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Numerical Modeling of the Ocean Circulation: From Process Studies to Operational Forecasting – The Mediterranean Example

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1. Introduction

The Earth is often referred to as the water planet, although water accounts for only 0.023% of the mass of the planet. Nevertheless, water is found mainly at or near the surface and in the atmosphere and therefore is a very prominent planetary feature when viewed from space. Water as a substance appears in all three physical phases – solid, liquid, and gas. Under the present day climatic conditions, ice is found mainly in the polar regions, at latitudes north of 60°N and south of 60°S. Liquid water is found in the hydrosphere which includes the oceans, marginal seas, lakes, and rivers. The oceans cover nearly 70% of the surface of the Earth, with an average depth of ~ 4000 m. Water vapor, the gaseous phase, appears in the atmosphere and accounts for up to 4% of the mass. The hydrologic cycle describes the continuing transfer of water among these three components. All three forms of water also play important roles in the climate system. Water vapor is the main absorber of infrared radiation and therefore is a major contributor to the greenhouse effect. Clouds and ice are the major factors that determine the albedo of the Earth and therefore are mostly responsible for the reflection of approximately 30% of the incoming solar radiation. The specific heat capacity of water is nearly four times that of air and therefore the oceans serve as a major heat reservoir and regulator of the climate system. Furthermore ocean currents are responsible for more than one third of the heat transport from the equator to the poles and therefore affect the horizontal temperature gradients in the atmosphere which are closely linked to the development of major weather systems on various temporal and spatial scales.

The oceans also serve as a major source of food and natural resources and are important for commerce and transportation. For hundreds and perhaps even thousands of years, mariners intuitively understood some of the salient features of the surface circulation in the most highly traversed parts of the ocean. In 1770 Benjamin Franklin and Timothy Folger published the first map of the Gulf Stream, the major ocean current that flows northward along the east coast of North America and then turns northeastward and flows across the North Atlantic Ocean. The purpose of this map was to help mail ships sailing from Europe to North America to avoid this current and thereby shorten the duration of their trip. Yet despite the interest in and the importance of the oceans, oceanography as a formal science is relatively young, being only slightly more than a century old. In the early years, it was primarily a descriptive science based on sparse and scattered observations. The quantitative aspects of physical and dynamical oceanography saw a major breakthrough with the publication of Henry Stommel's seminal work on the North Atlantic circulation (Stommel, 1948). With a simple mathematical model of the wind driven circulation he was able to elegantly explain the phenomenon of westward intensification (i.e., the formation of strong western boundary currents such as the Gulf Stream) as a result of the meridional variation of the Coriolis force.

The idea of using numerical models to further expand the understanding of the intricacies and complexities of the ocean circulation was introduced nearly twenty years later in the pioneering work of Bryan & Cox (1967). As with Stommel's research, they too investigated the circulation of the North Atlantic Ocean which at the time was the most highly observed ocean basin. The purpose of the model was to solve an initial value problem based on a simplified version of the Navier-Stokes equations. Through their model they were able to study the interaction between the wind driven and the thermohaline components of the circulation. Their work drew heavily from the experience of numerical weather prediction which took nearly thirty years to develop the capability of producing skillful forecasts beginning with Richardson's (1922) original concept but unsuccessful attempt and continuing to Charney et al. (1950) producing the first successful 24 hr forecast. As computational capabilities have increased exponentially over the past thirty years, so too has ocean modeling developed from a tool for simplified and focused process studies to fully operational forecasting systems. In this sense, the distinction between process studies (or simulations) and a forecasting system can be explained as follows. In the former, the goal is to understand the physical basis of the process without regard to reproducing specific details at any particular instant in time. In the latter, attention is focused on being able to produce the most accurate simulation of a particular realization of the flow at a specific time. The development of models for process studies and simulations was a necessary step in the development of forecasting systems. Furthermore, the useful range of a forecast, which is closely related to the limit of predictability, is limited by the chaotic behavior of the fluid flow. One the other hand, longer term simulations for the projection of future climate change is perhaps the most common example today of a process study. In both modern process studies and forecasting systems, the initial focus of model development has been the circulation, but today major progress has been made in developing components for simulating and predicting the fundamental biogeochemical processes of the oceanic ecosystem as well.

The goal of this chapter is to present an overview of modern ocean modeling as a tool for basic research as well as for operational forecasting. Considering the rapid developments and extensive experience of the Mediterranean oceanographic research community from recent years, we will use the Mediterranean as the prototype to explain and demonstrate these capabilities and successes in ocean modeling.

2. The governing equations and basic ocean dynamics

In order to fully appreciate the role and importance of numerical ocean models, it is helpful to first understand some of the basic dynamics of the ocean circulation. Mathematically, investigating the ocean circulation can be considered as solving an initial boundary value problem described by the Navier Stokes equations. These form a set of nonlinear, partial differential equations which describes the motion of any Newtonian fluid. The core of this is the three dimensional equation for the conservation of momentum which is essentially an expression of Newton's second law of motion. The two fundamental forces that must be considered are the pressure gradient force and gravity. For geophysical fluids, rotation of the Earth is also important and therefore Coriolis force must also be added to the equations. To complete the description of the motion equations for mass conservation (continuity) and for the conservation of internal energy must also be added. The latter can be expressed in terms of density or in terms of temperature and salinity. To make these equations more tractable and directly applicable to the ocean circulation, various simplifications and approximations are applied. These simplifications are usually based on a scale analysis of the various terms in the equations. The two most common approximations are: (1) the vertical extent or depth of the fluid layer is much smaller than the horizontal scale of motion, and (2) the Boussinesq approximation in which the density variations are assumed to be small compared to the mean value and are therefore neglected except in the buoyancy term of the equation. As a result of the first approximation, the vertical component of the conservation of momentum can be reduced to a diagnostic equation for hydrostatic balance (i.e., the vertical component of the pressure gradient force exactly balances gravity or the weight of the fluid). The second approximation, which is roughly equivalent to assuming that seawater is incompressible, means that mass continuity can be reduced to a diagnostic equation for the conservation of volume (i.e., three dimensional nondivergence). The final set of the governing equations (usually referred to as the primitive equations) in Cartesian coordinates (x, y, z), includes seven equations as follows:

