C for temperature.](image)
the sub-inertial range (Fig. 14). The first pronounced basin wide peak appears for temperature at 1/11.4 cph (Fig. 14c). It has not an analogue in the basin wide spectrum of SSH oscillations. On the opposite, the small peak at 1/24 cph in the SSH oscillations is not simulated for the T150 (compare Fig. 14a and b). This incoherence is not observed at this particular frequency in the local spectra in the western Black Sea (Fig. 12).

With some better time resolution, basin wide spectral maxima at 1/11.4 and 1/15–16 cph become clear, but only in the data for T150 (Fig. 14c, see also Table 1). The small peak at about 15–16 h in Fig. 14c indicates the inertial oscillations (period of 17 h for the Black Sea). Since we have smoothly varying seasonal wind forcing, these oscillations are not effectively excited, which explains why the corresponding spectral maximum is very low.

The model data give a very convincing proof that barotropic and baroclinic oscillations have equal periods at some specific frequencies in the low frequency band. The spectral maxima of SSH and T150 coincide at 28.6 and 53.2 h. These are the main sub-inertial frequencies for both fields. There is a very good agreement between the spectra calculated from the model data and the most energetic peak at about 27–30 h in the estimates of Demirov (1994). This author also finds spectral maxima at about 50 h and at about 66 h. The first one enters the frequency range between 45 and 55 h, where our model simulates increased amplitudes of the oscillations (see Table 1). We have a smaller number of energetic frequencies in the model in comparison to some previous estimates, perhaps due to the different damping properties of the models. As we will show later, model spectra in different locations show much larger diversity of the spectral maxima. In spite of the differences with the observations, the main maxima in our estimates and the ones of Demirov (1994) correspond to the same time scales (1 day and 2–3 days).

The numerical models provide large amount of data over wide areas that makes possible to analyse time–space characteristics of the barotropic and baroclinic oscillations. Due to tremendous difficulties to carry out appropriate experiments in situ, aiming to understand the appearances of waves over the whole surface of the sea, this issue is not sufficiently addressed in the literature. Though the results of model simulations discussed below do not necessarily correspond in all details to the processes in the real sea, our conclusions could give indications of the possible behaviour of the Black Sea wave system. We examine below the phase-amplitude properties of the most energetic oscillations for SSH (Fig. 15) and for T150 (Fig. 16). As in the Section 4.1, we use fast Fourier transformations to calculate phases and amplitudes over the whole basin. The analysis is done for the main frequencies: 1/11.4, 1/28.6 and 1/53 cph.

The characteristic length scales of temperature oscillations at frequency 1/11.4 cph can be roughly derived from the patterns in Fig. 16a and b. At this basin wide oscillation (Table 1) the length scale corresponding to the predominating Rossby radius of

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**Fig. 15.** Amplitude and phase properties of the SSH oscillations with period 43 h, see the notations in Fig. 6. (a) Amplitudes, $a = 3.5 \times 10^{-4} \text{ m}$, $c_i = 0.2 \times 10^{-3} \text{ m}$. (b) Phase. To more clearly visualise the space properties, phases from 0° to 180° are plotted with dash lines, and the ones from 180° to 360° with full lines, $c_i = 20°$. 
deformation is 50 km. The oscillations in the eastern most basin and in the western basin have very small amplitudes. Most of the baroclinic energy in this frequency range is along the western basin continental slope (note the very pronounced ‘intrusions’ in the outflow area) and in the central basin. The phase pattern (Fig. 16b) is very patchy and does not give clear indications for a dominating direction of the wave propagation in the interior basin. It shows, however, very strong coherency at distances of about the radius of deformation \( R_d \). Our estimates for the length scales compare well with the results of BI.
who found in their numerical experiments long internal gravity waves (wavelength of about 60–90 km) with periods of about 7–9 h.

