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A Fine Resolution Hybrid Coordinate Ocean Model (HYCOM) for the Black Sea with a New Solar Radiation Penetration Scheme

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A FINE RESOLUTION HYBRID COORDINATE OCEAN MODEL (HYCOM) FOR THE BLACK SEA WITH A NEW SOLAR RADIATION PENETRATION SCHEME

By

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To Dr. Harley E. Hurlburt of the Naval Research Laboratory:

“Sometimes swaying, sometimes crashing – thanks for the waves you have given me.”
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ABSTRACT

A 1/25° × 1/25° cos(lat), (long × lat) resolution eddy–resolving HYbrid Coordinate Ocean Model (HYCOM) is introduced for the Black Sea and used to examine the effects of ocean turbidity on upper ocean circulation features including sea surface height (SSH) and mixed layer depth (MLD) on annual mean and seasonal climatological time scales. HYCOM is a primitive equation model with a K–Profile Parameterization (KPP) mixed layer sub–model, and it uses a hybrid vertical coordinate that combines the advantages of isopycnal, terrain–following (sigma) and z–level coordinates in optimally simulating coastal and open–ocean circulation features. This model approach with ≈ 3.0 km resolution is applied to the Black Sea for the first time in the literature.

A newly–developed time–varying solar penetration scheme that treats attenuation as a continuous quantity is added to the model as will be described in this dissertation. This new scheme involves two bands for the top 10 m of the water column; thus, it is suitable for any Ocean General Circulation Model (OGCM) that has fine vertical layer resolution near the surface. With the new parameterization, the depth–dependent attenuation of subsurface heating in HYCOM is given by monthly mean fields for the attenuation of Photosynthetically Available Radiation ($k_{PAR}$) during 1997–2001 as constructed from Sea–viewing Wide Field–of–view Sensor (SeaWiFS) satellite data. Climatologically–forced HYCOM simulations with no assimilation of any ocean data are then used to show the importance of including spatial and temporal varying attenuation depths for the annual and monthly mean predictions of upper ocean quantities in the Black Sea.

Results are reported from three model experiments using different sets of $k_{PAR}$ values: (1) the monthly mean SeaWiFS $k_{PAR}$ data set to account for the seasonal and spatial variation of subsurface heating (standard simulation), (2) an unrealistically large value of $k_{PAR} = 99$ m$^{-1}$ to ensure the complete absorption of solar radiation at the sea surface, and (3) a basin–wide constant value of $k_{PAR} = 0.06$ m$^{-1}$ that is representative of clear water all over the Black Sea. Climatological wind stress and thermal forcing (air temperature at 10 m above the sea surface, air mixing ratio at 10 m above the sea surface, shortwave radiation at the sea surface, longwave
radiation at the sea surface) used in the model simulations were constructed from the 1.125° × 1.125° European Centre for Medium-Range Weather Forecasts (ECMWF) re-analysis product during 1979–1993, and 1.0° × 1.0° Fleet Numerical Meteorology and Oceanography Center (FNMOCS) Navy Operational Global Atmospheric Prediction System (NOGAPS) data during 1998–2002. The basic methodology is to force the model with monthly climatological atmospheric fields (i.e., wind and thermal forcing), but with the addition of representative 6-hourly wind stress anomalies, using the model SST and an accurate bulk formula to calculate latent and sensible heat fluxes.

It is first shown that most of the large differences between the basin-wide mean net heat flux at the sea surface and flux entering the mixed layer (mixed layer flux) in the Black Sea occur from April to October on the climatological time scales, and there is almost no difference between the two during the remaining months. The difference between the net heat flux at the sea surface and mixed layer flux is shortwave radiation absorbed below the mixed layer, which can be as large as 50 W m⁻² in summer because the MLD is very shallow (< 4 m) over the most of Black Sea. In particular, the model simulations reveal that the basin-wide mean flux below the mixed layer can be as large as 100 W m⁻² regardless of the atmospheric forcing used. Model results also show that using a constant attenuation depth of ≈ 17 m (clear water assumption), as opposed to using spatial and temporal varying attenuation depths, changes the surface circulation, especially in the eastern Black Sea. An unrealistic subsurface warming in the former results in a weakening of the mixed layer stratification and a deep MLD. As a result, the deep MLD off Sinop (at around 35.5°E, 42.5°N) weakens the surface currents regardless of the atmospheric forcing used in the model simulations.

Finally, the results in this dissertation suggest that, if the Black Sea turbidity is entirely or largely due to biology, a lack of nutrients (or other cues for loss of biomass) will have a significant effect on the overall circulation of the Black Sea. In this dissertation, it is not specifically examined how likely this is (i.e., how robust the bio-system is in the Black Sea).
CHAPTER 1

INTRODUCTION

The Black Sea is an almost enclosed sea connected to the Sea of Marmara and the Sea of Azov by the shallow and narrow Bosporus and Kerch Straits, respectively (Figure 1.1). The importance of the Black Sea extends far beyond its role as an enclosed sea because it constitutes a unique marine environment with a nearly land-locked shape, and minimal ventilation of the deep waters. In addition, the region presents challenging test areas to study oceanographic phenomena which are dominated by coastline orientation, coastline shape and topography, so it is an attractive test domain for ocean modeling studies. Thus, the Black Sea serves as a medium scale laboratory for investigation of a series of oceanographic phenomena that are common to different areas of the world ocean.

The most important current feature in the Black Sea is a cyclonic Rim Current located along upper continental slope (e.g., Sur et al. 1996; Oguz and Besiktepe 1999; Afanasyev et al. 2002). The Rim Current usually follows the topography of the shelf break, and it is accompanied by a series of anticyclonic mesoscale eddies as well as transient waves with an embedded train of mesoscale eddies propagating cyclonically around the basin. The meandering Rim Current implies important dynamical processes with regard to the cross-shelf exchanges. Sub-basin scale gyres in the Black Sea are usually connected by intense jets, a meandering Rim Current and a propagating eddy field. Previous observational studies (e.g., Oguz and Besiktepe 1999; Afanasyev et al. 2002) and fine resolution eddy-resolving Ocean General Circulation Model (OGCM) studies (e.g., Oguz et al. 1995; Staney and Beckers 1999; Staney and Staneva 2000; Kourafalou and Staney 2001; Staneva et al. 2001) examined upper ocean quantities, mainly the Rim Current, using different approaches. None of these OGCM studies directly indicated what kind of turbidity they used in the mixed layer or how they treated the solar radiation attenuation in the model simulations. Thus, the missing part of the Black Sea OGCM studies is an examination of upper ocean features with respect to solar radiation penetrating into and below the mixed layer. Such an examination is necessary for the Black Sea because the sharp density stratification near the surface inhibits the ventilation of
Figure 1.1. The geography of the Black Sea. Also shown are rivers discharged into the Black Sea. Only major rivers named on the map are used in the model simulations. The narrowest and widest points in the Sea of Marmara (SM) are 1.2 km and 6.4 km, respectively, and depth varies from 50 to 91 m in the main channel. Bosporus is approximately 31 km long, with an average width of 1.5 km. It is only 0.7 km at its narrowest point with a sill depth of 30 m. The length of the Dardanelles is about 70 km with a general width ranging from 1.2 km to 2 km. Two coastal cities (Trabzon and Sinop) used in the text are also shown.

sub–pycnocline waters of the Black Sea. As a result, the euphotic structure is usually confined to the mixed layer, which is directly affected by the air–sea interaction processes. Since the region is biologically active (e.g., Kideys et al. 2000; Konovalov et al. 2001; Oguz et al. 2002), it is necessary for an OGCM to take water turbidity into account, even if it is not coupled with a biological model. For example, variations in the attenuation of light–absorbing pigment may change the vertical distribution of heating in the mixed layer, thereby affecting upper ocean dynamics.

With the availability of a satellite–sensed diffuse attenuation coefficient at 490 nm ($k_{490}$) data set (McClain et al. 1998), it is now possible to determine the ocean turbidity globally at high spatial resolution and use it as part of the heat flux forcing in an OGCM. By using such a data set, the time–varying solar penetration schemes (e.g., Morel and Antonie 1994; Murtugudde et al. 2002; Morel and Maritorena 2001) can treat attenuation as a continuous quantity, which
is an improvement over the use of a few discrete attenuation values corresponding to classical Jerlov water types (Jerlov 1976). However, matching attenuation of Photosynthetically Available Radiation ($k_{\text{PAR}}$) to Jerlov water attenuation profiles is needed because the original single decay formulation (e.g., $Q_{\text{sol}}(z)/Q_{\text{sol}}(0) = 0.49 \exp(-z k_{\text{PAR}})$) as used in OGCMs (e.g., Murtugudde et al. 2002; Kara et al. 2003a) is only accurate for depths deeper than 10 or 20 m.

Unlike ocean models, with a bulk mixed layer, which have vertical resolution of 10 m or coarser near the surface (e.g., Schneider and Zhu 1998; Wallcraft et al. 2003), OGCMs which have many vertical levels near the surface (e.g., Bleck et al. 2002) often have MLDs much shallower than 10 m and also need to distribute solar radiation within the mixed layer. Thus, a modified version of the single decay formulation is required. For this purpose, a new solar subsurface heating formulation is developed in this dissertation. The new parameterization treats attenuation as a continuous quantity (depending only on depth and $k_{\text{PAR}}$), but it is more accurate (i.e., closer to Jerlov curves from the surface downward for representative $k_{\text{PAR}}$ values) in the 0–20 m range.

The main focus of this study is to set up an ocean model with hybrid coordinates for the Black Sea, using the newly-developed solar radiation penetration scheme. Effects of subsurface heating are examined on circulation dynamics of the region, including its coastal and shelf regions and the interactions of these regions with the interior of the basin and air–sea interactions. The model is applicable to studies concerned with coastal and interior locations of the Black Sea because the hybrid coordinate is isopycnal in the open, stratified ocean, but makes a dynamically smooth transition to a terrain–following coordinate in shallow coastal regions, and to $z$–level coordinates in the mixed layer and/or unstratified seas.

This dissertation is organized as follows: In chapter 2, spatial and temporal characteristics of attenuation depths from the remotely-sensed data in the Black Sea are introduced, as well as a new formulation representing attenuation of solar radiation with depth. In chapter 3, basic features of HYCOM are given, along with set up of the Black Sea model and the atmospheric forcing used in the model. In chapter 4, the effects of ocean turbidity on the model simulations are discussed. Finally, the summary and conclusions are presented.
CHAPTER 2

SOLAR RADIATION PENETRATION

Ocean solar radiation has a strong spectral dependence with the red and near-infrared radiation absorbed near the sea surface and shorter wavelengths absorbed at greater depths (e.g., Lalli and Parsons 1997). In general, the solar irradiance at the ocean surface ranges in wavelength from about 300 to 2800 nm, and is composed of three general regions: the ultra–violet (UV) below 400 nm, the visible 400–700 nm, and the infrared (IR) above 700 nm. Approximately half of the solar irradiance occurs in the infrared, and most of this is absorbed and converted to heat near the ocean surface. The remaining incident solar irradiance that penetrates to depths greater than 1 m is predominantly in the visible and UV, and this is regulated through optical absorption by the water, phytoplankton, and suspended particles. This latter portion of the spectrum is referred to as Photosynthetically Available Radiation (PAR) because it is the light available for photosynthesis by phytoplankton (e.g., Liu et al. 1994). PAR is defined as the 350–700 nm range of the spectrum, and accounts for 43–50% of the solar irradiance at the sea surface (Rochford et al. 2001). The vertical PAR distribution is a direct response to the intensity of incident solar irradiance flux entering at the sea surface (e.g., Austin and Petzold 1986) and its attenuation with depth ($k_{\text{PAR}}$).