Horizontal momentum

$$\frac{\partial u}{\partial t} + u\frac{\partial u}{\partial x} + v\frac{\partial u}{\partial y} + w\frac{\partial u}{\partial z} - fv = -\frac{1}{\rho_0}\frac{\partial p}{\partial x} + DIFF(u) \quad (1)$$

$$\frac{\partial v}{\partial t} + u\frac{\partial v}{\partial x} + v\frac{\partial v}{\partial y} + w\frac{\partial v}{\partial z} + fu = -\frac{1}{\rho_0}\frac{\partial p}{\partial y} + DIFF(v) \quad (2)$$

Where u, v, w are the velocity components in the x, y, z directions, t is time, ρ is the density (the subscript 0 indicates the mean value), $f = 2\Omega sin\varphi$ is the Coriolis parameter (Ω is the rotation rate of the Earth and φ is the latitude), p is the pressure, and $DIFF(\psi)$ is the diffusion given by $DIFF(\psi) = \frac{\partial}{\partial x}\left(A_h\frac{\partial\psi}{\partial x}\right) + \frac{\partial}{\partial y}\left(A_h\frac{\partial\psi}{\partial y}\right) + \frac{\partial}{\partial z}\left(A_z\frac{\partial\psi}{\partial z}\right)$, where A_h and A_z are the horizontal and vertical diffusion coefficients, respectively;

Vertical momentum (hydrostatic equation)

$$\frac{\partial p}{\partial z} = -\rho g \quad (3)$$

Where g is gravity;

Mass continuity

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0 \quad (4)$$

Conservation of internal energy (can be written in terms of density or temperature and salinity)

$$\frac{\partial T}{\partial t} + u\frac{\partial T}{\partial x} + v\frac{\partial T}{\partial y} + w\frac{\partial T}{\partial z} = DIFF(T) \quad (5)$$

$$\frac{\partial S}{\partial t} + u\frac{\partial S}{\partial x} + v\frac{\partial S}{\partial y} + w\frac{\partial S}{\partial z} = DIFF(S) \tag{6}$$

Where T and S are the temperature and salinity, respectively;

Equation of state

$$\rho = \rho(S, T, p) \tag{7}$$

Details of the derivation of the governing equations can be found in any text book on geophysical fluid dynamics such as Cushman-Roisin (1994). In order to solve the equations it is necessary to specify appropriate spatial boundary conditions and the top, the bottom and the sides of the domain as well as initial conditions. There is no general formulation of the boundary conditions since they depend upon the particular problem being addressed. Examples of boundary conditions at the top include wind stress for the momentum equations, or heat and mass fluxes for the internal energy equations. The bottom boundary conditions usually consist of frictional drag and no vertical mass flux. Lateral boundary conditions may be as simple as no flow at the coastline or some type of wave radiation condition at an open lateral boundary which allows waves to escape with no reflection (e.g., Orlanski, 1976).

The equations as they appear above describe a wide range of atmospheric and oceanic motions (except sound waves which are filtered out by the Boussinesq approximation). To study particular phenomena or processes, they can be further simplified, usually through additional scale analysis which leads to neglecting other terms. In some cases analytical solutions can be found, but in most cases numerical approaches are necessary. A very powerful and widely used simplification of Eq. (1) and (2) is geostrophic flow in which the local time derivative, the nonlinear advections terms, and diffusion are neglected. The remaining leading order terms, which roughly balance each other, are the Coriolis force and the horizontal pressure gradient force (last term on the left hand side and first term of the right hand side, respectively). The immediate implication is that the currents must flow parallel to the isobars (lines of constant pressure) rather than from high pressure to low pressure zones as in non-rotating fluids. This also means that the currents can be diagnosed directly from the pressure or mass field. The practical importance of this is that it is much easier and cheaper to measure the mass field variables (i.e., temperature, salinity, and pressure) than to measure the motion field (currents). Consequently the vast majority of physical oceanographic measurements consists of the three dimensional distribution of temperature and salinity. Combining the geostrophic equations with the hydrostatic equation allows us to compute the vertical shear of the currents due to horizontal pressure or density gradients. The two main weaknesses of the geostrophic approximation are that it breaks down in tropical areas where the Coriolis force is very weak, and it does not allow for temporal changes.

Another common method for simplifying the equations is to reduce their spatial dimensionality. For example, the primary external forcing of the ocean originates in the atmosphere and is applied from above (winds and heat flux). Consequently, the vertical gradients of the primary dependent variables in Eq. (1), (2), (5), and (6) are much larger than the horizontal gradients. It is therefore quite common to study the importance of this stratification through the use of one-dimensional water column models. A classic example is the study of the wind forced surface boundary layer by Ekman (1905) in which Eqs. (1) and (2) are reduced to steady state equations balancing the Coriolis force with the vertical component of diffusion. The solution is the so called Ekman spiral for the surface layer in which the current magnitude decays with depth and the current vector rotates clockwise with depth in the northern hemisphere. Another example is the investigation of vertical convective mixing and its role in the deepening of the surface mixed layer through the use of the one-dimensional version of Eq. (5) and (6) in which the local time derivative is balanced by the vertical component diffusion (e.g., Martin, 1985).