The phase-amplitude characteristics of the oscillations at 1/43 cph (Fig. 15a and b) illustrate that the model simulates a rotational structure with anticyclonic sign in the shelf area. Oscillations have lower amplitudes on the shelf than in the basin interior (Fig. 15a). The particular choice of the phase lines does not allow to identify the sign of rotation since this type of plotting was accepted to better visualise standing wave patterns.

The patterns in Fig. 15a and b correlate very well with the phase-amplitude characteristics of T150 at 1/28 cph (Fig. 16c and d). Comparison between the same characteristics at frequencies 1/28 and 1/53.2 cph, where the two fields show spectral maxima (see Fig. 14a and b), indicates that no pronounced correlation exists. Different explanations could be given to this discrepancy. The first plausible one is that no exact correspondence of patterns is to be expected if the areas where the waves propagate are too different, as in the case discussed above. The oscillations of the SSH reach the shelf and propagate on it with a speed which is different from the one in the deep sea. The internal waves propagate mainly in the deep part of the basin. The impact of the shelf on the characteristics of barotropic oscillations could result in substantial distortions at some particular frequencies and in decreasing the coherency in space between internal and external mode oscillations. At some neighbouring frequencies, where the energy of barotropic oscillations is substantially smaller, coherency in space is more clearly observed (Fig. 15, frequency of 1/43 cph). This can lead to the speculation that oscillations in the pycnocline have the leading role in exciting the low frequency oscillations of SSH.

The second explanation of the absence of a clear coherency between barotropic and baroclinic oscillations over most of the sea takes into account the different origin of their excitation. The variance map (Fig. 8a) indicates that small scale features dominate the SSH pattern along the western coast. As we demonstrated in Section 4.2.2, strong coherency between the two types of oscillations is simulated there. In these areas, the exchange between the barotropic and baroclinic modes becomes important, contributing to the coherent appearance of phase-amplitude characteristics.

An important change in the wave characteristics of T150 is observed with decreasing frequency from 1/11.4 cph to 1/28.6 cph. The wave pattern in the western basin dominates the solution at this frequency. This is illustrated by the six maxima in the amplitude and three zones with equal phases (Fig. 16c and d). The corresponding wave length is about 80–90 km. For the wave-numbers in the same range (wavelength 40–100 km) BI found periods of about 1.5–3 days, and phase speed of about 20 to 50 cm s⁻¹. As can be easily calculated from Fig. 16c, our estimates do not differ much from the earlier estimates.

The line with equal phases in the western-most area tends to follow the coast, and to form a strip, where oscillations are coherent. Amplitudes at the shelf/continental slope edge have maximum values, particularly in the location south of Cape Kaliakra and to the west of Crimea Peninsula. Amplitudes in the area of the main wave have minimum along the Turkish coast. However, this minimum is still much larger than the magnitudes of oscillations in the central and eastern basin. Obviously, oscillations at this frequency are not important for these areas.

In the narrow section of the sea, between the Turkish coast and the Crimean Peninsula, phase lines tend to change direction from NW–SE in the western basin to N–S in the central basin. Oscillations along the Anatolian and Caucasus coasts are in phase for very long distances (large length scales). In the eastern-most Black Sea, the oscillations tend to form again (as in the western Black Sea) phase patterns with scales of about 100 km.

These results find an excellent support by the analysis of BI on the 12 h barotropic wave. As seen from Fig. 16c and d, wave vectors in the continental slope area are normal to the isobaths. This also compares well with the observations of BI, who found that the wave vector has a south eastward direction. This could give arguments to speculate that wave shedding from the coastal areas, induced by oscillations of the halocline, can contribute to the generation of internal waves in the basin interior. This seems a plausible mechanism, taking into consideration the results of the local analyses, given in the preceding sections.
The excitation of oscillations with 12 h periodicity is explained in the work of BI by the tidal forcing of the Black Sea. However, without having this forcing in the model, we succeeded to simulate a similar south easterly propagating wave at approximately two times lower frequency.