As the light passes through water, it is both scattered and absorbed with different wavelengths of the visible spectrum penetrating to different depths. The influence of solar radiation differs depending on water type (Simonot and Le Treut 1986) because attenuation depth in the ocean is quite variable over the global ocean (e.g., Kara et al. 2003a). While fully spectral representations are available (Morel and Antonie 1994; Morel and Maritorena 2001), most models use simple solar irradiance approximations, such as single or bi–modal exponential parameterizations (Paulson and Simpson 1977; Zaneveld and Spinrad 1980; Simpson and Dickey 1981). The reason is that, for depths greater than 10 m, the penetrative solar flux can be accurately determined by resolving just the 300 nm to 745 nm spectral region, which is well represented by a single exponential. By using this approach, OGCM studies (Schneider and Zhu 1998; Murtugudde et al. 2002; Kara et al. 2003a) only considered solar radiation that penetrates below the top layer or level of the model
for climate simulations. In these cases, the red and near-infrared radiation is completely absorbed within the surface layer (e.g., minimum MLD of 10 m).

2.1 A New Subsurface Heating Parameterization

For accurate results within the top 10 m of the water column an OGCM with high vertical resolution needs at least two attenuation depth scales for parameterization of penetrating radiation. Traditional two-band Jerlov schemes differ significantly from single-band $k_{\text{PAR}}$ between about 3 m and 15 m, with $k_{\text{PAR}}$ giving lower attenuation than Jerlov near the surface (Figure 2.1). This is primarily due to the Jerlov split between "red" (absorbed near surface) and "blue" (more deeply penetrating) light. The single $k_{\text{PAR}}$ scheme assumes that the 51% non-visible (primarily IR) light is absorbed at the surface and treats the remaining 49% (i.e., PAR) as one band which is biased towards blue. The classical Jerlov types include both visible and IR light in its near surface red band and therefore have much less blue light, with the amount of blue light dependent on the turbidity. For example, the percentage of light left at about 2 m depth for Jerlov I, IA, IB, II and III types is 42%, 38%, 33%, 23% and 22% (all primarily in the blue spectrum), respectively (Table 2.1).

The starting point to find a new formulation for use in OGCMs is the traditional Jerlov water types at 2 m (Jerlov 1976). It is assumed that these numbers are appropriate for $k_{\text{PAR}} = 1/23, 1/20, 1/17, 1/14$ because these are the blue extinction coefficients from existing schemes. Assuming that PAR only represents the blue band, the fit is then to determine a ratio, $\gamma = Q_{\text{blue}}(0)/Q_{\text{sol}}(0)$ such that

$$Q_{\text{sol}}(2)/Q_{\text{sol}}(0) = (1 - \gamma) \exp(-2/0.5) + \gamma \exp(-2k_{\text{PAR}}),$$

matches the Jerlov table. Note that $\exp(-2/0.5) = 0.0183$ (1.83%) so this must be allowed in the fit. It is found that $\gamma = \max(0.27, 0.695 - 5.7k_{\text{PAR}})$ by assuming that I, IA, IB is more important than II (Figure 2.2). Thus, the HYCOM solar subsurface heating formulations are expressed as follows:

$$Q_{\text{sol}}(z)/Q_{\text{sol}}(0) = (1 - \gamma) \exp(-z/0.5) + \gamma \exp(-zk_{\text{PAR}}),$$

$$\gamma = \max(0.27, 0.695 - 5.7k_{\text{PAR}}).$$

This approach comes from matching $k_{\text{PAR}}$ to the traditional Jerlov scale for the fraction in each band. The percentage of shortwave radiation under $k_{\text{PAR}}$ extinction calculated from this approach is provided in Table 2.2.
Figure 2.1. Percentage of shortwave radiation remaining below sea surface as a function of depth. The \( k_{\text{PAR}} \) values used in HYCOM are compared to the ones obtained from the traditional Jerlov approach (Jerlov I, IA, IB, II, III and mud) and the ones used in a coarse vertical resolution OGCM, NLOM, (\( k_{\text{PAR}} = 0.05, 0.06, 0.08, 0.12, 0.17 \text{ m}^{-1} \) and mud). The HYCOM \( k_{\text{PAR}} \) values of 0.04345, 0.0500, 0.0588, 0.0714, 0.1266 and 2.00 \text{ m}^{-1} \) correspond to Jerlov I, IA, IB, II, III and mud cases, respectively. Note that HYCOM uses a 0.5 m e–folding depth for the red spectrum so as \( k_{\text{PAR}} \) approaches 2.00 \text{ m}^{-1}, it matters little what fraction is in the each band because both bands have small 0.5 m e–folding depth.
Table 2.1. Percentage of shortwave radiation remaining under the Jerlov extinction. The percentage values are given for each Jerlov class.

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<tr>
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<th>Jerlov IB</th>
<th>Jerlov II</th>
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In the new formulation (2.2), the red penetration scale is 2 (i.e., e-folding depth of 0.5 m), the blue penetration scale is \( k_{\text{PAR}} \), and a linear formula for the fraction of solar radiation in the blue band is used. The deep extinction rate is given by \( \gamma \exp(-z k_{\text{PAR}}) \). The new generalized HYCOM \( k_{\text{PAR}} \) values were already shown in Figure 2.1 for comparisons with the traditional Jerlov waters and the \( k_{\text{PAR}} \) approach based on a single band as used in Naval Research Laboratory Layered Ocean Model (NLOM) with an embedded mixed layer (Wallcraft et al. 2003).

It should be noted that although Lewis et al. (1990) and Siegel et al. (1995) indicated solar transmission at around 20 m depth can exceed 40 W m\(^{-2}\) for a surface irradiance of 200 W m\(^{-2}\), it is shown here that this is possible only if \( k_{\text{PAR}}=0.04 \), and if \( k_{\text{PAR}}=0.2 \), 200 W m\(^{-2}\) becomes 1 W m\(^{-2}\) at 20 m (see Table 2.2). Earlier studies (e.g., Ohlmann et al. 1998) indicated that solar
Figure 2.2. Percentage changes of $\gamma = \frac{Q_{\text{blue}}(0)}{Q_{\text{sol}}(0)}$ with $k_{\text{PAR}}$. Note that PAR is assumed to represent primarily the blue spectrum.

radiation in the visible band within the mixed layer below 10 m can be represented with a single exponential profile to within 10% accuracy. Here, it is also found that if 73% of the solar radiation is in the red exponential band, then as little as 4% of the total is left at 10 m.

The new parameterization presented here (i.e., $k_{\text{PAR}}$ based Jerlov-like 2-band scheme) is somewhat consistent with the one presented by Morel and Antonic (1994) but computationally less expensive for global applications. Morel and Antonic (1994) use a 3–band scheme (IR, red and blue) with PAR that includes both red and blue bands; while, the red band in the parameterization presented here consists of IR and red bands (i.e., non PAR plus part of PAR).

### 2.2 Remotely-Sensed Attenuation Coefficients

The OGCM that will be used in this dissertation (see chapter 3) takes solar radiation penetration into account through monthly mean attenuation depths, so that the subsurface heating can be prescribed in the model. For this purpose a monthly mean attenuation depth climatology was constructed using a remotely sensed $k_{390}$ data set (McClain et al. 1998) as acquired from the
Table 2.2. Percentage of shortwave radiation remaining under the HYCOM $k_{\text{PAR}}$ extinction based on $\gamma = Q_{\text{blue}}(0)/Q_{\text{sw}}(0)$ values. The percentage values are given for $k_{\text{PAR}}$ values of 0.0400, 0.0435, 0.0500, 0.0588, 0.0714, 0.1266, 0.2000, and 2.0000 m$^{-1}$.

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Sea-viewing Wide Field-of-view Sensor (SeaWiFS) Project during the 1997–2001 observation period. SeaWiFS ocean color data provide information on radiances emitted from the sea surface at selected wavelengths. From this information the wavelength dependence of the diffuse attenuation coefficient can be constructed using the PAR portion of the solar spectrum. Here, details of how this monthly climatology was constructed are given.

At the time this study was initiated the SeaWiFS project provided $k_{490}$ data collected from 1 October 1997 through to 4 January 2001 (McClain et al. 1998). These data have been averaged and binned by the SeaWiFS Project at NASA Goddard Space Flight Center (NASA GSFC) into daily, weekly, monthly, and annual means, and been made available as part of their Level–3 binned
data products (Darzi 1998). The monthly mean $k_{490}$ data, which are simple averages of the inferred $k_{490}$, were acquired from the available scenes of the 1997–2001 observation period. The details on the algorithms used to infer and bin the $k_{490}$ from the sensor observations are available in SeaWiFS technical reports (Firestone and Hooker 1998), and will not be given here. These data products have a resolution of 9 km ($\approx 1/12^\circ$) but with an incomplete spatial and temporal coverage. This data set was processed using the SeaWiFS Data Analysis System (SeaDAS) software (Fu et al. 1998) to produce gridded fields over the Black Sea.

As with all remotely sensed data, the SeaWiFS $k_{490}$ suffers from problems of data voids because of cloud coverage and/or infrequent sampling by satellite sensors, as well as incorrect and biased observations (noise) because of sensor limitations. The SeaWiFS algorithm used here comes from the initial processing of the distributed data archive and does not have the corrections that are applied for turbid water regions in the reprocessed data. Even with the large number of scenes used to construct the monthly $k_{490}$ data sets, they contain ocean regions of sparse data.

To fill in these data voids in the monthly $k_{490}$, the first step is to interpolate over time using a one-dimensional periodic cubic spline with a tension factor of $\sigma = 5.0$ for each location having a data void during the year. This tension factor was found to be optimal after a few tests. In general, a tension factor greater than zero causes the cubic spline to be blended with or converge to a linear spline based on the same set of points, and a tension factor of unity causes the cubic spline to coincide with the linear spline. To exclude the incorrect values that can arise due to spurious oscillations in the cubic splines, interpolated values that fall outside a range of values uniquely determined for each data void location were rejected. A suitable range is found to be 25% of the minimum value to 125% of the maximum value of those monthly mean $k_{490}$ values used to construct the spline coefficients. Data voids in a given month are left unaltered as a first step if the interpolated value is rejected.

After temporal filling, the remaining data voids are replaced by next applying a statistical interpolation (SI) method (Daley 1991) in space for each individual monthly mean, with an exponential correlation function assumed for the covariance (see Appendix A). This approach successfully yields monthly $k_{490}$ means of complete coverage. However, the data sets still suffer from noise problems inherited from the satellite observations.

To remove the noise, a 3-step process is applied to the $k_{490}$ data. First, the data is checked again for anomalously large or small values. Using the same interpolating grid boxes as in the SI, the mean and standard deviation are calculated for all data values except the one at the box center. The data value at the central location is considered anomalous if it falls outside of the one-standard
deviation interval for this distribution. Its replacement value is determined by applying SI again. The latest data values are used when making these replacements, and can include SI values inserted from earlier in the search. Second, the condition \( k_{490} \geq 10^{-5} \text{ m}^{-1} \) is imposed on all the data values to avoid an infinite \( 1/k_{490} \) depth scale, should there be such a future need in an OGCM application. The minimum limit of \( k_{490} = 10^{-5} \text{ m}^{-1} \) is chosen so \( 1/k_{490} \) is comparable to the maximum ocean depth. Third, to remove the remaining minor residual noise, a \( 3 \times 3 \) spatial filter with a 1:2:1 weighting (\( F = 1.0, 2.0, 1.0, 2.0, 4.0, 2.0, 1.0, 2.0, 1.0 \)) is then passed once over the data to smooth it to a level suitable for HYCOM applications.

Using these monthly \( k_{490} \) data sets, the monthly \( k_{\text{PAR}} \) climatology is constructed with the Zaneveld et al. (1993) regression relations. The monthly mean \( k_{\text{PAR}} \) fields were then computed from the \( k_{490} \) and spatially interpolated to the HYCOM domain as shown in Figure 2.3.