In contrast to the vertical column models, other processes in which the horizontal variations are important or of interest can be investigated using two dimensional, depth integrated versions of the equations. To study the wind driven gyres in the upper ocean, (Stommel (1948) and Munk (1950) both started with the geostrophic form of Eqs. (1) and (2) with the addition of a frictional drag term as an alternative to horizontal diffusion. They took advantage of the non-divergence of the geostrophic flow to recast the equations as a single equation for vorticity. Through the solutions of the equations they were able to explain the underlying dynamics of the observed circulation in the North Atlantic Ocean. The general anticylonic (clockwise) gyre was driven by the curl of the wind stress (i.e., change in the direction from the easterly Trade Winds in the tropics and subtropics to the Westerlies in the mid-latitudes). The appearance of the intense western boundary current (i.e., the Gulf Stream) was a result of intensification due the accumulation of anticylonic relative vorticity by the wind stress and anticylonic planetary vorticity due to meridional variations of the Coriolis parameter but bounded by the damping effect of friction with the east coast of North America.

The various examples presented above are meant to demonstrate some of the basic and salient features of ocean processes which have been investigated over the past 100 years through the use of various simplified versions of the governing equations for geophysical fluid dynamics. It does not even scratch the surface of the vast body of scientific literature in this rapidly expanding and exciting field of study. An in depth survey of these processes can be found in many of the excellent modern textbooks published in recent years such as Vallis (2006).

3. Numerical ocean modeling

As noted in the introduction, the rapid development of computer technology over the past few decades has encouraged the massive development and advancement of numerical ocean models since the original effort of Bryan & Cox (1967). The main advantage of numerical modeling as compared to simplified process studies is that the numerical models are based on the more complete form of the governing equations presented in the previous section. This allows us to investigate more complex flow regimes and processes than in the past. In fact some of the simplifications such as hydrostatic balance in the governing equations are also being removed in recent models, thereby restoring the full time dependent equation for the vertical component of velocity. This is driven by the interest in and capabilities to investigate and simulate smaller scale processes. A model has the potential to fill in the many gaps left by limited in situ observations, subject of course to the computational and mathematical limitations of any model. One disadvantage of using more and more complex models is that it becomes more difficult to isolate and understand specific dynamical processes and thereby we develop the tendency to use a model as a black box. Even when running the most complex models we must never lose sight of what exactly the model is doing. Successful completion of a simulation does not guarantee proper results.

We must always critically examine the results to be sure the model is doing what it should. With this in mind we present a very brief survey of the most commonly used numerical methods used in ocean modeling today.

The governing equations presented in the previous section form a set of time dependent, hyperbolic partial differential equations. Numerical methods for solving such equations have been developed and have appeared in the mathematical literature over many years. As noted in the introduction, the approach to constructing numerical ocean models and the choice of particular methods has benefited greatly from and closely followed the development of atmospheric models and numerical weather prediction, which preceded ocean modeling by 10-15 years. The most common method used today in ocean models is finite differencing. In recent years finite elements have also become popular. For various reasons, spectral methods have not been widely used, perhaps due to the associated difficulties of dealing with irregular boundaries (i.e., coastlines). Without loss of generality, we will use the finite difference method to illustrate the fundamental principles of a typical approach to ocean modeling. A detailed presentation of ocean modeling methodology and applications can be found in the recent books of Kampf (2009, 2010).

The first step in developing a model is to define the domain of interest and divide it into a discrete set of grid points in space. The goal of the model is to approximate the continuous equations with a compatible set of algebraic equations which are solved at the grid points. This requires accurate methods for representing the first and second order spatial derivatives based on the values of the relevant variables at the grid points. If we consider a dependent variable y as a function of the spatial coordinate x, say $y(x)$, then the gradient or first derivative $\partial y/\partial x$ can be approximated from a Taylor series expansion around the i-th grid point x_i as

$$\frac{\partial y}{\partial x} \approx \frac{y_{i+1}-y_i}{\Delta x} + O(\Delta x) \tag{8}$$

or

$$\frac{\partial y}{\partial x} \approx \frac{y_i-y_{i-1}}{\Delta x} + O(\Delta x) \tag{9}$$

which are referred to as the forward and backward differencing schemes, respectively, and where Δx is the grid spacing. These schemes are first order accurate as indicated by the truncation error $O(\Delta x)$. By averaging these two schemes we obtain the more accurate centered differencing scheme

$$\frac{\partial y}{\partial x} \approx \frac{y_{i+1}-y_{i-1}}{2\Delta x} + O(\Delta x^2) \tag{10}$$

which is second order accurate. Higher order schemes that are even more accurate can be formed by various weighted combinations of the respective Taylor series expansions, although most models use centered differencing. The second derivative is approximated by the three point stencil

$$\frac{\partial^2 y}{\partial x^2} \approx \frac{y_{i+1}-2y_i+y_{i-1}}{\Delta x^2} \qquad (11).$$

The location of the dependent variables is a matter of choice. They can all be co-located at the grid points, or can be located on a staggered grid in which certain variables are shifted by half of a grid point. A commonly staggered grid is the Arakawa C-grid in which the velocity components are shifted one half of a grid point in their respective directions relative to the mass variables (Arakawa & Lamb, 1977). This arrangement ensures that the numerical equations will preserve certain integral properties of the continuous equations such as energy conservation.