Substantial differences can be found in the phase-amplitude properties of T150 oscillations at the two most energetic frequencies (compare Fig. 16e and f, frequency 1/53.2 cph and Fig. 16c and d, frequency 1/28.6 cph). The oscillations with 53 h periods indicate a concentration of wave energy in the western most part of basin. Length scales are comparable with the ones of the 28.6 h oscillations (compare Fig. 16d and f). As in the case of the 11.4 h oscillation, no pronounced basin scale features can be found in the wave phase patterns, except for the stronger coherence between the oscillations along the Anatolian and Caucasian coasts. The simulation of very large amplitudes of the oscillations with 53 h period in front of the Bulgarian coast (see also Fig. 8b and Fig. 13) is coherent with the estimates of Demirov (1994).

4.2.4. Some illustrations on the spatial diversity of SSH and temperature spectra

The spectral composition of oscillations is different for the different Black Sea areas. Therefore, comparisons of basin wide averaged spectra (as the ones discussed in Section 4.1 with local spectra derived from experimental data could be sometimes controversial. Analysing model spectra in separate locations can, however, contribute to a better understanding of wave appearances in the sea. With the analysis of the basin-wide amplitudes done in Section 4.2.3, we partially elucidated this issue. Now, we will focus on some locations, known from observations as areas with extreme variance of the pycnocline.

Fig. 17. Model spectra for the SSH in the points off Cape Kaliakra in the winter perturbation experiment. Locations of the points where spectra are plotted are 30 km apart one from another (two model grid intervals). The spectrum in the upper plot is in the point (i, j) = (9, 20), see Fig. 1. Each following plot is eastward from the preceding one. Horizontal axes give the period of waves (hours), vertical axes give the amplitude (cm). The length of these axes corresponds to 2 cm.
The area of Cape Kaliakra is characterised by very strong upwelling events (Trukhchev et al., 1985). This is often explained with the increasing contribution of nonlinearity in this area, which is due to specific topography, coastline aspects and the attacking angle of the current (e.g., Demirov, 1994). We give below an example on the space dependence of model spectra along the zonal section at 43°27’N (Fig. 17).

The 53.2 h oscillation dominates the solution on the shelf. Its relative weight in the energy spectrum increases with increasing distance from the coast, and reaches maximum at 29°15’E (just to the left of the continental slope, as seen from north to south). About 60 km to the east of this location the oscillation at 1/28.6 cph frequency start to dominate the solution. This is the standing wave in the western Black Sea (Fig. 16c and d). This result proves again that the transformation of the type of oscillations occurs at the continental slope/shelf edge. The intense oscillations in the basin interior, illustrated in Section 4.2.3 by the well-pronounced wave patterns at 1/28.6 cph, decrease in intensity. The energy thus tends to concentrate over the continental slope/shelf edge at lower frequency. As shown previously by the analyses of barotropic and baroclinic oscillations, this is accompanied by large amplitudes of the pycnocline oscillation. Breaking internal waves in this location might be the reason for internal mixing and explain observed formation of periodic strong upwelling phenomena, excited by external (wind) forcing.

A ‘drift of the spectral maxima’ is found by BI for different events in the frequency range between 6.5 to 46.5 h, caused by extreme amplitudes of the pycnocline. These authors speculate that local processes are perhaps influenced by waves, originating from remote locations, and carrying their specific characteristics. Changing spectral composition and the contribution of different frequencies in the spectrum could also be due to the amplification of some frequencies excited by the atmospheric forcing. The simplified forcing in our model and the predominating variability due to the internal dynamics, does not allow to search for similarities supporting all experimental results. Nevertheless, model oscillations in some locations are characterised by peculiarities (Fig. 18), which deserve to be commented, though they have very small amplitude. We choose as an illustration the temperature fields since SSH does not show spectral peaks in the low frequency range (no similarity with curves in Fig. 18 are observed). Along with the main oscillations of temperature, we find oscillations with periods of about 29 h, 3.5–4 days, 11 days (point E) and 11.7 days and 14.7 days (point F). These longer period oscillations are typical for the synoptic driving (BI). They can be easily amplified in the model if the corresponding external forcing is included. This will be analysed in a future study.