Attenuation depth values (\( 1/k_{\text{PAR}} \)) usually range from \( \approx 3 \) to 10 m (\( k_{\text{PAR}} \) values of \( \approx 0.3 \) to 0.1 \( \text{m}^{-1} \), respectively) in all months but values in summer (i.e., July and August) are relatively small. In comparison to the interior of the Black Sea, the coastal waters typically have small \( k_{\text{PAR}} \) values (i.e., less turbidity) especially in spring and summer. This is partly due to the fresh water from the rivers around the region. The easternmost region of the Black Sea has the lowest turbidity, and western part of the region (\( \approx \) west of 33\(^\circ\)E) has the largest turbidity. This is generally true for all months. The transparency of the western region has maximum values between 44\(^\circ\)–45\(^\circ\)N latitudes and 30\(^\circ\)–33\(^\circ\)E longitudes. Regions where turbidity is the largest usually occur in the western part during November and December.

The monthly time series of \( k_{\text{PAR}} \) at some selected locations (Figure 2.4) clearly reveal that there is not a strong seasonal variation in the Black Sea. However, \( k_{\text{PAR}} \) values in summer are generally smaller in comparison to the ones in other months. At some locations, \( k_{\text{PAR}} \) can even be very large. For example, \( k_{\text{PAR}} \) can be as large as 0.48 \( \text{m}^{-1} \) and 0.34 \( \text{m}^{-1} \) at (29.1\(^\circ\)E, 41.4\(^\circ\)N) and (31.0\(^\circ\)E, 45.5\(^\circ\)N), respectively. It is noted that one should be interested in regions with small \( k_{\text{PAR}} \) values (i.e., \( k_{\text{PAR}} < 0.30 \text{ m}^{-1} \)) because they may induce a change in upper ocean quantities, and all large \( k_{\text{PAR}} \) values (i.e., \( k_{\text{PAR}} > 0.30 \text{ m}^{-1} \)) are essentially uniform (see Figure 2.1).

It should also be emphasized that strong density stratification in the Black Sea effectively inhibits vertical mixing and ventilation of sub-pycnocline waters from the surface. This results in permanent contamination in the region. Similarly, the Black Sea experiences large volumes of nutrients and contaminants from the Danube, Dnieper and Dniester along the northwestern shelf (e.g., Mee 1992). All of these factors support the idea that the Black Sea OGCM studies require ocean turbidity at high spatial resolution, and they should be part of the heat flux forcing, so that
Figure 2.3. Climatological monthly mean attenuation of photosynthetically available radiation ($k_{\text{PAR}}$) over the Black Sea from January through December. These fields were processed from the remotely-sensed monthly attenuation coefficient at 490 nm ($k_{490}$) which were acquired from SeaWiFS data set. The climatology has high spatial resolution of 9 km ($\approx 1/12^\circ$) over the Black Sea during 1997–2001.

the time-varying solar penetration schemes can treat attenuation as a continuous quantity. This is an improvement over using a few discrete attenuation values corresponding to classical Jerlov water types (Jerlov 1976).

### 2.3 Shortwave Radiation and Attenuation Depth

The main interest is in the amount of shortwave radiation that penetrates through the mixed layer. Thus, spatial variations of the annual mean of shortwave radiation over the sea surface are
Figure 2.4. Climatological annual cycle of $k_{\text{PAR}}$ (m$^{-1}$) at five different locations over the Black Sea.

shown in Figure 2.5a,b. These fields were obtained from two different sources: (1) European Centre for Medium-Range Weather Forecasts (ECMWF) re-analysis product (ECMWF 1995; Gibson et al. 1997), and (2) Fleet Numerical Meteorology and Oceanography Center (FNMOC) Navy Operational Global Atmospheric Prediction System (NOGAPS) data (Rosmond et al. 2002). An annual mean was formed during 1979–1993 for ECMWF and 1998–2002 for NOGAPS. In chapter 3, ECMWF and NOGAPS outputs will be used to investigate model sensitivity to the choice of atmospheric forcing product. While spatial distributions of shortwave radiation are different between the ECMWF and NOGAPS, they usually have large values in the interior in comparison to coastal regions. The basin averaged annual mean shortwave radiation values are 135 and 163 W m$^{-2}$ for ECMWF and NOGAPS, respectively. The amount of shortwave radiation that enters the mixed layer depends on the attenuation depths as mentioned in chapter 2a.

The annual mean $k_{\text{PAR}}$ processed from the SeaWiFS data is shown in Figure 2.5c. The large $k_{\text{PAR}}$ values are a typical feature of the attenuation coefficient in the Black Sea. The decreased transparency over the most of the Black Sea is due to the greater biological productivity that occurs from increased availability of nutrients trapped in the euphotic zone since there is not much ventilation and the mixed layer is usually shallow over the most of the region. For the
Figure 2.5. Climatological annual mean of various quantities over the Black Sea: (a) shortwave radiation at the sea surface (W m$^{-2}$) constructed from the 1.125° × 1.125° ECMWF re-analysis product (1979–1993), (b) shortwave radiation at the sea surface (W m$^{-2}$) constructed from the 1° × 1° NOGAPS product (1998–2002), (c) attenuation of Photosynthetically Available Radiation (m$^{-1}$) processed from the 1/12° SeaWiFS $k_{490}$ data (1997–2001) as explained in the text, and (d) ocean mixed layer depth (m) calculated from GDEM data based on the layer definition of Kara et al. (2000). Both ECMWF and NOGAPS provide relatively high frequency atmospheric forcing. Note that the color bars for (a) and (b) are same.

present application, the greatest interest is in those situations where subsurface heating can occur within/below the mixed layer. Large $k_{\text{PAR}}$ values might produce complete (or almost complete) absorption of the solar radiation within the mixed layer. In the western part, $k_{\text{PAR}}$ values are usually greater than 0.25 m$^{-1}$ which corresponds to an attenuation depth of 4 m. This implies that a turbidity increase in the region causes increased heating within the mixed layer because the climatological MLD is relatively shallow (Figure 2.5d). Note that the MLD is calculated from monthly mean temperature and salinity data obtained from the new 1/4° Generalized Digital Environmental Model (GDEM) climatology (NAVOCEANO, personal communication)
that includes 78 levels in vertical. The mixed layer depth (MLD) is defined as the depth at the base of an isopycnal layer, where the density has changed by a variable amount of $\Delta \sigma_t = \sigma_t(T + \Delta T, S, P) - \sigma_t(T, S, P)$ from the density at a reference depth of 1 m based on a $\Delta T=0.5^\circ$C, where $S$ is the salinity, $T$ is the temperature and $P$ is pressure which is set to zero. The reader is referred to Kara et al. (2000) for a more detailed definition of MLD.
CHAPTER 3

HYCOM DESCRIPTION

HYCOM is a generalized (hybrid isopycnal/terrain-following (σ)/z-level) coordinate primitive equation model as described in Bleck (2002), in detail. The hybrid coordinate extends the geographic range of applicability of traditional isopycnic coordinate circulation models (e.g., Bleck et al. 1992) toward shallow coastal seas and unstratified parts of the ocean. Typically, HYCOM behaves like σ-coordinate model in shallow, unstratified coastal regions, like a z-level coordinate model in the mixed layer and other unstratified regions, and like an isopycnic-coordinate model where the ocean is stratified. The model makes a dynamically smooth transition between the coordinate type via the layered continuity equation. The choice of coordinate type varies in time and space, and the optimal choice is updated every time step.

The hybrid coordinate is an important feature for the Black Sea because of the interaction between very shallow coastal regions where the Rim Current is usually present and the interior region. Hybrid coordinates are also used for meteorological modeling. For example, the Rapid Update Cycle (RUC), a hybrid-isentropic weather prediction model, has been in use since the mid-1990s. The capability to assign additional coordinate surfaces to the oceanic mixed layer gives the option of using the K-Profile Parameterization (KPP) sub-model (Large et al. 1994) for vertical mixing in HYCOM. It provides mixing from surface to bottom, smoothly matching the large surface boundary layer diffusivity/viscosity profiles to the weak diapycnal diffusivity/viscosity profiles of the interior ocean (Large et al. 1997).

HYCOM contains a total of five prognostic equations: two momentum equations for the horizontal velocity components, a mass continuity or layer thickness tendency equation and two conservation equations for a pair of thermodynamic variables, such as salt and temperature or salt and density. The model equations, written in \((x, y, s)\) coordinates, where \(s\) is generalized vertical coordinate, are
\[
\frac{\partial \mathbf{v}}{\partial t_s} + \nabla_s \frac{\mathbf{v}^2}{2} + (\zeta + f) \mathbf{k} \times \mathbf{v} + \left( \frac{\partial s}{\partial t} \right) \frac{\partial \mathbf{v}}{\partial s} + \nabla_s M - p \nabla_s \alpha = \]

\[
-g \frac{\partial \tau}{\partial p} + \left( \frac{\partial p}{\partial s} \right)^{-1} \nabla_s \cdot \left( \nu \frac{\partial \mathbf{v}}{\partial s} \right), \tag{3.1}
\]

\[
\frac{\partial}{\partial t_s} \left( \frac{\partial p}{\partial s} \right) + \nabla_s \cdot \left( \frac{\partial \mathbf{v}}{\partial s} \right) + \frac{\partial}{\partial s} \left( \frac{\partial p}{\partial t} \frac{\partial \theta}{\partial s} \right) = \nabla_s \cdot \left( \nu \frac{\partial \mathbf{v}}{\partial s} \theta \right) + \mathfrak{h}_\theta, \tag{3.2}
\]

\[
\frac{\partial}{\partial t_s} \left( \frac{\partial p}{\partial s} \right) + \nabla_s \cdot \left( \mathbf{v} \frac{\partial p}{\partial s} \right) + \frac{\partial}{\partial s} \left( \frac{\partial p}{\partial t} \frac{\partial \theta}{\partial s} \right) = 0, \tag{3.3}
\]

where \( \mathbf{v} = (u, v) \) is the horizontal velocity vector, \( p \) is pressure, \( \theta \) represents any one of the model’s thermodynamic variables, \( \alpha = 1/\rho \) is the potential specific volume, \( \zeta = \partial v/\partial x_s - \partial u/\partial y_s \) is the relative vorticity, \( M = g z + p \alpha \) is the Montgomery potential, \( g z = \phi \) is the geopotential, \( f \) is the Coriolis parameter, \( \mathbf{k} \) is the vertical unit vector, \( \nu \) is a variable eddy viscosity/diffusivity coefficient, and \( \tau \) is the wind- and/or bottom-drag induced shear stress vector. The variable \( \mathfrak{h}_\theta \) represents the sum of diabatic source terms, including diapycnal mixing acting on \( \theta \). Subscripts show which variable is held constant during partial differentiation. Distances in \( x, y \) direction, as well as their time derivatives \( u \) and \( v \), respectively, are measured in the projection onto a horizontal plane.

The prognostic equations in the model are time-integrated using the split-explicit treatment of barotropic and baroclinic modes (Bleck and Smith 1990). The split-explicit approach has proven to be advantageous for executing ocean models on massively parallel computers because it does not require solution of an elliptic equation. Isopycnal diffusivity and viscosity values, including the one used for thickness diffusion (interface smoothing), are formulated as \( u_d \Delta x \) where \( \Delta x \) is the local horizontal mesh size and \( u_d \) is of order 0.01 m s\(^{-1}\). In regions of large shear, isopycnal viscosity is set proportional to the product of mesh-size squared and total deformation (Smagorinsky 1963), the proportionality factor being 0.2.