In the time integration of the equations, the computational stability of the differencing scheme must also be considered. This means that the time step and spatial grid spacing must be chosen in such a way as to properly resolve the motion of the fastest moving waves that can be simulated by the model (usually the free surface or external mode gravity wave). Most models use some type of centered explicit or split explicit scheme that is also second order accurate. In explicit schemes all variables can be advanced to the forward time step based on the values known at the present and/or backward time step. In split explicit schemes, the terms in the momentum equations that are identified with the fastest moving waves are integrated separately with a shorter time step. This is done to improve the numerical efficiency and execution time of the model. An alternative is to use a fully implicit time scheme in which advancing the model to the forward step involves simultaneously knowing the values of the variables at the forward, present, and backward time steps. This has the advantage of the scheme being absolutely stable (i.e., no numerical amplification) but the disadvantage of being computationally cumbersome and slow due to the need to invert a large tridiagonal matrix at every time step.

The final point to consider in the construction of an ocean model is how to account for the unresolved scales of motion. The grid spacing or resolution of a model limits the explicitly resolved processes. Strictly speaking, from a mathematical perspective the shortest length scale that can be explicitly resolved by a model is $2\Delta x$. Practically however, scales shorter than ~4-6 Δx are often misrepresented due to numerical damping or phase speed errors which may be an inherent characteristic of certain differencing schemes. However there are also important processes which may occur on scales smaller than the grid spacing which affect the larger scale flow. The classic example of this is vertical convective mixing forced by the wind or by induced by static instability when the water is cooled from above. This mixing will transport various properties of the water but is accomplished by small scale turbulent eddies which are not usually explicitly resolved. Such processes are accounted for by adding sub grid scale parameterizations, often in the form of a diffusion term but with a diffusion coefficient that is several orders of magnitude larger than the value for molecular diffusion. The quasi-empirical method for computing these eddy diffusion coefficients is usually referred to as a turbulence close scheme (e.g., Mellor & Yamada, 1982; Pacanowski & Philander, 1981). An analogous term is usually included for horizontal mixing. Finally, ecosystem models, which are becoming more common components of ocean models, require the addition of advection-diffusion equations, similar to Eqs. (5) and (6), for the relevant biogeochemical variables in addition to all of the biogeochemical processes which are treated computationally the same as sub grid scale processes.

4. The Mediterranean Sea – a laboratory ocean basin

Since the early 20th century the Mediterranean Sea was known to be a concentration basin where excess evaporation drives a basin wide thermohaline cell in which less saline water enters from the Atlantic Ocean through the Strait of Gibraltar (Nielsen, 1912). This surface water becomes more saline and denser. It sinks to a depth of ~ 250 m and then returns to the strait where it is carried by a subsurface outflow back to the Atlantic. During the past 25-30

years scientific interest in the oceanography of the Mediterranean Sea was renewed for various reasons. As a result of intensive field campaigns, it became clear that the circulation is far more complex than originally envisioned. It is now known that the Mediterranean Sea functions as a mini-ocean with dynamical processes occurring over a broad spectrum of spatial and temporal scales ranging from the basin wide thermohaline cell, driven by deep water formation, with a time scale of tens of years to energetic mesoscale eddies varying over a period of several weeks to months (e.g., Millot, 1999; Robinson & Golnaraghi, 1994).
Following the new description of the circulation that emerged from these programs, various numerical models were applied to the Mediterranean to further investigate the processes that drive the circulation. Initially, low resolution, basin wide models were used to study the climatological mean circulation of the entire Mediterranean (e.g., Roussenov et al., 1995; Zavatarelli & Mellor, 1995). Other models focused on particular process studies such as deep water formation (e.g., Wu et al., 2000) and/or the sub-basin circulation, and were used to study the response of the general circulation to interannual atmospheric variability (e.g., Korres et al., 2000). Most recently, a rather unique and fascinating phenomenon that occurred in the Eastern Mediterranean involved an abrupt shift in the source region of deep water formation from the Adriatic Sea to the Aegean Sea during the 1990's. This has been called the Eastern Mediterranean Transient (Roether et al., 2007). Several models have been used to simulate the evolution of this process (e.g., Lascaratos et al., 1999; Samuel et al., 1999) in response to changes in atmospheric forcing. As the data and research models provided new understanding of the circulation, and as observational systems and computer technology advanced, by the late 1990’s it was decided to apply this new knowledge to the problem of operational ocean forecasting. An up to date review of the present understanding of the Mediterranean circulation can be found in Brenner (2011).
In the next two sections we will present some examples of both process studies and ocean forecasting taken from some of our most recent research efforts. This represents only a small fraction of many of the ongoing investigations being conducted by many scientists around the Mediterranean. In no way is this intended to be an exhaustive survey. It is simply a small sample meant to demonstrate the state-of-the-art of applications of numerical ocean models. It is mainly out of convenience that we take examples from our own personal experience of research in the Mediterranean.