4.3. The propagation of oscillations along the coast. Model evidence for coastal trapped waves

The propagation of oscillations along the coast is illustrated below at frequencies 1/72 and 1/51 cpd
The fast Fourier transformation (FFT) analysis is based on a model output for the SSH for a 1 year integration. With accumulating statistics for a rather long period, we aim to present some typical characteristics of the oscillations in the low frequency range. We see from Fig. 19a, that maximum amplitudes are trapped against the coast. The Caucasian coast acts as a wave guide, particularly at $1/72$ cpd. The same conclusion could be made for the western coast as well, but the area of maximum amplitudes at $1/72$ cpd is much more diffused towards the open sea. There are very clear indications, particularly in the southern and eastern Black Sea, for oscillations propagating with the coast on their right at frequencies of $1/72$ cpd. The tongue-like phase patterns along the Caucasus coast exemplifies this, and give an idea about the direction in which the waves rotate.

In a large region west of Crimea Peninsula (extending from the northwestern shelf to the basin interior) the anticyclonic rotation starts to dominate the solution. These changes in the type of rotation could be due to the abrupt change of the width of the slope west of the Crimea Peninsula, which results in changing waves properties, or in exciting oscillations with different frequencies.

To have an insight in the variety of the oscillations, we show in Fig. 19c and d phase-amplitude characteristics at $1/51$ cpd. This frequency is very close to the frequency where spectral maxima are observed both for SSH and temperature (Fig. 14). Again, we see the pronounced cyclonic rotation in the eastern most Black Sea. However, rotational structures are less clearly observed in the area south of the Crimea Peninsula than at a frequency of $1/72$ cpd. The anticyclonic rotation at $1/72$ cpd in the

![Fig. 19. Amplitude and phase properties of the SSH. Amplitudes are plotted as in Fig. 15, phases are plotted as in Fig. 6. (a) Amplitudes at $1/72$ cpd, $ma = 1.4 \cdot 10^{-3}$ m, $ci = 0.2 \cdot 10^{-3}$ m, (b) phases at $1/72$ cpd, (c) amplitudes at $1/51$ cpd, $ma = 1.3 \cdot 10^{-3}$ m, $ci = 0.2 \cdot 10^{-3}$ m, (d) phases at $1/51$ cpd.](image-url)
shelf area is replaced by cyclonic one at 1/51 cpd, so simultaneous existence of opposite rotations at different frequencies is possible.

Oscillations in the Bosphorus inflow area (north of Sakaria canyon), with the coast on their right are noteworthy at 1/51 cpd. One important difference between the coastal trapped waves and the oscillations of the SSH with higher frequencies is that the amplitudes of the first ones are at least an order of magnitude higher (compare the magnitudes in amplitudes plots in Fig. 19 with Figs. 15 and 6). This is, however, mostly due to the longer integration time, with a model forced with an annual signal. This makes possible the amplification of residual effects over the year. This issue is the central topic of Section 5.

5. Forced oscillations

5.1. Seasonal variability of the sea surface elevation as a result of variability in the river run-off

To give an idea on both time and space variations of SSH response to the variability in the river run-off, we show SSH anomaly along zonal section at 43°13′ N (Fig. 20). The anomaly is calculated by subtracting at each time and in all points the value of SSH at the beginning of the year. This type of presentation of the results is chosen to enhance the sign of variations of SSH. In spring, anomaly tends to increase in all locations (positive values), which is a result of filling the basin with larger amount of fresh water. The SSH anomaly shows maxima in

![Fig. 20. Temporal variability of SSH at 43°13′ N. The numbers on the horizontal axis (J) are model grid points, and the ones on the ordinate (K), the time step. To convert the axes in length and time, units are to be multiplied by Δx = 15 km (see also Fig. 1) and Δt = 4 days, which is the time discretisation of this plot.](image-url)
June (about 1 month after the maximum of the river discharge is reached) over the whole surface of the Black Sea. The time when the SSH reaches its minimum values varies in a longer range, from September–October in the eastern-most Black Sea to December in the central part. Obviously, local processes strongly affect the temporal evolution of SSH, as seen by the different types of propagating signals in different parts of the sea. This requires a more detailed study, which is the one of the issues addressed in the paper by Stanov and Beckers (1998). However, in the present paper, which is focused on process studies, we want to address the impact which different types of oscillations might have on the Black Sea circulation/stratification.