HYCOM uses KPP mixed layer model, which is particularly attractive for several reasons. It contains improved parameterizations of physical processes in the mixed layer, including non-local effects. It calculates the mixing profile from the surface to the bottom of the water column, and thus provides an estimate of diapycnal mixing beneath the mixed layer. It has also been designed to run with relatively low vertical resolution, an advantage for OGCMs. It provides mixing throughout the water column with an abrupt but smooth transition between the vigorous mixing within the surface boundary layer and the relatively weak diapycnal mixing in the ocean interior. In the ocean interior, the contribution of background internal wave breaking, shear instability mixing,
and double diffusion (both salt fingering and diffusive instability) are parameterized in the KPP model. In the surface boundary layer, the influences of wind-driven mixing, surface buoyancy fluxes, and convective instability are parameterized. The KPP algorithm also parameterizes the influence of nonlocal mixing of temperature and salinity, which permits the development of countergradient fluxes.

The upper ocean fields analyzed throughout this dissertation are sea surface height (SSH) and mixed layer depth (MLD). The SSH in HYCOM is a diagnostic quantity, which is sum of the Montgomery potential and barotropic pressure. The MLD is a diagnostic quantity which depends on a specified density difference with respect to the surface. In general, the MLD is the first depth at which the density jump with respect to the surface is the equivalent of 0.2°C. By converting to density one can allow for salinity driven mixed layers. The conversion is done at the surface (i.e., with respect to sea surface temperature and sea surface salinity) so HYCOM converts the temperature jump to a local density jump and then finds the shallowest depth. The KPP mixed layer model is applied to every horizontal point independently. This sometimes leads to a very noisy MLD. A 9-point smoothing on the MLD fields is used throughout the analysis. This is a standard uniform grid 9-point smoother, and it does not allow for the differences in area between grid cells.

3.1 Black Sea Model

This section describes the Black Sea model set up based on HYCOM. A brief explanation for model resolution, model constants and atmospheric forcing used in the model simulations are also given.

3.1.1 Model Resolution and Constants

The Black Sea model used here has $[1/25° \times 1/25° \cos(\text{lat}), \text{long} \times \text{lat}]$ resolution on Mercator grid. Longitudinal and latitudinal array sizes in the model are 360 and 206, respectively. This Mercator grid has square grid cells with a resolution of $(0.04 \times \cos(\text{lat}) \times 111.2 \text{ km})$, i.e., $\approx 3.4 \text{ km}$ at the southern and $3.1 \text{ km}$ at the northern coast.

There are a total of 15 hybrid layers in the model, and the target density ($\sigma_l$) values corresponding to layers 1 through 15 are 6.00, 9.00, 10.00, 11.00, 12.00, 12.80, 13.55, 14.30, 15.05, 16.20, 16.80, 16.95, 17.05, 17.15 and 17.20, respectively. These were primarily based on the basin averaged climatological mean temperature and density fields (see section 3.1.3), with each layer
Table 3.1. Constant parameters used in the HYCOM Black Sea model.

<table>
<thead>
<tr>
<th>Value</th>
<th>Description of the constant used in the Black Sea Model</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.1</td>
<td>deformation–dependent Laplacian viscosity factor</td>
</tr>
<tr>
<td>5.e-3</td>
<td>diffusion velocity for Laplacian momentum dissipation (m s(^{-1}))</td>
</tr>
<tr>
<td>67.e-4</td>
<td>diffusion velocity for biharmonic momentum dissipation (m s(^{-1}))</td>
</tr>
<tr>
<td>0.05</td>
<td>diffusion velocity for biharmonic thickness diffusion (m s(^{-1}))</td>
</tr>
<tr>
<td>5.e-2</td>
<td>diffusion velocity for Laplacian temp/saln diffusion (m s(^{-1}))</td>
</tr>
<tr>
<td>2.e-3</td>
<td>coefficient of quadratic bottom friction</td>
</tr>
<tr>
<td>10.0</td>
<td>thickness of bottom boundary layer (m)</td>
</tr>
<tr>
<td>0.02</td>
<td>minimum density jump across interfaces (kg m(^{-3}))</td>
</tr>
<tr>
<td>0.2</td>
<td>equivalent temperature jump across mixed–layer (°C)</td>
</tr>
<tr>
<td>30.0</td>
<td>reference mixed–layer thickness for SSS relaxation (m)</td>
</tr>
<tr>
<td>0.3</td>
<td>KPP critical bulk richardson number</td>
</tr>
<tr>
<td>0.7</td>
<td>KPP value for calculating shear instability</td>
</tr>
<tr>
<td>50.e-4</td>
<td>KPP max viscosity and diffusivity due to shear instability (m(^2) s(^{-1}))</td>
</tr>
<tr>
<td>1.e-4</td>
<td>KPP: background/internal wave viscosity (m(^2) s(^{-1}))</td>
</tr>
<tr>
<td>1.e-5</td>
<td>KPP background/internal wave diffusivity (m(^2) s(^{-1}))</td>
</tr>
<tr>
<td>10.e-4</td>
<td>KPP salt fingering diffusivity factor (m(^2) s(^{-1}))</td>
</tr>
<tr>
<td>1.9</td>
<td>KPP salt fingering factor</td>
</tr>
<tr>
<td>98.96</td>
<td>KPP value for nonlocal flux term</td>
</tr>
<tr>
<td>10.0</td>
<td>KPP value for nonlocal flux adjustment term</td>
</tr>
<tr>
<td>1.8</td>
<td>KPP value for turbulent shear contribution to bulk Ri number</td>
</tr>
<tr>
<td>5.0</td>
<td>KPP value for turbulent velocity scale</td>
</tr>
</tbody>
</table>

Increasingly thick, the abyssal layer being the thickest. These target density values were found to be optimal in comparison to other layer density sets based on a series of tuning experiments. Based on the layer selection, the wide and shallow shelf area in the northwestern part is resolved by the first few model layers.

Constant parameters used in the Black Sea model is provided in Table 3.1. The sensitivity of the Black Sea model to the choice of KPP parameters was investigated by undertaking a tuning exercise to find an optimal set in providing a realistic sea surface temperature (SST) from realistic atmospheric forcing over as much of the Black Sea as possible. This was done by comparing the monthly satellite–based SST climatology to monthly averages from the model, which is similar to a statistical tuning methodology, such as introduced in Wallcraft et al. (2003). Overall, the original KPP constants reported in Large et al. (1994) worked best in the Black Sea model although it has also been confirmed that new KPP constants (e.g., KPP value for turbulent shear contribution to bulk Ri number and KPP value for calculating shear instability) suggested in Smyth et al. (2002) did not cause any significant changes in the model simulations.
Figure 3.1. The bottom topography (m) and the model coastline used in HYCOM simulations. Note that the bottom topography includes 1-minute DBDB–V data. The model land-sea boundary is the 10–m isobath which is shown with light brown in color. The Sea of Azov is excluded in the model simulations.

3.1.2 Bottom topography

Most previous Black Sea model studies used UNESCO bathymetric maps to represent bottom topography (e.g., Staney and Stenava 2001; Staney and Staneva 2000), and a few studies used the Earth Topography Five Minute Grid (ETOP05) data (e.g., Oguz et al. 1995; Staneva et al. 2001). While ETOP05 (NOAA 1988) is usually adequate for deeper parts of the ocean, it is less reliable over continental shelves (< 200 m deep) because of mismatches in the chart isobaths used in constructing the data set. As a result, it poses a problem for the Black Sea because of the existence of the wide continental shelf.

The bottom topography in the Black Sea model (Figure 3.1) was constructed from Digital Bathymetric Data Base–Variable resolution (DBDB–V) which was developed by Naval Oceanographic Office (NAVOCEANO) to support the generation of bathymetric chart products. It provides bathymetric data to be integrated with other geophysical and environmental parameters.
for ocean modeling. The digitally rendered contours are put through a gridding routine. This routine takes the values that fall within a grid node area of influence, utilizing a multi-stage minimum–curvature spline algorithm. The addition of Digital Bathymetric Data Base–Variable resolution (DBDB–V) data (NAVOCEANO 1997; 2001) alone introduces significant improvements into the final topography used in the model. DBDB–1.0 is the 1′ (1 minute) subregion of NAVOCEANO’s DBDB–V that covers the Mediterranean Sea including the Aegean Sea and Black Sea. After interpolation to the Black Sea model’s grid, it was also smoothed twice with a 9-point smoother to reduce energy generation at small scales.

In general, there is an almost flat abyssal plain, a flat wide shelf in the northwestern part and a steep continental slope along the Turkish coast (see Figure 3.1). The abyssal plain of the Black Sea is ≈ 2000 m deep, and the maximum depth is ≈ 2204 m. The basin does not include large gulfs or islands but one of the most interesting features is the narrow continental shelf. The northwestern part of the Black Sea is covered by a relatively wide shelf communicating with the deep interior basin through a relatively wide continental slope. This shelf is less than 200 m deep and gradually narrows toward the southwest. The major shelf region of the Black Sea is only found along the western coast where it also receives the freshwater discharges from the Danube, Dnieper and Dniester.

3.1.3 Temperature and Salinity Initialization

Previous modeling studies of the Black Sea were limited to use of local data sets constructed from sparse observations for the initial temperature and salinity climatology (e.g., Altman et al. 1987; Truchchew and Demin 1992). There is also uncertainty in the quality of these existing data sources in the Black Sea (e.g., Staneva and Stanev 1998). A well-documented commonly used subsurface temperature and salinity climatology from the 1° × 1° World Ocean Atlas 1994 (Levitus et al. 1994; Levitus and Boyer 1994) does not cover the Black Sea region. While a recently updated World Ocean Atlas (Conkright et al. 2002) includes monthly mean temperature and salinity climatologies, it is still at 1° × 1° grid resolution, which does not resolve the coastal and shelf regions. Thus, the Black Sea suffers from lack of quality fine resolution subsurface temperature and salinity climatologies which can be used for model initialization.

In this study, HYCOM is initialized using the temperature and salinity from the Modular Ocean Data Assimilation System (MODAS) climatology developed at the NRL (Fox et al. 2002) because it provides increased horizontal resolution. The climatology has variable grid resolution: 1/8° near
Figure 3.2. Annual mean of basin averaged potential temperature, salinity and potential density profiles obtained from the MODAS climatology. Standard depth levels in the climatology are 0, 10, 20, 30, 50, 75, 100, 125, 150, 200, 250, 300, 400, 500, 600, 700, 800, 900, 1000, 1100, 1200, 1300, 1400, 1500, 1750, 2000, 2200 m.

land, 1/4° over shallow shelves, and 1/2° in the open ocean. This makes the MODAS climatology a candidate to use in studying the surface circulation. In general, the MODAS climatology is one of the current Navy standard tools for production of three dimensional grids of temperature and salinity, and derived quantities such as density (Harding et al. 1999). The MODAS includes both a static climatology and a dynamic climatology. The static climatology represents the historical averages, while the dynamic climatology combines near real-time observations of sea surface height and sea surface temperature data with other local observations from ships, aircraft, or buoys to produce a three-dimensional analysis of the ocean temperature and salinity structure. The final products are quality-controlled gridded analysis fields as outputs.

Climatological potential temperature, salinity and density fields from MODAS are output at 27 depth levels ranging from 0 to 2200 m for the Black Sea (Figure 3.2). The MODAS climatology does not include temperature/salinity below 1750 m in the Black Sea because the climatology is relaxed to the World Ocean Atlas 1994 at deeper layers, which is void of data in this region. In
this case, a simple linear extrapolation was applied to the temperature and salinity data, and then the profiles were extended down to \( \approx 2200 \) m.