5. Process studies and simulations

In this section we present an example from some of our recent and ongoing Mediterranean modeling research. It is a one dimensional (vertical) coupled hydrodynamics-ecosystem model for a typical point located in the Eastern Mediterranean Sea. The goal of this model is to investigate the fundamental biogeochemical processes and the influence of the annual cycle of vertical mixing upon them. The hydrodynamic part of the model is a one dimensional version of the Princeton Ocean Model (POM) originally described by (Blumberg & Mellor, 1987). POM is a three dimensional, time dependent model based on the primitive equations with the Boussinesq and hydrostatic approximations as described above in Section 2. It also included a free surface, which turns the continuity equation, Eq. (4), into a time dependent equation for the height of the free surface. POM contains full thermodynamics as well as the turbulence closure sub-model of (Mellor & Yamada, 1982). It is forced at the surface through the boundary conditions which specify the wind stress, heat flux components, and fresh water flux. In the vertical column version all horizontal advection and diffusion are neglected and the focus is on the role of vertical mixing only. Complex ecosystem or biogeochemical models are young relative to hydrodynamic models and are therefore in a stage of rapid development. For this particular study we have used the Biogeochemical Flux Model (BFM) described by (Triantafyllou et al., 2003 and Vichi et al., 2003). The model simulates several classes of phytoplankton, zooplankton, the carbon cycle, and the nitrogen cycle. The specific coupling of the models and implementation for the southeastern Mediterranean Sea presented here is based on the work of Suari (2011). In terms of the hydrodynamics, the main challenge in running the one dimensional model for the eastern Mediterranean was to account for the inflow of relatively fresh Atlantic Water which prevents unrealistic increases in salinity, which would cause the model to eventually become unstable. This was solved by adding a relaxation term in which the simulated salinity profile was nudged towards monthly mean climatological profiles. The model was configured with 40 unevenly spaced layer from the surface to a depth of 600 m and was forced at the surface with a repeating annual cycle that consisted of daily mean winds and heat fluxes that were computed from the multiyear average of the data taken from the NCEP/NCAR reanalysis covering the period from 1950-2006 (Kalnay et al., 1996). The model was run for 50 years with this perpetual year forcing. The purpose of such experiments is to assess the long term behavior and stability of the system without regard to the high frequency or inter annual variability.

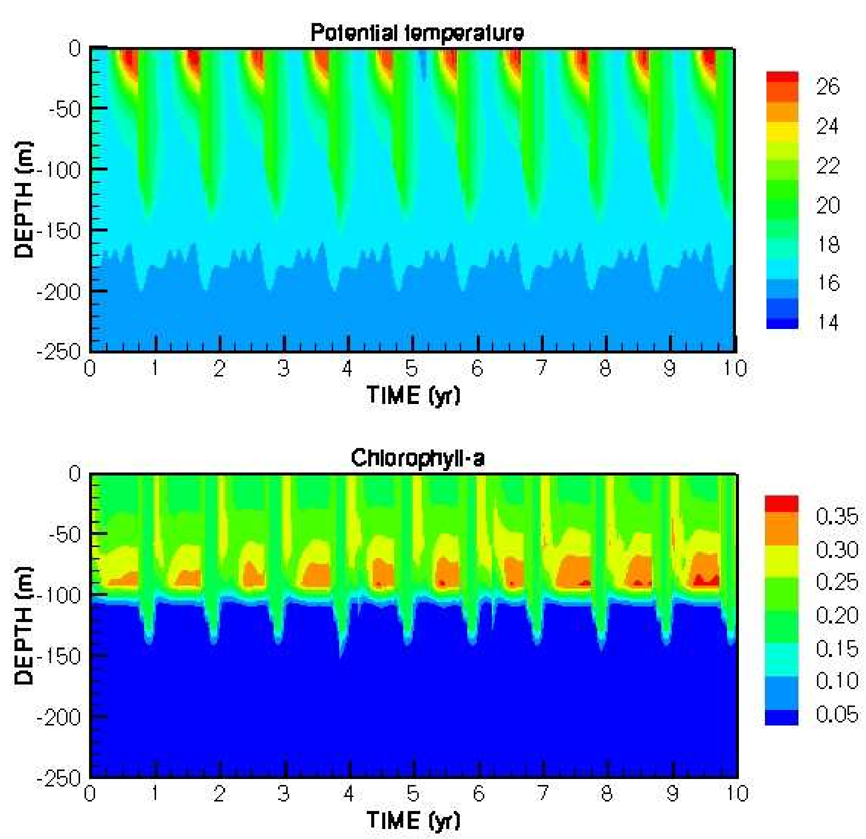

Fig. 1. Time series plot of: potential temperature (°C) in the upper panel and chlorophyll-a (μg L^{-1}) in the lower panel from the last 10 years of a 50 year simulation of a one dimensional coupled hydrodynamic-ecosystem model for the eastern Mediterranean.

The results presented in Fig. 1 show the potential temperature (upper panel) and the chlorophyll-a (lower panel) from the last ten years of the fifty year simulation. By this point the model has passed through the spin-up phase and produces a relatively stable repeating annual cycle. The surface temperature varies between a maximum of approximately 26°C in summer and a winter minimum of 16.6°C. The shallow surface mixed layer, in which the temperature is relatively high and uniform, is clearly visible in summer when it extends from the surface to a depth of 30 m. It is driven mainly by wind mixing, which generates enough turbulent kinetic energy to mix the water against the density gradient. By autumn the surface begins to cool and as a result the water column begins to mix vertically due to free convection, as indicated by the deepening of the green shaded contour. The free convection is driven by gravitational instability of the water column due to cooling from above. By late winter (early to mid March) the mixed layer has deepened to it maximum extent of 190 m as indicated by the uniform cyan contour extending from the surface from late February through mid March. This cycle and the values of the temperature and mixed layer depth are consistent with the observations from this region (e.g., Hecht, et al., 1988; Manca et al., 2004; Ozsoy et al., 1993).