5.2. Sensitivity analysis on the impact of friction in the baroclinic part on the SSH variability with annual periodicity

In most of the remainder of the paper, we will consider the effects resulting from smoothing the barotropic part or from using a different model topography. We will illustrate different model responses by analysing simulated data in points G, H, and F. Point G is in the front of the Strait of Bosphorus, point H is on the continental slope, south of the shelf area, and point F is in the eastern Black Sea. SSH in these locations, simulated in experiment C (Fig. 21a) clearly shows the important contribution of the seasonal signal to the SSH variability. Some specific features in the appearances of this variability are observed in the different locations. This practically supports the results shown in Fig. 20. However, the variability shown in Fig. 21a represents more clearly some details related to the phase differences.

The data shown in Fig. 21 are not filtered, therefore, the high frequency oscillations and their amplitudes could be identified by the thickness of the curves. We have chosen the friction on the barotropic part in experiment C such that the oscillations with amplitudes from a couple of millimeters to about 1 cm are not too much affected by this friction because this is the amplitude of oscillations which is practically observed in the Black Sea in calm weather.

Fig. 21. (a) SSH oscillations in C-experiment. The locations of the points G, H, F are shown in Fig. 1. Comparisons between SSH, simulated in C-experiment and in SB experiment in points: H—(b) and G—(c).
Fig. 21 (continued).
The amplitude of oscillations of SSH with annual periodicity in experiment SB (smoothed bottom) is almost the same as in the C-experiment (Fig. 21b and c). However, we see from Fig. 21b (location H, on the continental slope) that the whole curve in the experiment SB is 1–3 cm lower. Unlike this location, we see almost the same model behaviour of the SSH in location G (in the Bosphorus outflow area, Fig. 21c) during the first half of the year. In fall, SSH in the experiment SB is about 1 cm lower.

The difference in the time evolution of SSE in the two experiments shows that smoothing the topography could create come inconsistencies in the simulated seasonal variability. The differences between the two simulations increase in the transition area between the shelf and the open sea (location H). This is another indication that smoothing bottom relief (in order to avoid errors arising from the application of \( \sigma \)-co-ordinate system to regions with large continental slopes), could affect not only the appearance of high frequency motion and residuals from them in a short run (this is well-known from earlier studies, see Haidvogel and Brink, 1986), but also the low frequency variability simulated in the ocean models.

5.3. The impact of high frequency oscillations on the seasonal variability of SSH

We will now investigate the residual effects, which high frequency oscillations might have on the annual variability of SSH, using the results of experiment C and experiment FFS. In the presence of stronger explicit friction in the barotropic part (experiment C), the oscillations of the SSH decrease in amplitude. This might reduce the interaction of barotropic part and the bottom. Studies on the residual effects from transient processes has been previously done by Haidvogel and Brink (1986). Similar effects in the Black Sea model could result from the variability in the river discharge. This issue has been examined in more detail by Stanev and Beckers (1998). Here, we remind only that the river discharge, which is specified in the model as time dependent, shows a clear seasonality with maximum in April (Fig. 22a). The simulated barotropic transport in the Bosphorus Strait has an opposite phase, which is seen from the lower curves in the same figure. With reducing the friction on the barotropic part in the FFS experiment, the forcing in the strait of Bosphorus, which is dependent on the actual SSH simulated in the model, slightly changes (compare the difference between the full and dash line in Fig. 22a).