In general, the Black Sea can be described as a two-layer system with a thin low salinity surface layer overlying a relatively uniform higher salinity deep layer (see Figure 3.2). Murray et al. (1991) and Özsoy et al. (1993) showed that the Cold Intermediate Layer (CIL) is characterized by temperatures less than 8°C, which occurs within a typical range of about 50 to 180 m overlying the main pycnocline where vertical stratification is maintained by the salinity gradient. This layer is located under the summer thermocline and above the halocline. Salinity increases from 17.8 psu at the sea surface to 23.0 psu at 500 m. In the northwestern shelf freshwater from the rivers (e.g., Danube) help to maintain the low salinity in the surface waters.

### 3.1.4 Wind and Thermal Forcing

The model reads in the following time-varying atmospheric forcing fields: wind stress and thermal forcing (air temperature and air mixing ratio at 10 m above the sea surface, net shortwave radiation and net longwave radiation at the sea surface). In this dissertation, the HYCOM simulations use wind/thermal forcing constructed from two different archived operational weather center products: (1) \( 1.125^\circ \times 1.125^\circ \) ECMWF Re–Analysis (ERA) climatology (1979–1993), and (2) \( 1^\circ \times 1^\circ \) NOGAPS climatology (1998–2002).

All model simulations are performed using climatological monthly mean forcing fields. However, a high frequency component is added to the climatological forcing because the mixed layer is sensitive to variations in surface forcings on time scales of a day or less (e.g., Wallcraft et al. 2003; Kara et al. 2003b) and because the future goal is to perform simulations forced by high frequency interannual atmospheric fields from operational weather centers. Hybrid winds consist of monthly ECMWF (or NOGAPS) plus ECMWF wind anomalies. Construction of the ECMWF hybrid winds is briefly described here. The same procedure is applied to NOGAPS. For HYCOM wind stress forcing \( \bar{\tau}_{HYCOM} \), 6–hourly intra–monthly anomalies from ECMWF are used in combination with the monthly mean wind stress climatology of ECMWF (NOGAPS) interpolated to 6 hourly intervals. The 6 hourly anomalies are obtained from a reference year. For the purpose the winds from September 1994 through September 1995 (6–hourly) are used, inclusive, because they represented a typical annual cycle of the ECMWF winds, and because the September winds in 1994 and 1995 most closely matched each other. The 6–hourly September 1994 and September 1995 wind stresses are blended to make a complete annual cycle, which is denoted by \( \bar{\tau}_{ECMWF} \).
The ECMWF wind stresses are calculated from ECMWF 10 m winds using the bulk formulae of Kara et al. (2002). Monthly averages are first formed from the September 1994 through September 1995 ECMWF wind stresses ($\bar{\tau}_{ECMWF}$) to create the ECMWF wind stress anomalies ($\bar{\tau}_A$), and are then linearly interpolated to the time intervals of the 6-hourly ECMWF winds to produce a wind stress product ($\bar{\tau}_f$). The anomalies are then obtained by applying the difference $\bar{\tau}_A = \bar{\tau}_{ECMWF} - \bar{\tau}_f$. Scalar wind speed is obtained from the input wind stress and therefore has 6-hourly variability. Note that NOGAPS is used directly, and the high frequency component ($\bar{\tau}_h$) is added to the interpolated monthly mean wind stress from NOGAPS.

The net surface heat flux that has been absorbed (or lost) by the upper ocean to depth $z$, $Q(0)$, is parameterized as the sum of the downward surface solar irradiance ($Q_{sol}$), upward longwave radiation ($Q_{LW}$), and the downward latent and sensible heat fluxes ($Q_L$ and $Q_S$, respectively).

$$Q(z) = (Q_{sol}(0) - Q_{sol}(z)) - Q_{LW} + Q_L + Q_S,$$

(3.4)

where $Q_{sol}$ as described in chapter 2. Here $Q_{sol}(z)$ is the amount of solar radiation that penetrates to depth. The rate of heating/cooling of each layer is simply obtained by evaluating (3.4) at the bottom and top of the layer, with only $Q_{sol}(z)$ non-zero below layer 1. The model reads in monthly mean $k_{PAR}$ fields produced from SeaWiFS $k_{PAR}$ data to include effects of turbidity as explained in chapter 2, in detail.

Latent and sensible heat fluxes at the air–sea interface are calculated using efficient and computationally inexpensive bulk formulae that include the effects of dynamic stability (Kara et al. 2002). The combination of accuracy and ease of computation makes this method preferred for computing air–sea fluxes in the HYCOM. Note that both sensible and latent heat fluxes are calculated using top layer temperature at each model time step. Radiation flux (shortwave and longwave fluxes) is so dependent on cloudiness that this is taken directly from ECMWF (or NOGAPS) for use in the model.

The annual climatological SST cycle is built into the model to a limited extent. Including air temperature and model SST in the formulations for latent and sensible heat flux automatically provides a physically realistic tendency towards the correct SST. If the model SST is too high/low, the flux is reduced/increased relative to that from the correct SST. The trend towards reality is typically sufficient on its own to keep the model SST approximately on track.

River input to HYCOM is treated as precipitation. This works independently of any other surface salinity forcing. Looking more closely at the largest rivers in a given ocean model domain is important to represent evaporation and precipitation effects properly. However, the problem is
Table 3.2. Climatological annual mean flow values obtained from four climatological data sets for the rivers discharged into the Black Sea. The source for the Perry data is Perry et al. (1996). Other two data sets are available online. The NRL data set is similar to RivDIS except a scale factor that is obtained from Perry et al. (1996) to correct the annual mean. Note that 1 Sv = 10⁶ m³ s⁻¹ ≈ 32000 km³ y⁻¹.

<table>
<thead>
<tr>
<th>River</th>
<th>Perry (m³ s⁻¹)</th>
<th>RivDIS (m³ s⁻¹)</th>
<th>UCAR (m³ s⁻¹)</th>
<th>NRL (m³ s⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Danube</td>
<td>6365.0</td>
<td>6499.0</td>
<td>6413.6</td>
<td>6114.1</td>
</tr>
<tr>
<td>Dniepr</td>
<td>1630.8</td>
<td>1482.2</td>
<td>1483.7</td>
<td>1630.0</td>
</tr>
<tr>
<td>Rioni</td>
<td>409.7</td>
<td>408.6</td>
<td>402.7</td>
<td>409.7</td>
</tr>
<tr>
<td>Dniestr</td>
<td>326.3</td>
<td>375.2</td>
<td>324.3</td>
<td>326.3</td>
</tr>
<tr>
<td>Sakarya</td>
<td>217.3</td>
<td>192.3</td>
<td>192.8</td>
<td>217.6</td>
</tr>
<tr>
<td>Kizilirmak</td>
<td>180.5</td>
<td>201.8</td>
<td>202.2</td>
<td>202.2</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>River</th>
<th>Perry (km³ y⁻¹)</th>
<th>RivDIS (km³ y⁻¹)</th>
<th>UCAR (km³ y⁻¹)</th>
<th>NRL (km³ y⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Danube</td>
<td>203.7</td>
<td>205.0</td>
<td>205.2</td>
<td>195.7</td>
</tr>
<tr>
<td>Dniepr</td>
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<td>46.7</td>
<td>47.5</td>
<td>52.2</td>
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<td>Rioni</td>
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<td>12.9</td>
<td>12.9</td>
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</tr>
<tr>
<td>Dniestr</td>
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<td>5.8</td>
<td>6.4</td>
<td>6.5</td>
<td>6.5</td>
</tr>
</tbody>
</table>

Knowing that one data source is more accurate than another. There are several readily available river discharge climatologies, and climatological annual mean river flow values constructed from these sources are given in Table 3.2 for each river discharged into the Black Sea (Table 3.3). Note that there are also other rivers which are not listed in the table, such as the Southern Bug (Ukranie) and Kamtehiya (Bulgaria), discharged into the Black Sea; however, they are not input to the model because the contribution from these rivers is very small.

HYCOM reads in monthly mean river discharge values. The monthly mean RivDIS climatology (Vörösmarty et al. 1997; 1998) is preferred for use in HYCOM because it gives river inflow values at the mouth of the river. A total of 6 major rivers is used as precipitation forcing. The Danube River has the largest discharge into the Black Sea with a river flow of 6365.0 m³ s⁻¹ (0.006365 Sv). In the simulations described here HYCOM does not include the direct Bosporus inflow, causing an imbalance in evaporation minus precipitation. This imbalance in the model was handled by adding a negative river precipitation (i.e., a river evaporation) for the Bosporus. In summary, the model treats rivers as a “runoff” addition to the surface precipitation field. The flow is first applied to a
Table 3.3. Time periods during which climatological river discharges were constructed. Total number of years for the climatology is also included. Note that Perry data set was constructed using the annual mean from individual years so the total number of years represents the total of these individual years; while, the other two data sets were constructed using monthly mean discharge values.

<table>
<thead>
<tr>
<th>River</th>
<th>Perry Climatology</th>
<th>RivDIS Year</th>
<th>RivDIS Climatology</th>
<th>UCAR Year</th>
</tr>
</thead>
</table>

Table 3.4. HYCOM simulations used in this dissertation. All simulations were performed using 64 processors on several massively parallel supercomputers. The model uses climatological wind and thermal forcing (i.e., air temperature at 10 m, air mixing ratio at 10 m, shortwave and longwave radiation) constructed from European Centre for Medium Range Weather Forecasts (ECMWF) re-analysis product during 1979–1993 and Fleet Numerical Meteorology and Oceanography Center (FNMOC) Navy Operational Global Atmospheric Prediction System (NOGAPS) data during 1998–2002.

<table>
<thead>
<tr>
<th>Expt</th>
<th>$k_{PAR}$</th>
<th>Description of the experiment</th>
<th>Forcing</th>
</tr>
</thead>
<tbody>
<tr>
<td>expt 1</td>
<td>Variable</td>
<td>Spatial and temporal attenuation depths</td>
<td>ECMWF</td>
</tr>
<tr>
<td>expt 2</td>
<td>99 m$^{-1}$</td>
<td>All solar radiation absorbed at the surface</td>
<td>ECMWF</td>
</tr>
<tr>
<td>expt 3</td>
<td>0.06 m$^{-1}$</td>
<td>Constant attenuation depth in the Black Sea</td>
<td>ECMWF</td>
</tr>
<tr>
<td>expt 4</td>
<td>Variable</td>
<td>Spatial and temporal attenuation depths</td>
<td>NOGAPS</td>
</tr>
<tr>
<td>expt 5</td>
<td>99 m$^{-1}$</td>
<td>All solar radiation absorbed at the surface</td>
<td>NOGAPS</td>
</tr>
<tr>
<td>expt 6</td>
<td>0.06 m$^{-1}$</td>
<td>Constant attenuation depth in the Black Sea</td>
<td>NOGAPS</td>
</tr>
</tbody>
</table>

single ocean grid point and smoothed over surrounding ocean grid points, yielding a contribution to precipitation in m s$^{-1}$.

3.1.5 Model Simulations

Climatologically-forced model simulations that use three different sets of $k_{PAR}$ values are performed to investigate the effects of ocean turbidity on the upper ocean quantities. Table 3.4 gives a brief explanation of each simulation. For expt 1, spatially and monthly varying $k_{PAR}$ values interpolated to the HYCOM grid are used. For expt 2, all solar radiation is absorbed at the sea surface by using a unrealistically large $k_{PAR}$ value of 99.9 m$^{-1}$. For expt 3, the ocean turbidity over
the Black Sea is set to a constant, \( k_{\text{PAR}} = 0.06 \, \text{m}^{-1} \), which is a representative value for clear water (e.g., Kara et al. 2003a). These experiments (i.e., expts 1, 2 and 3) use wind/thermal forcing from ECMWF; while, expts 4, 5 and 6 are essentially twins of expts 1, 2 and 3 but use wind/thermal forcing from NOGAPS. All HYCOM simulations presented in this dissertation were performed with no assimilation of any oceanic data except initialization from climatology and relaxation to sea surface salinity.