The lower panel of Fig. 1 shows that chlorophyll-a (concentration in $\mu M\ L^{-1}$), which is the proxy for phytoplankton biomass, is confined to the upper part of the water column where there is sufficient light for photosynthesis. Nutrients (mainly nitrate, phosphate, and silicate) are also necessary for the cells to function. The nutrients are injected into the upper layers from below the nutricline (begins at ~150-200 m and extends to ~ 600 m) during deep winter mixing or during wind induced upwelling events. They are rapidly depleted form the photic zone when photosynthesis commences. Chlorophyll-a in the figure exhibits a pattern that is typical for an oligotrophic sea such as the eastern Mediterranean. During spring and summer it is confined mainly to the upper 90-100 m. During the deep mixing in the latter part of winter, the phytoplankton are transported deeper by convective mixing. A combination of factors leads to reduced photosynthesis and biomass concentration during this period. Sun light is less intense and the phytoplankton spend less time in the photic zone. Also due to the deeper mixing they are distributed over a larger volume and therefore the concentration is lower as indicated by the cyan contour. The warmer colors indicate two important features on the marine ecosystem. In early spring the yellow contours show a layer of relatively high chlorophyll concentration extending from the surface to a depth of ~80 m and which lasts for 2-3 weeks. This phenomenon is referred to as the spring bloom. It occurs shortly after the end of the winter (i.e., end of net surface cooling) and the onset of net surface heating in the spring. As a result the free convective mixing ceases and the phytoplankton remain in the upper layers. At this time nutrients are abundant due to the import of high nutrient waters from the deeper layers during winter. These two factors combined with the increasing intensity of the sunlight lead to a rapid increase in photosynthesis and therefore a substantial increase in chlorophyll-a concentration. The nutrients are consumed by the photosynthetic activity of the cells. Since the nutrient source in deep water has been cut off by the cessation of free convection, the nutrients in the photic zone are rapidly depleted and the bloom ends within a few weeks. This is indicated by the transition to the green contours. Later in the summer a subsurface layer with high chlorophyll-a concentration appears at a depth of 70-90 m (yellow and orange contours). This phenomenon referred to as the deep chlorophyll maximum, DCM, is due to the complex interaction between light intensity, leakage of nutrients from the nutricline, and the density stratification. Its occurrence is quite common in the oligotrophic Mediterranean Sea (e.g., Estrada et al., 1993; Yacobi et al., 1995). The simulated pattern and values of chlorophyll-a concentration are consistent with observed values for this region reported by (Manca et al., 2004 and Yacobi et al., 1995).

6. Ocean forecasting

In this section we present another example of the powerful use and application of numerical ocean models as part of an operational forecasting system. In contrast to process studies or simulations which are designed to help us understand the particular dynamical process of interest, the goal of a forecasting system is to provide the most accurate prediction of the circulation at a particular instant in time, but within the constraint of producing the forecast in reasonably short period of time so that it considered to be useful. Clearly a 24 hour forecast that requires 24 hours of computer time has no value. A balance must therefore be reached between the acceptable level of forecast error and the time required to produce the forecast. Furthermore in a forecast system, in addition to the model itself, the specification of the initial conditions is a central consideration. Experience from numerical weather prediction has shown that during the first few days the forecast errors depend mainly on errors in the initial conditions, whereas at longer forecast lead times model errors and uncertainties have a larger impact on forecast errors. In addition to collecting data, accurate mathematical methods are necessary for interpolating the observations to the model grid while creating a minimal amount of numerical noise. This entire procedure, referred to as data assimilation (e.g., Kalnay, 2003), will not be discussed here. Our focus will be on the numerical model itself.

The development of the Mediterranean Forecasting System, MFS, began in 1998 as a cooperative effort of nearly 30 institutions with the goal of producing a prototype operational forecasting system and to demonstrate its feasibility. The project included components of in situ and remotely sensed data collection, data assimilation and model development. The model development component was structured to include a hierarchy of nested models with increasing resolution. The overall system was driven by the coarse resolution, full Mediterranean model. At the next level, sub-basin scale models, which covered large sections of the western, central, and eastern Mediterranean with a threefold increase in resolution, were nested in the full basin model. Nesting is the procedure through which the initial conditions were interpolated to the higher resolution grid, and the time dependent lateral boundary conditions were extracted from the coarser grid model. Finally, very high resolution local models for specific regions were nested in the sub-basin models with an additional two to threefold increase in resolution. An overall description of the prototype system and its implementation can be found in (Pinardi et al., 2003). While the initial model development focused on mainly climatological simulations with the nested model, the next phase led to the pre-operational implementation of short term forecasting with all three levels of models. This system has evolved into Mediterranean Operational Ocean Network, which is perhaps one of the most advanced operational ocean forecasting systems today (MOON, 2011). It routinely provides daily forecasts for the circulation at all scales and the ecosystem at the larger scales.

One component of MOON is a high resolution local model for the southeastern continental shelf zone of the eastern Mediterranean. The model was developed initially within MFS (Brenner, 2003) and has subsequently gone through a number of improvements and refinements. The version presented here is described in detail by (Brenner et al., 2007). It is based on the full three dimensional, primitive equations Princeton Ocean Model which was described above in Section 3. The horizontal grid spacing is 1.25 km and there are 30 vertical levels distributed on a terrain-following vertical coordinate. Data for the lateral boundary conditions are extracted from a sub-basin, regional model which covers most of the Levantine, Ionian, and Aegean basins. The domain and bathymetry of the model are shown in Fig. 2. The mathematical formulation of the boundary conditions along the two open boundaries consists of specifying the normal and tangential components of the horizontal velocity at all boundary grid points and the tracers (temperature and salinity) at inflow points. At outflow points the boundary values are extrapolated from the first interior grid point using a linearized advection equation.