The difference between simulations in the two experiments are better pronounced in the SSH (Fig. 22b–d). Simulations in point G (Bosphorus inflow area) almost coincide during the warm part of the year. However, C-experiment gives deeper positions of SSH in the beginning of winter, whereas FFS-experiment gives deeper SSH in the late fall. These differences between SSH in experiments C and FFS cause slight differences in the barotropic forcing in the Bosphorus location during winter. This is the reason for the deviations between dashed and full lines in Fig. 22a.

Summer maxima of the SSH coincide in C and in FFS experiments in location H (on the shelf), however, winter values in FFS experiment are about 2 cm lower (Fig. 22c). On the opposite, winter values in FFS experiment are about 1 cm higher in location F, which is in the eastern Black Sea (Fig. 22d). This shows that the effects resulting from the friction, which we impose on the barotropic part, have a clear regional dependence. The rule is that the differences between the results in the two experiments are larger in winter.

It follows from the above analysis that reducing the amplitudes of short periodic SSH oscillations (or allowing them to increase unrealistically) could negatively influence simulated variability with much lower periods. As known from earlier studies (see Haidvogel and Brink (1986)), and as demonstrated by the differences in the two simulations, model data depend on the interaction between the variable flows (having different amplitudes in the two experiments) and the topography. During winter, when the vertical stratification is more homogeneous, (barotropic theory could be applied during this time, at least for the shallow shelf), the bottom relief affects strongly the barotropic part of solution, and some important residual effects could appear. In the warm part of the year, the seasonal thermocline has almost the same depth throughout the sea and shields the bottom (including the bottom of the shelf) from direct oscillations from above. This might explain why the estimates from the two experiments are closer during the warm part of the year, and motivates to study the
impact of stratification on the temporal variability with seasonal and interannual time scales (Stanev and Beckers, 1998). Unlike this paper, we focus now on the impact of short periodic oscillations on the thermohaline structure of the model.

5.4. The impact of short periodic oscillations on the formation/erosion of halocline

The problem on how important the barotropic oscillations could be for the circulation of the baroclinic ocean is still disputable. We demonstrated that these oscillations could perturb the pycnocline. These perturbations have different appearances on the shelf, over the continental slope, or in the deep basin (Fig. 17). Substantial differences in the amplitudes are simulated in the western and eastern Black Sea. Extremely strong oscillations in front of Cape Kaliakra (Fig. 17) are coherent with the existing understanding about the local extrema in the oscillations of the pycnocline. In a nonlinear system, the increase

Fig. 22. (a) The river discharge (upper curve) and the time filtered outflow through the Strait of Bosphorus (lower curves). (b, c, d) SSH in three locations (b) location G, (c) location H, (d) location F. Full lines, C-experiment; dash lines, FFS-experiment.
of amplitudes can result in breaking waves and generation of turbulence (see Fig. 11b). In such a way, the impact of barotropic oscillations on the ocean stratification would become significant.

We remind here that the Black Sea stratification is extremely stable, which reduces severely the vertical mixing. The small mixing explains the persistency of the CIL. Breaking waves could contribute to entrainment of surface water in the CIL, which will finally affect the water masses. Since the CIL has a rather small thermic inertia, it could become affected by the mixing due to the oscillations of pycnocline. Thus, oscillations of the free surface, enhancing the oscillations of pycnocline, could present an efficient mechanism for generating internal mixing. For a sea with a very weak vertical mixing this can have important consequences.

We now discuss the results of the following two model runs: the C-experiment, where we use standard friction on the barotropic part, and the FFS-experiment, with much less friction. Both experiments are initialised with the results of the C-experiment. The evolution of the vertical stratification throughout the year is illustrated for point H, over the continental shelf break, and point F, in the eastern Black Sea (Fig. 23). With this choice, we aim to illustrate the variability in a location, adjacent to the area of the cold intermediate water formation, and in another one, which is sufficiently far from this area.

The internal gravity waves start to undulate the isotherms in the warm part of the year and to disappear, when the cooling starts again (Fig. 23). This is very well-pronounced in the shallow location, where the seasonal thermocline practically vanishes in the cold part of the year. In summer, the thermocline in this location is thicker and deeper than in the eastern Black Sea. This is due to much stronger mixing, simulated on the shelf, which is easily seen from the model’s turbulent kinetic energy fields.