The simulations were performed using several massively parallel supercomputers. The model array size is 360 \( \times \) 206, and performing a 1-month simulation takes \( \approx 2.3 \) wall-clock hours using 64 IBM SP POWER 3 processors or \( \approx 0.7 \) wall-clock hours using 64 HP/COMPACT SC45 processors. The model was run until it reached statistical equilibrium using 6 hourly climatological forcing as described earlier. A linear regression analysis was performed for domain averaged parameters (layer temperature, salinity, potential and kinetic energy, etc.) to investigate statistical equilibrium in each layer, and is expressed numerically as % change per decade. The model is deemed to be in statistical equilibrium when the rate of potential energy change is acceptably small (e.g., < 1% in 5 years) in all layers. It takes about 5–8 model years for a simulation to reach equilibrium.
CHAPTER 4

RESULTS AND DISCUSSION

Sensitivity of model results to water turbidity is examined with a particular focus on upper ocean currents and sea surface height (SSH). All model results are presented based on annual and monthly means that were constructed from the last 4 years of the model simulations. At least a 4-year mean was needed because the 3.0 km HYCOM simulations are nondeterministic and flow instabilities are a major contribution at this resolution.

4.1 Mixed Layer Flux

In this subsection the basin–wide mean heat flux and mixed layer flux over the Black Sea (Figure 4.1) is investigated using the model simulations introduced earlier (see Table 3.4). The mixed layer flux is the heat flux applied to the mixed layer in the model simulations. The difference between the basin–wide mean heat flux at the sea surface and mixed layer flux is shortwave radiation absorbed below the mixed layer. Both expt 2 (ECMWF wind/thermal forcing) and expt 5 (NOGAPS wind/thermal forcing) assume all radiation at the sea surface. Thus, basin wide mean heat flux at the sea surface is equal to the mixed layer flux, so they are not shown.

As expected, basin–wide mean heat flux is zero for each experiment (see Figure 4.1) because the Black Sea has nearly closed boundaries which facilitates zero heat flux at the sea surface. In any case, a closed basin Black Sea model must have a zero annual net heat flux. An initial non–zero annual net heat flux will eventually disappear because of SST adjustments. Although basin–wide mean heat flux is zero for each simulation, the heat flux variation throughout the year is different for each simulation because the model calculated sensible and latent heat fluxes, which use model SST, are different.

Figure 4.2 shows the shortwave radiation absorbed below the mixed layer for each simulation. Overall, the shortwave radiation absorbed below the mixed layer is much larger in expt 3 than expt 1, and similarly, it is also larger in expt 6 than expt 4. Thus, the use of the clear water constant attenuation depth clearly results in less penetration flux as opposed to the simulations which use
Figure 4.1. Climatological basin–wide mean heat flux and mixed layer flux in the Black Sea: (a) expt 1 which uses ECMWF wind/thermal fluxes and space/time varying $k_{\text{PAR}}$, (b) expt 2 which uses ECMWF wind/thermal forcing and a clear water constant $k_{\text{PAR}}$ value of 0.06 m$^{-1}$, (c) expt 4 which uses NOGAPS wind/thermal fluxes and spatial $k_{\text{PAR}}$, and (d) expt 6 which uses ECMWF wind/thermal forcing and clear water constant $k_{\text{PAR}}$ value of 0.06 m$^{-1}$. 
Figure 4.2. Climatological basin–wide mean flux difference between the net heat flux at the sea surface and the heat flux applied to the mixed layer in the model simulations (i.e., mixed layer flux) for the experiments shown in Figure 4.1: (a) expts 1 and 3 when HYCOM was forced with ECMWF wind/thermal fluxes, and (b) expts 4 and 6 when HYCOM was forced with ECMWF wind/thermal fluxes. The difference between the net heat flux and mixed layer flux for expts 2 and 4 are zero because both simulations assume all radiation at the sea surface so they are not shown in the figure.

SeaWiFS–based spatial and temporal varying attenuation depths regardless of the atmospheric forcing product used in the simulations.

There is usually heat loss from October through mid–March in all simulations. The mixed layer flux is almost identical to the mean heat flux at the sea surface during these two time periods. On the contrary, there is usually a gain starting from mid–March through October, and the net heat flux reaches its maximum in summer which is due mostly to the shortwave radiation being large (not shown here). The amount of mixed layer flux into the mixed layer is clearly different when using spatial and temporal varying attenuation depths (expt 1) in comparison to the clear water.
constant attenuation depth case (expt 3) which uses the same ECMWF wind/thermal forcing. In the latter, there is less mixed layer flux in summer. Similarly, a comparison of expt 4 and expt 6 also reveals the same type of result.

Overall, the basin–wide difference between the net heat flux at the sea surface and mixed layer flux (i.e., shortwave radiation absorbed below the mixed layer) can be as large 50 W m$^{-2}$ during summer months. In fact, the basin–averaged mean difference for expts 1 and 3 calculated January through December are 17 and 41 W m$^{-2}$, and the basin–averaged mean difference for expts 4 and 6 are 19 and 48 W m$^{-2}$, respectively.

### 4.2 Annual Mean Analysis

The surface circulation in the Black Sea is dominated by meanders, eddies and dipole structures (e.g., Oguz and Besiktepe 1999). This is evident from surface current and SSH snapshots from the 3.0 km HYCOM simulations (Figure 4.3). Given that the Rossby radius of deformation in the Black Sea is $\approx 20–25$ km, it is clear that the radii of coastal eddies in the model range from one to a few times the radius of deformation and are especially prevalent in the eastern half of the Black Sea. This is somewhat consistent with the results reported from a 1/12° ($\approx 9.0$ km) $z$–level OGCM (Staneva et al. 2001) which was forced with monthly mean atmospheric parameters (Sorkina 1974; Altman and Kumish 1986).

When using different atmospheric forcing products, the location and number of eddies in the simulations are quite different in some regions, especially in the easternmost of the region, In the case of ECMWF wind/thermal forcing (expt 1) there are almost no eddies present near the Sea of Azov; while, two small eddies exist in the case of NOGAPS wind/thermal forcing (expt 4). In addition to the differences in the sizes and locations of the small eddies, the SSH values for expts 1 and 4 are also different by up to $\approx 8$ cm in some regions, especially near the northeastern coast.

The HYCOM simulations do not show significant changes in the surface circulation and SSH when all radiation is absorbed in the surface layer (expts 2 and 5) rather than using the solar radiation penetration scheme involving realistic attenuation depths (expts 1 and 4). This is because the Black Sea is very turbid. On the other hand, the clear water constant attenuation depth assumption (i.e., $k_{PAR}= 0.06$ m$^{-1}$ $\approx 17$ m) in expts 3 and 6 shows very different surface circulation structure and SSH variability, mostly east of 39°E where attenuation coefficients (depths) are small (large) in comparison to other parts of the region (see Figure 2.5).
Figure 4.3. Snapshots of sea surface currents (cm s⁻¹) overlain on sea surface height (cm) in the Black Sea on March 2: Left panels from top to bottom, (a), (b) and (c), are for expts 1, 2 and 3 when HYCOM was forced with ECMWF wind and thermal fluxes. Similarly, right panels from top to bottom, (d), (e) and (f), are for expts 4, 5 and 6 when HYCOM was forced with NOGAPS wind and thermal fluxes. Note that the length of the reference velocity vector is 15 cm s⁻¹. See chapter 3 in the text for construction of the wind and thermal forcing fields.
Annual mean surface currents and SSH values for expts 1 through 6 are shown in Figure 4.4. The clear water constant attenuation depth assumption (expts 3 and 6) results in different current structure in comparison to the standard cases (expts 1 and 4) in the eastern Black Sea. The SSH variability in the eastern gyre and off Sinop are different. In general, as indicated in Ozsoy and Unluata (1997), the coastal eddies which have time scales of only a few days (see Figure 4.3) usually merge, causing the larger eddies to form on longer time scales as this is evident from the HYCOM annual mean surface current fields. There is a permanent current system encircling the Black Sea basin cyclonically over the continental slope zone. It is accompanied by a series of persistent anticyclonic mesoscale eddies as well as transient waves with mesoscale eddies propagating cyclonically around the basin (Figure 4.4a,d).

The large scale cyclonic circulation in the Black Sea is driven by the wind stress curl. This is a well-known feature of the Black Sea as previously noted by other OGCM studies (e.g., Stanev and Beckers 1999; Kourafalou and Stanev 2001). However, it is also clear that this feature is largely independent of the atmospheric forcing product choice. These results are not entirely consistent with previous OGCM studies. For example, Stanev and Beckers (1999) used a three-dimensional primitive equation GeoHydrodynamics Environment Research (GER) 3D model with a resolution of 15.0 km x 12.0 km. They reported the lowest SSH values in the eastern cyclonic gyre, indicating substantial differences in the wind stress since this pattern is controlled by the wind stress curl. The 3.0 km HYCOM with a finer resolution than the GHER 3D shows the realistic lowest values in the western cyclonic gyre for both ECMWF and NOGAPS forcing cases (expts 1 and 4, respectively). These results also indicate importance of using wind/thermal forcing from operational models.

Another interesting feature of the surface currents shown in all HYCOM simulations is a well organized anticyclonic current structure (see Figure 4.4) between 34°E and 36°E south of 43°N (also known as Sinop eddies) in the eastern Black Sea, which is missing from the annual mean surface current map of Stanev and Beckers (1999). Previously, analyzing available hydrographic data, Oguz et al. (1993) showed that the Sinop anticyclonic eddy emerged consistently as dominant features of the Black Sea circulation. This is consistent with the result obtained from HYCOM.

It should be noted that based on results from the global NLOM with an embedded mixed layer (Kara et al. 2003a), it was found that the standard spatial and temporal varying $k_{\text{PAR}}$ simulation was closer to the clear water constant attenuation depth case ($k_{\text{PAR}}=0.06 \text{ m}^{-1}$) because the global ocean is not very turbid on average and because globally NLOM's mixed layer was relatively deep (including a 10 m minimum). However, in some regions global NLOM results were reasonable with
Figure 4.4. The annual mean sea surface currents (cm s\(^{-1}\)) overlain on annual mean sea surface heights (cm) in the Black Sea: Left panels from top to bottom, (a), (b) and (c), are for expts 1, 2 and 3 when HYCOM was forced with ECMWF wind and thermal fluxes. Similarly, right panels from top to bottom, (d), (e) and (f), are for expts 4, 5 and 6 when HYCOM was forced with NOGAPS wind and thermal fluxes. Note that the length of the reference velocity vector is 15 cm s\(^{-1}\). The annual mean was formed using model years 5 through 8, and each field has basin wide mean SSH anomaly of zero.
all the radiation absorbed in the mixed layer. In the Black Sea, the standard experiments which use space/time variation in attenuation depths (expts 1 and 4) are much closer to the experiments which assume all radiation absorbed at the surface (expts 2 and 5, respectively) because of its high turbidity. Thus, an OGCM will need to use a spatially varying turbidity.

A zonal temperature cross-section analysis is performed along 42.62°N (Figure 4.5) to explain the differences between expts 1, 2 and 3 (also expts 4, 5 and 6) with respect to stratification. This section was chosen because it crosses major current systems including the mean eddy off Sinop (see Figure 4.4). The main purpose is to see how the subsurface heating affects annula mean stratification and surface circulation. Figure 4.5 shows dramatically the importance of turbidity in stratification, MLD and SST. Absorbing all radiation at the surface (expts 2 and 5) does not cause any significant difference in comparison to the standard cases (expts 1 and 4). This shows only a small effect from realistic mixed layer flux. There is obviously little shallow stratification when using a constant attenuation coefficient value of 0.06 m−1 (see Figure 4.5c,f) as expected. This is because a significant amount of shortwave radiation remains below the sea surface. For example, ≈ 20% is absorbed below 20 m (see Figure 2.1). Thus, using the constant clear water attenuation depth, rather than realistic attenuation depths from SeaWiFS, results in excessive warming below the mixed layer. For example, the depth of the 11.5°C isotherm is ≈ 20 m for expt 1; while, it is ≈ 35 m for expt 3.