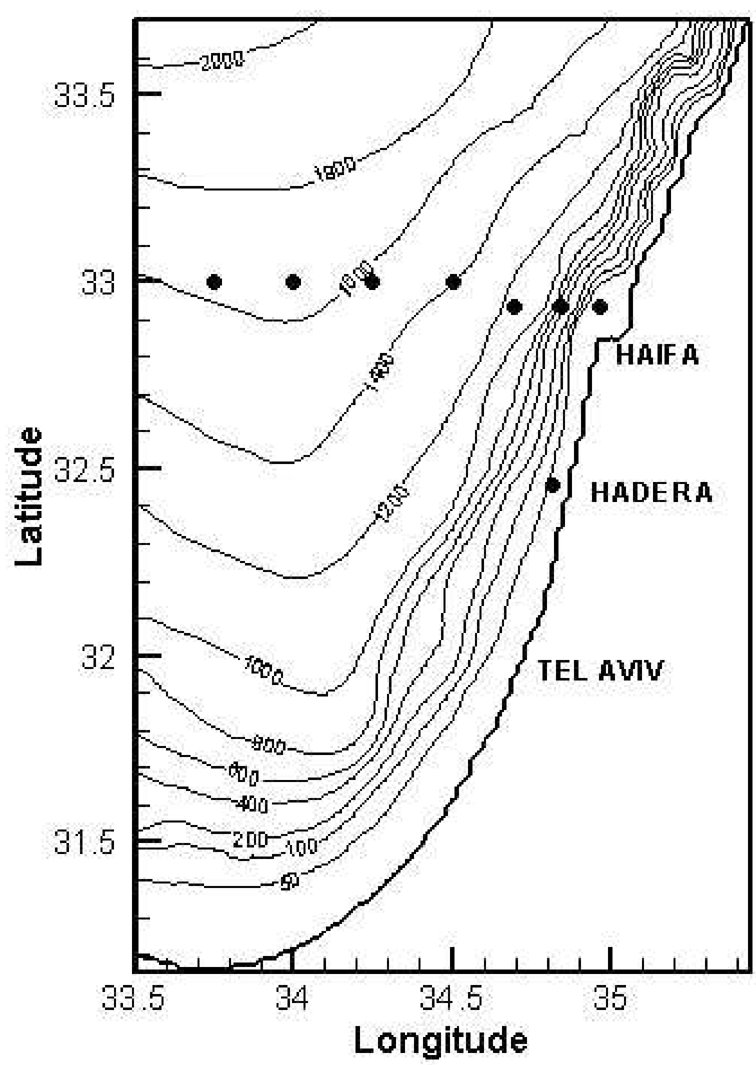

Fig. 2. Domain and bathymetry of the high resolution southeastern Levantine model. Dots indicate locations were observations were available for verification.

This model runs daily and produces forecasts of the temperature, salinity, free surface, and currents out to four days. As noted previously, the primary goal of a forecasting system is to produce the best possible prediction of the circulation at a specific instant in time. Thus forecast verification is an important aspect of assessing the usefulness of the system. In comparison to the atmosphere, ocean observations are extremely limited. The best spatial and temporal data are provided by satellites but are generally limited to sea surface temperature (SST) and sea surface height. The former are usually available several times daily while the latter are limited to approximately weekly, depending upon the path of the satellite. Other measurements are available from ships of opportunity or from fixed buoys but these are sporadic and limited in both time and space. With all of these reservations in mind, in the next few figures we present some examples of the verification of the forecasts produced by this model. In Fig. 3 we show the forecast skill of SST for a one year period as a function of forecast lead time. The skill scores used are the domain averaged root mean square error (RMSE) and the anomaly correlation coefficient. The former measures the magnitude of the forecast error while the latter provides a measure of the pattern error. The figure shows the forecast skill for the high resolution shelf model (red line) and for the coarser resolution regional model (green line). We also include the error for a persistence forecast (i.e., no change from the initial conditions), which is considered to be the minimum skill forecast. From both the RMSE and the anomaly correlation it is clear that the forecast skill degrades as the forecast length increases. The value added to the forecast by the high resolution model is substantial as it outperforms the regional model in both scores (i.e., lower error magnitude and higher pattern correlation). Both models also manage to significantly beat persistence for RMSE, but the regional model is only marginally better in the pattern correlation.

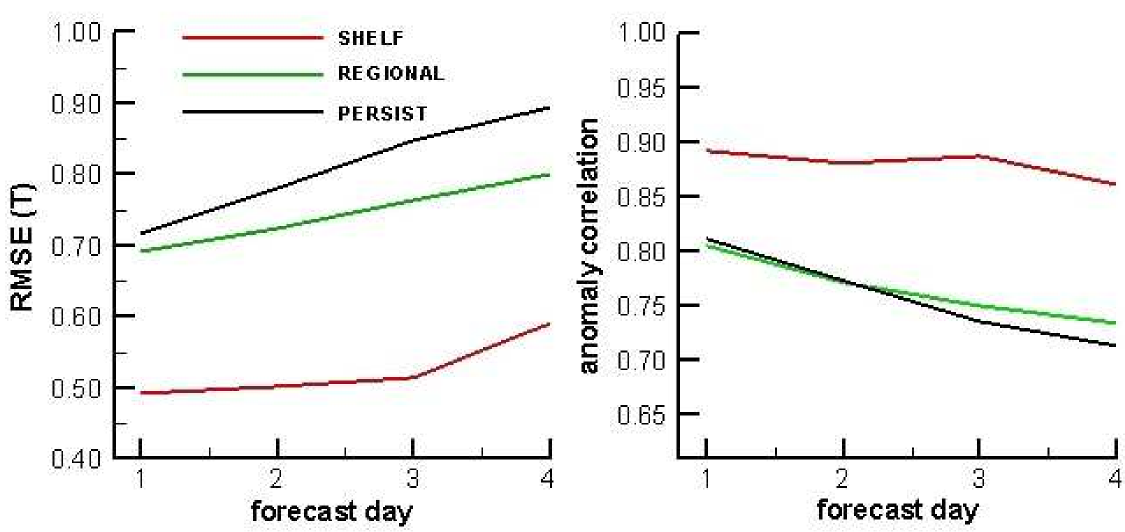

Fig. 3. Forecast skill for one year of forecasts in terms of root mean square error and anomaly correlation coefficient.