We see in both locations the winter ventilation from December to early March, (vertical isolines) reaching depths of about 50–60 m in location F, and...

![Fig. 23. Time evolution of the temperature profile in point H (a and b) and in point F (c and d) simulated in C-experiment (a and c) and FFS-experiment (b and d).](image-url)
the bottom in location H. The isotherms marking the position of the CIL (8°C water) reach sea surface in location F for a very short time in February. The ventilation in location H is well-pronounced from day 30 to day 70. This gives rise to the formation of the CIL, one of the major thermohaline peculiarities of the Black Sea stratification. The warming in spring and in summer results in the formation of the
seasonal thermocline, reducing the vertical diffusion and shielding the CIL waters from mixing with the surface waters.

The CIL persists throughout the year in both experiments. It is noteworthy that 7.75°C isotherms cannot be found in Fig. 23 from day 110 to the day 280. The re-appearance of this isotherm after the second date, which is practically before the next cooling starts, means that cold waters are transported advectively to this location from the area of their origin. This is coherent with the existing concepts of the propagation of the cold water in this sea.

The surface gravity waves have different amplitudes in the two experiments, which is due to different friction in the barotropic part. The oscillations, which they excite in the halocline, have different magnitudes in the different parts of the sea. We proved this in Section 4.2, but we also see it clearly now in Fig. 23, where the amplitudes of internal waves have much lower magnitudes in the eastern than in the western Black Sea. These two factors (regional dependency of magnitudes of oscillations and their dependency on the model friction) will result in quite diverse appearances of the impact of barotropic solution on the formation of water masses.

As seen from the differences in model data in experiments C and FFS, the seasonal thermocline is thinner and shallower when the oscillations of SSH are suppressed (Fig. 23a and c). Minimum temperature of the cold intermediate water (CIW) is 6°C in C-experiment and 7°C in FFS experiment. The axis of CIL deepens slightly in C-experiment throughout the year. This deepening is better pronounced in FFS experiment (reaching about 80 m in August in location F). The increase in the cold water content in location F during the late summer and early fall in experiment FFS, is not simulated in experiment C (Fig. 23c and d).

One could expect that such pronounced changes in the thermal structure of the sea would have pronounced impact on the general circulation, but this has not been observed. Indeed, the CIL is dynamically a rather passive layer (Stanev, 1990) and could be regarded as a good ‘tracer’ for the mixing processes at intermediate depths. From the above analysis of CIL, it became obvious that friction on the barotropic part affects vertical stratification. This is a consequence of increasing magnitudes of the enhanced oscillations in the pycnocline, which in a nonlinear system tend to create additional mixing.

6. Interdisciplinary implications

The modification of the mixing behaviour and the presence of particular energetic oscillations is not only interesting from a physical point of view, but also may have implications for the biological system. A straightforward effect can be illustrated on the example of a particle which is forced to oscillate in a vertical plane by an internal wave. The up and downward motions will eventually bring it nearer to the surface into light conditions which are favorable to primary production. So, even if all other parameters of the problem remain the same, but all state variables (e.g., temperature, salinity, phytoplankton concentration) are moved with the water parcel, the light conditions may change, favoring phytoplankton growth.

The illumination at sea surface $I_0(t)$ depends on the moment of the day and may be schematized in spring by

$$I_0 = \max\left[0, \sin\left(2\pi t/D\right)\right]$$

where $t$ starts at the rising sun and $D$ is the length of the day. For the depth $d$ of the water parcel, the available light decreases to:

$$I = I_0(t)\exp(-kd)$$

where $k$ is the decay coefficient.