In general, using a constant clear water constant attenuation depth assumption in the HYCOM simulations (expts 3 and 6) results in a relatively deep MLD in comparison to the standard simulations (expts 1 and 4) which used realistic attenuation depths (Figure 4.6). The reason is that the warming of the water column below the mixed layer leads to a weakening of the stratification and rapid deepening of the mixed layer. This is evident from the deep MLD between 34°E and 37°E, resulting in disorganized current structure off Sinop especially in expt 6.

It is noted that expts 1 and 2 (similarly expts 4 and 5) produce a MLD that is never deeper than the compensation depth along 42.62°N latitude belt; while, MLD from expt 3 (similarly expt 6) is clearly deeper than the compensation depth off Sinop. Here the compensation depth $D_C = \ln(0.01)/k_{PAR}$ is defined as the depth at which the PAR decreases to 1% of its surface value (e.g., Lalli and Parsons 1997). In the present context $D_C$ represents the maximum depth for solar heating of the upper ocean. When the MLD exceeds the maximum $D_C$ value off Sinop, the thermal energy from solar heating is lost through penetration. It is also noticed that NOGAPS wind/thermal forcing(expt 4) results in stronger currents off Trabzon in comparison to the ECMWF wind/thermal
forcing case (expt 1), showing effects of atmospheric forcing on the model simulations. For example, the shortwave radiation from NOGAPS is significantly larger than the one from ECMWF at around 40–41°E while MLDs are similar to each other.

In comparison to the climatological MLD (see Figure 2.5d), the root–mean–square (RMS) differences are \( \approx 5.7, 5.6 \) and 10.1 m for expts 1, 2 and 3, respectively. This means that using a constant attenuation depth value of 17 m in expt 3 may result in a RMS increase in MLD as large as \( \approx 77\% \) (i.e., from 5.7 to 10.1 m). Similar RMS differences are found for expts 4, 5 and 6 with values of 6.3, 6.3, and 10.9 m, respectively. Thus, it is concluded that using a constant clear
Figure 4.6. The annual mean MLD calculated along 42.62°N latitude in the Black Sea: (a) expts 1, 2 and 3 when the HYCOM was run with ECMWF wind/thermal forcing, and (b) expts 4, 5 and 6 when the HYCOM was run with NOGAPS wind/thermal forcing. The compensation depth as defined in the text is $D_C = \ln(0.01) / k_{PAR}$. 
water attenuation depth value yields unrealistically deep MLDs. Given the fact that there is strong
stratification near the surface (see Figure 4.5a,d), the MLD RMS difference value of 10 m results in
significantly large changes in the subsurface temperatures and thermal stratification (Figure 4.5c
and f).

4.3 Seasonal Analysis

This section examines seasonal variability of sea surface currents and subsurface temperatures
on climatological time scales. Two specific months, February and June, are chosen for analysis
because the difference between the net surface heat flux and mixed layer flux is very small/large
in February/June as demonstrated earlier (see Figure 4.2). All model results are presented based
on monthly means in February and June that are constructed from the last 4 years of the model
simulations.

There are differences in the number and magnitude of the coastal eddies in the eastern part
of the Black Sea in February among all simulations (Figure 4.7). This is especially evident when
using space/time varying attenuation depths as opposed to clear water constant attenuation depth
(i.e., expt 1 vs expt 3 and expt 4 vs expt 6). The use of a clear water constant attenuation depth
results in more eddies. While there are not significant differences in the sea surface currents and
SSH values when comparing standard cases to all radiation at the surface cases (i.e., expts 2 and
5) on the annual time scales, it is clear that there are differences in February, and this is especially
true for the simulations forced with NOGAPS wind and thermal fluxes. For example, the Sinop
eddy in expt 4 is not as strong as the one in expt 5, and there are two small eddies in expt 4.

In comparison to the annual mean of the surface current field (see Figure 4.4), there is a
less well-organized cyclonic eastern gyre in February as evident from the standard simulations.
However, the cyclonic western gyre with large negative SSH values is evident in both standard
simulations. The anticyclonic Trabzon eddy that exists in expt 1 is not seen in expt 2 either.
In general, the eastern and western gyre are evident in standard HYCOM simulations. This is
consistent with the results reported by Truckchew and Demin (1992) who suggested that there
are two organized closed cyclonic circulation cells, one of which covers the interior of the eastern
side, and the other that is located within the southwestern part of the basin during winter.
An observational study by Öğuz and Besiktepe (1999) also confirmed that the Rim current is
identified as a well-defined meandering jet confined to the steepest topographic slope and associated
cyclonic-anticyclonic eddy pairs. It has the form of a highly energetic and unstable flow system.
Figure 4.7. Mean sea surface currents (cm s\(^{-1}\)) overlain on annual mean sea surface heights (cm) in the Black Sea in February: Left panels from top to bottom, (a), (b) and (c), are for expts 1, 2 and 3 when HYCOM was forced with ECMWF wind and thermal fluxes. Similarly, right panels from top to bottom, (d), (e) and (f), are for expts 4, 5 and 6 when HYCOM was forced with NOGAPS wind and thermal fluxes. The climatological February mean was formed using model years of 5 through 8. The length of the reference velocity vector is 15 cm s\(^{-1}\).
Not surprisingly, subsurface temperatures in expts 3 and 6 are quite different in the eastern part (east of 38°E) in comparison to the standard cases in February, expts 1 and and 4, respectively (Figure 4.8). The direct effect of less turbidity would be to put more heat below the mixed layer and that would tend to warm up the mean mixed layer temperature. This is clearly evident from the clear water constant attenuation depth assumption (i.e., \( k_{\text{PAR}} = 0.06 \text{ m}^{-1} \approx 16.7 \text{ m} \)). There might be indirect effects that change this, but most indirect effect arguments also imply warming at depth with less turbidity. For example, subsurface temperatures in expts 3 and 6 are \( \approx 1^\circ \text{C} \) warmer than the ones in expts 1 and 4.

While there were no significant differences between expt 4 and expt 5 on annual time scales, it is clear from west of 34°E that absorption of all radiation at the sea surface may result in cooler temperatures seasonally (e.g., in February). In this case it should be noted that MLD in expt 4 is also deeper than the one in expt 5 west of 34°E (Figure 4.9). Absorption of all radiation at the sea surface (expts 2 and 5) clearly results in deep MLD along 42.6°N latitude belt, and in the case of MLD being the deepest off Sinop, it causes strong sea surface currents around 36°E. This finding is consistent with the one obtained from annual mean analysis. The MLDS in expts 3 and 6 are also deepest at around 36°E, causing similar sea surface current magnitudes off Sinop.

The MLD from all experiments exceeds the \( D_C \), which is calculated as \( D_C = \ln(0.01)/k_{\text{PAR}} \), along 42.62°N latitude belt in February (see Figure 4.9). Since the MLD exceeded the maximum \( D_C \) value at all longitudes, the thermal energy from solar heating that was lost from mixed layer to below the mixed layer is entirely entrained into the mixed layer. Previously, for the annual mean analysis, the \( D_C \) was usually deeper than the MLD at all longitudes, clearly indicating that the Black Sea experiences strong seasonal variability.

The effects of water turbidity on the upper ocean currents is easily seen in June (Figure 4.10) as the \( D_C \) is much deeper than the MLD which is \( \approx 3 \text{ m} \) all over the Black Sea, and shortwave radiation is very large (Figure 4.11). The location and magnitude of the Sinop eddy remains the same regardless of the water turbidity when using ECMWF wind/thermal forcing (expts 1, 2 and 3); while the same is not true for the NOGAPS wind/thermal forcing cases. Absorption of all radiation at the sea surface (expt 5) and the use of clear water constant attenuation depth value of \( \approx 17 \text{ m} \) (expt 6) yields a different current structure off Sinop. There is no evidence of the Trabzon eddy in any of the simulations in June. In general, as in the winter case, the shallow northwestern part of the region has a limited number of eddies in comparison to other regions of the basin.

A significant change in the thermal stratification is evident when using the clear water constant attenuation depth along 42.6°N latitude–belt (Figure 4.12). In this case, there is cooler subsurface
temperatures, which is mostly independent of the atmospheric forcing (expts 1 and 4 vs expts 3 and 6, respectively). Previously, in winter expts 3 and 6 yielded warmer temperatures than expts 1 and 4, respectively (see Figure 4.8). Note that the MLD in June is very shallow along 42.6°N latitude–belt and never exceeds the $D_C$ (Figure 4.13).

When analyzing the annual mean subsurface temperatures along 42.6°N latitude–belt (see Figure 4.5), it was shown that overall, there was net subsurface warming in the clear water constant attenuation depth case in comparison to the standard cases regardless of the atmospheric forcing used in this study (i.e., ECMWF or NOGAPS). However, in the seasonal analysis, one clearly notes...
Figure 4.9. Mean MLD in February calculated along 42.62°N latitude in the Black Sea: (a) expts 1, 2 and 3 when the HYCOM was run with ECMWF wind/thermal forcing, and (b) expts 4, 5 and 6 when the HYCOM was run with NOGAPS wind/thermal forcing. The climatological February mean was formed using model years of 5 through 8. The compensation depth as defined in the text is $D_{C} = \ln(0.01)/k_{\text{PAR}}$. 
Figure 4.10. Mean sea surface currents (cm s\(^{-1}\)) overlaid on annual mean sea surface heights (cm) in the Black Sea in June averaged over 4 years (model years 5 through 8): Left panels from top to bottom, (a), (b) and (c), are for expts 1, 2 and 3 when HYCOM was forced with ECMWF wind and thermal fluxes. Similarly, right panels from top to bottom, (d), (e) and (f), are for expts 4, 5 and 6 when HYCOM was forced with NOGAPS wind and thermal fluxes. The length of the reference velocity vector is 15 cm s\(^{-1}\).
**Figure 4.11.** Climatological mean of various fields over the Black Sea: (a) Ocean mixed layer depth (MLD) in February; (b) compensation depth \(D_C\) in February; (c) MLD in June; (d) \(D_C\) in June; (e) and (f) shortwave radiation from ECMWF and NOGAPS, respectively, in February; (g) (and (h) shortwave radiation from ECMWF and NOGAPS, respectively, in June. Note that (a) through (d) are constructed from GDEM.
Figure 4.12. Cross-section of temperature along 42.62°N from surface to 80 m depth in June. Results from expts 1 through 6 are shown in panels (a) through (f), respectively. The climatological June mean was formed using model years of 5 through 8.

that there is warming in winter and cooling in summer. The net seasonal warming and cooling effects would cancel each other on the annual mean time scales. This can be explained by seasonal and annual mean $D_C$ values along 42.6°N latitude–belt (Figure 4.14). The annual mean $D_C$ is usually between the $D_C$ in February and $D_C$ in June. The MLD always exceed the $D_C$ in February but the MLD never exceeds the $D_C$ in June along 42.6°N latitude–belt. Thus, thermal energy that was lost from mixed layer to below the mixed layer through solar penetration is always gained in February but not in June. On the other hand, MLD in February is much deeper (e.g., > 40 m) than the one in June; although, attenuation depth values differ only by 8 m or so.
Figure 4.13. Mean MLD in June calculated along 42.62°N latitude in the Black Sea: (a) expts 1, 2 and 3 when the HYCOM was run with ECMWF wind/thermal forcing, and (b) expts 4, 5 and 6 when the HYCOM was run with NOGAPS wind/thermal forcing. The climatological June mean was formed using model years of 5 through 8. The compensation depth as defined in the text is $D_C = \ln(0.01)/k_{\text{PAR}}$. Note that the MLD is much shallower than the $D_C$ in June; so $D_C$ values shown are divided by 7 (i.e., $D_C/7$). In other words, the $D_C$ values shown in the figure must be multiplied by 7 to obtain actual values.
Figure 4.14. The climatological mean compensation depth calculated from the SeaWiFS $k_{\text{PAR}}$ values along 42.62°N latitude in the Black Sea in February, June and annual mean as well. This figure is intended to explain differences in mixed layer flux for monthly time scales versus annual mean.

Finally, model–data comparisons are made to examine HYCOM performance in predicting the MLD (Table 4.1). The basin averaged mean MLD difference can be as large as 14.8 m between expt 6 and expt 4. It is interesting that the difference is in the winter, even though the winter MLD is typically deeper than the winter $D_C$. This is expected because MLD is deeper in winter.

Summer MLDs are very shallow (e.g., between 3 and 6 m) all over the Black Sea, so only shown are MLD differences in February (Figure 4.15). The clear water constant attenuation depth assumption (i.e., expts 3 and 6) results in large MLD differences ($> 25$ m) over the Black Sea in comparison to standard simulations (expts 1 and 4).

Close to the coastal regions, the model MLD is very deep. One reason for this is the mismatch in the ECMWF and NOGAPS land–mask used for forcing HYCOM. For example, ECMWF forcing treats some sea points over land, causing problems in representing sea points near land (not shown here). This occurs because the atmospheric fields are on a relatively coarse grid (e.g., 1.125°) but HYCOM is on a much finer grid (e.g., 1/25° here). Hence, any coastal ocean modeler needs to be aware of such a possible problem when using any atmospheric product. From examination (not
Figure 4.15. Mean MLD difference between various experiments in February. While (a), (b) and (c) show differences when the model was forced with climatological ECMWF thermal/wind fluxes, (d), (e) and (f) show differences when the model was forced with climatological NOGAPS wind/thermal fluxes. The expts 1 and 4 represent the standard simulations which use spatial and temporal varying attenuation of Photosynthetically Available Radiation ($k_{PAR}$). Note that the observed climatological MLD (Clim.) used in (c) and (f) is based on 1/8° GDEM climatology, and it is interpolated onto the model domain to calculate differences at each grid point. All model simulations were performed with no assimilation of any ocean data except relaxation to salinity at the sea surface. The contour interval in the color bar is 2.5 m.
Table 4.1. Basin averaged mean MLD differences between various experiments. Also included are differences between the model MLD obtained from standard simulations which use space/time varying attenuation depths (i.e., expts 1 and 4) and climatological MLD constructed from GDEM temperature and salinities over the Black Sea. Minimum and maximum basin wide difference values seen over the Black Sea are also provided.

<table>
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<tr>
<th>MLD difference</th>
<th>Feb</th>
<th>Apr</th>
<th>Jun</th>
<th>Aug</th>
<th>Oct</th>
<th>Dec</th>
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<tr>
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<td>0.2</td>
<td>0.7</td>
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<td>0.7</td>
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<td>0.7</td>
<td>0.7</td>
<td>4.5</td>
<td>10.4</td>
</tr>
<tr>
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<td>-0.4</td>
<td>-0.8</td>
<td>-1.3</td>
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<tr>
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<td>0.4</td>
<td>0.5</td>
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<td></td>
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shown), it is obvious that the ECMWF masking algorithm favors land. In the interior, the MLDs from the model are usually deeper than the ones from the observed MLD climatology constructed from GDEM. The reason is that in HYCOM, using 0.2°C as the MLD criteria, is slightly deeper than KPP's turbulent boundary layer depth. This means that the KPP is very efficient at removing vertical density shear within its turbulent boundary layer.

In some cases, (e.g., recently shallowed boundary layer) the MLD calculated with 0.2°C is much deeper than KPP's boundary layer depth. However, in a observed climatology (e.g., MLD based on GDEM profiles) there is not a boundary layer with almost no variation in density within the mixed layer because it is an average of many such boundary layers. Thus, using a criterion larger than 0.2°C makes sense. The problem is relating an instantaneous MLD to a climatological MLD (i.e., knowing which temperature jump value across mixed-layer to use in each case). The other
problem is that a large value will put the mixed layer at the bottom at some regions. Thus, a
temperature jump of 0.5° was used across mixed-layer to construct the MLD climatology from the
GDEM temperature and salinity profiles.
CHAPTER 5

CONCLUSIONS

This study presents the results of fine resolution (≈ 3.0 km) HYbrid Coordinate Ocean Model (HYCOM) simulations of the Black Sea. The hybrid coordinate extends the geographic range of applicability of traditional isopycnic coordinate circulation models toward shallow coastal seas and unstratified parts of the ocean.

The purpose of this work is two-fold: (1) to develop a solar radiation penetration scheme which uses attenuation depths from remotely-sensed data, and (2) to set up an ocean model with hybrid coordinates for the Black Sea which uses this solar radiation penetration scheme for examination of the impact of subsurface heating on the upper ocean circulation structure. The solar radiation penetration scheme presented in this dissertation treats attenuation as a continuous quantity and is applicable to any OGCM that has fine vertical resolution near the surface. The global climatological monthly mean $k_{\text{PAR}}$ fields used in the parameterization of the solar radiation penetration are derived from the remotely-sensed Sea-viewing Wide Field-of-view Sensor (SeaWiFS) data during 1997–2001. These monthly fields provide the first complete datasets of subsurface optical fields that can be used for Black Sea model applications related to subsurface heating and bio-optical processes. In particular, these fields have been processed to have smoothly-varying and continuous coverage necessary for use in the Black Sea model applications.

Results are reported from three model experiments using different sets of $k_{\text{PAR}}$ values. The current version of the model does not include assimilation of any ocean data except initialization to climatology and relaxation to sea surface salinity. The HYCOM simulations show that assuming all shortwave radiation absorbed at the surface yields annual mean sea surface currents which are similar to those using space/time varying turbidity fields. It is found that a single Jerlov water class cannot be used for solar penetration in the Black Sea. In other words, this indicates that a basin-scale model which would typically use clear water (Jerlov IA) will not work in the turbid Black Sea in predicting upper ocean quantities, in particular sea surface currents and mixed layer depth (MLD). It is concluded that on climatological annual mean time scales any ocean model
study of existing Black Sea circulation needs to use either all shortwave radiation absorbed at the surface or to use a realistic turbidity via attenuation depths from remotely-sensed data (e.g., SeaWiFS). The results also clearly suggest that if the Black Sea turbidity is entirely or largely, due to biology, a lack of nutrients (or another causes for a loss of biomass) will have a significant effect on the overall circulation of the Black Sea. However, a specific examination of Black Sea biosystem robustness is not made here.

Results presented in this dissertation reveal that the direct effect of the including space/time varying attenuation depths in the Black Sea is the shallowing of the MLD. This occurs because the heat is not deposited below the MLD in contrast to the clear water constant attenuation depth simulation. The deepening of the MLD when using the clear water attenuation depth increases the heat capacity of the upper ocean. The increase in heat capacity is responsible for changes in the surface currents. When using the clear water constant attenuation depth assumption of \(\approx 16.7\) m over the Black Sea, the MLD becomes unrealistically deep. This also changes sea surface current structure, especially in the easternmost of the region (e.g., off Trabzon).

In general, using a constant clear water constant attenuation depth assumption in the HYCOM simulations results in a relatively deep MLD in comparison to the standard simulations which used space/time varying attenuation depths. The reason is that the warming of the water column below the mixed layer leads to a weakening of the stratification and rapid deepening of the mixed layer. This results in disorganized current structure off Sinop. It is also noted that while there are not significant changes on the sea surface circulation between the standard simulations and the ones assuming all radiation absorbed at the sea surface on the annual time scales, seasonal analysis clearly shows differences, indicating the effects of water turbidity on monthly time scales.

This dissertation is confined to examination of upper ocean quantities (sea surface currents, sea surface height and MLD) with respect to subsurface heating on the annual/monthly mean time scales. Further examination will reveal impacts of subsurface heating on sea surface temperature. Finally, it should be noted that the Black Sea model is being developed as part of an ongoing global HYCOM planned for transition to operational use in 2006.
APPENDIX A

STATISTICAL INTERPOLATION

Statistical interpolation (SI) is a minimum variance method that seeks to minimize the expected error in the analysis field (Lorenc 1981; Daley 1991). The method determines an interpolated analysis field value $f_A$ at a location $\vec{r}_i$ from a weighted sum of known observed and background field values $f_O$ and $f_B$, respectively, at other locations ($\vec{r}_j, j = 1, \ldots, n; \vec{r}_j \neq \vec{r}_i$) via

$$f_A(\vec{r}_i) = f_B(\vec{r}_i) + \sum_{j=1}^{n} W_{ij} [f_O(\vec{r}_j) - f_B(\vec{r}_j)],$$

(A.1)

where $W_{ij}$ is a covariance weight vector for the location $\vec{r}_i$ (Daley 1991, equation (4.2.9)). Since there are no in-situ observations for the mean $k_{490}, f_O = 0$. Similarly, $f_B(\vec{r}_i) = 0$ because there is no background value at $\vec{r}_i$. The weight vector is determined by solving the matrix equation

$$\sum_{k=1}^{n} B_{jk} W_{ki} = C_{ji},$$

(A.2)

where

$$B_{jk} = \langle \Delta f_B(\vec{r}_j) \Delta f_B(\vec{r}_k) \rangle$$

(A.3)

$$C_{ji} = \langle \Delta f_B(\vec{r}_j) \Delta f_B(\vec{r}_i) \rangle$$

(A.4)

are the covariances in the field value errors $\Delta f_B$. The $n \times n$ symmetric matrix $B$ is the error covariance matrix among the background field values at the non-interpolated locations ($j, k = 1, \ldots, n$), and the $n$ vector $C$ is the corresponding error covariance vector of the background field values at non-interpolated locations with respect to the interpolated location (Daley 1991, equation (4.2.10)). For lack of information on the root-mean-square errors of the mean $k_{490}$, the covariances are assigned correlation functions of

$$B_{jk} = \exp \left[ -|\vec{r}_j - \vec{r}_k| / r_0 \right]$$

(A.5)

$$C_{ji} = \exp \left[ -|\vec{r}_j - \vec{r}_i| / r_0 \right]$$

(A.6)

with a distance scale of $r_0 = 1^\circ$. The value of $r_0$ was chosen to be consistent with the grid resolution.
The data filling of the SeaWiFS \(k_{490}\) using SI proceeds by searching for data voids along latitude transects. Searches along longitude proceed from west to east, and then from east to west. An interpolating grid box of \(5 \times 5\) centered at the data void is used for the SI in all but the last 2 latitude rows at the poleward boundaries. A \(3 \times 3\) box is used for the second last rows, and a \(3 \times 2\) box for the boundaries. Data voids at non-central locations in the interpolating box are excluded when performing the SI.
REFERENCES


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BIOGRAPHICAL SKETCH

A. Birol Kara

A. Birol Kara was born in Trabzon, Turkey, in 1969. Upon completing his BSc at the Istanbul Technical University, he started the MSc degree at the Department of Meteorology, Florida State University, in 1995. After graduation, he worked with Dr. Harley Hurlburt of the Naval Research Laboratory, Stennis Space Center, during 1998–2001. He returned to the Florida State University to complete his Ph.D. in meteorology in 2001. He has a few scientific journal articles published on various subjects: atmospheric boundary layer, North Atlantic hurricane climatology, air–sea interaction and ocean mixed layer.