In the presence of internal waves with amplitude $A$ and period $T$, the water parcel is subjected to an intensity of light

$$I = I_0(t)\exp\left[-k\left(d + A\sin(2\pi t/T + \phi)\right)\right].$$

We can compare the above two intensities using parameters typical for the Black Sea (depth of the water parcel of 50 m, a wave amplitude of between 10 and 20 m, and a period of 10.5 h). In the case of a spring situation and the parameters of ecological system of Gregoire et al. (1998), we get the situation of Fig. 24. There we see that the light intensity felt by a moving parcel can give a net rise to growth rates of up to $0.4/d$, whereas the non-moving parcel at most reaches $0.1/d$. As the parcel remains for several hours in this high growth rate region, we may
Fig. 24. Time evolution light intensity for a fixed water parcel at 50 m and a water parcel moved by a wave. In addition, the light intensity levels for growth rates of $0.1/d$, $0.2/d$ and $0.4/d$ are shown.

expect a strong response to the presence of an internal wave. One must, however, keep in mind that the biological system, which is subjected to this coherent movements, may react by moving itself relatively to the water parcel, but since not all state variables can move, the relative movement may also change the system behaviour.

We see that a significant change in the growth rates may be possible, caused by moving parcels into different light conditions. The residual effects will not cancel, since the evolution in each parcel is strongly nonlinear.

A more conventional effect on biological systems is also the modification of the mixed layer depth and the mixing of CIL waters. As we have shown, the presence or absence of waves makes the model change its representation of the mixed layer and in consequence, the average nutrient concentration will be affected.

Since the wave effects may be important for biological models, it would be interesting to test the two effects (movement of water parcels by internal waves and modification of the mixed layer) in a 3D model Gregoire et al. (1998), by performing similar experiments as those described in the present paper. However, it will be difficult to differentiate the two effects in such an experiment, except if vertical diffusion is artificially kept unmodified to analyse the pure effect of the parcel displacements by waves.

Locally, these effects are known to be important (e.g., New and Pingree (1990)), but if it turns out that the effects are neither negligible at basin scale, then it is also clear that daily atmospheric data definitely should be applied, not only because they modify the mixed layer structure but also because atmospheric perturbations trigger the waves shown here.

7. Conclusions

In a complicated physical system like the ocean, the appearances of physical processes are practically
a consequence of various interactions between oscillations with different time and length scales. The fundamental issue how important are the short-time scale processes, or the ones with small wavelengths for the ocean circulation is far from being solved. Numerical models give a very powerful tool to formulate different scenarios to, at least partially, elucidate the possible impact of various oscillations on the ocean circulation/stratification.

We now face an increased interest in developing models with detailed description of thermodynamic processes and the oscillations of the SSH, which can contribute to further understanding the barotropic–baroclinic oscillations in the ocean and their interpretations in realistic ocean conditions. The present work presents a small contribution in this field. We decided here to illustrate the appearances of processes as simulated by the model, rather than to look deeply into the physics of the different processes. The last is more appropriate for the idealised studies, where the results are not ‘contaminated’ by a large number of processes and interactions between them. Our strategy was motivated by our intention to address the above issue for one real oceanic basin, which provides wide enough spectrum of processes to be simulated by single simplified model. Further work is foreseen with including more realistic forcing, or carrying out experiments which aim to investigate physical processes.

The present results could be regarded as preliminary, but even in such quality they give some motivation to claim that (at least for the particular area, the Black Sea, which is a typical coastal basin) the interaction between barotropic and baroclinic oscillations could have pronounced impact on the circulation/stratification. We gave an illustration on how short periodic oscillations affect long term processes of erosion of halocline, which contributes to the specific water mass formation. Nonlinear processes, and breaking waves, particularly along the western cost and shelf, are possible candidates to explain how the mixing in the Black Sea works. Another issue which still remains not enough elucidated is the upwelling. It is associated in some areas with extreme oscillations of the halocline and penetration of CIW into the mixed layer. This could provide the physical background for the biological productivity, which recently attracted much attention in the Black Sea region. Since upwelling is intimately related to the waves in the pycnocline, models like the one used here should address further this key issue for the Black Sea circulation.

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References