While the skill of the SST forecast is impressive, it is also important to validate the ability of the model to predict the subsurface fields. Unfortunately here the data are much more limited in space and time. In Fig. 4 we show a scatter plot of the predicted versus the observed temperature taken from a sea level measurement station located offshore near Hadera (see map in Fig. 2 for location). The instrument was located at a depth of ~15 m below the surface and the bottom depth is ~27 m. The comparison shown here also covers a one year period. Overall the comparison is excellent with a correlation coefficient of nearly 0.97. During winter (low temperatures) and summer (high temperatures) the points tend to be roughly evenly scatter above and below the regression line thus indicating that there is no clear bias in the forecasts. During the transition seasons of spring and autumn (mid range temperatures), there is a strong tendency for the model to under predict the temperature and therefore develop a cold bias. This is most likely due to the more rapid temperature changes during the transitions seasons as compared to summer or winter.

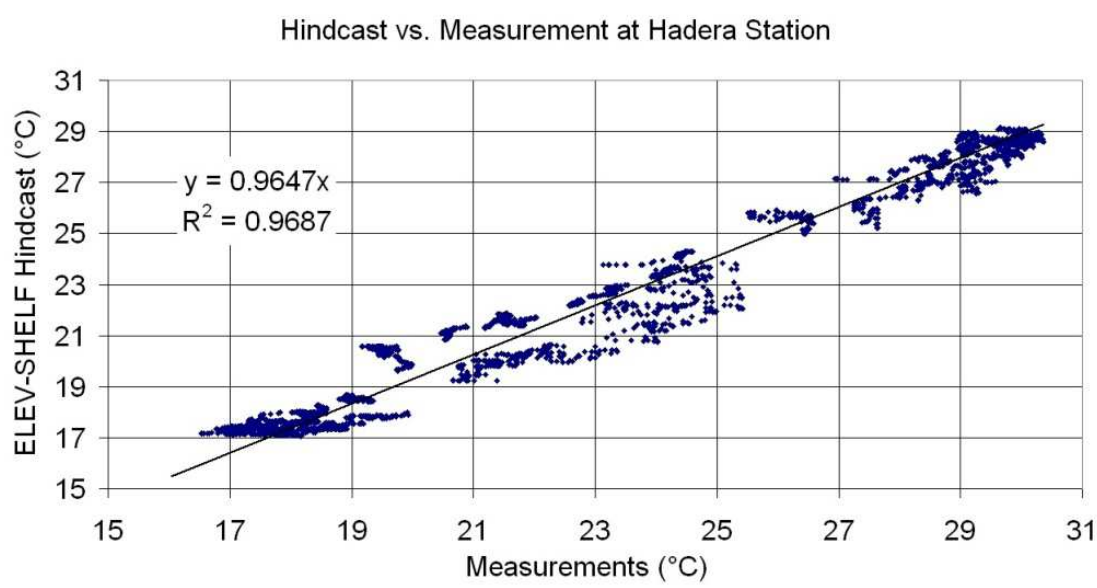

Fig. 4. Scatter plot of the predicted versus measured temperature at a depth of 15 m at an offshore station.

Finally, as a measure of the spatial distribution of the prediction of the subsurface fields, a comparison was made between all measurements collected during a single, one day cruise in the late summer along a transect of points that extend westward from Haifa (see Fig. 2 for location). The measurements were obtained from an instrument that measures nearly continuous profiles of temperature and salinity from the surface to the bottom or to a depth of 1000 m, whichever is deeper. From below the surface mixed layer, the model did an excellent job of predicting the temperature and salinity at all depths and stations along the transect. In the mixed layer the model showed a warm bias with simulated temperatures that were too high by 1-2°C. This error is probably due to the specification of surface heat fluxes that were too high and/or winds that were too weak which prevented the model from creating a deep enough mixed layer. The high resolution forecast was significantly better than the regional model forecast in this area which again demonstrates the value added by a high resolution model. It should be noted however that this comparison was conducted for a single forecast only.

7. Conclusion

In this chapter we have presented a concise overview of more than 40 years of research and development of numerical ocean circulation models. The pioneering work of Bryan & Cox (1967) set the stage for subsequent model development. The rapid development of computer technology of the past two decades has been a major factor allowing for the design of increasingly more complex and realistic models. By complementing field data and the associated gaps, numerical ocean models have proven to be an indispensible tool for enhancing our understanding of a wide range and variety of processes in oceanic hydrodynamics. Consequently, most modern oceanographic studies will almost always include a highly developed modeling component. Models are routinely used for processes studies and as the central component operational ocean forecasting systems as demonstrated by the examples presented.

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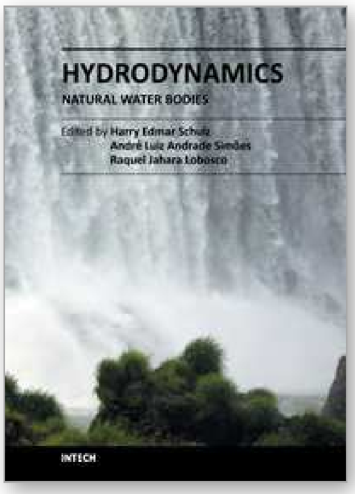

Hydrodynamics - Natural Water Bodies

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The knowledge of the characteristics of the fluids and their ability to transport substances and physical properties is relevant for us. However, the quantification of the movements of fluids is a complex task, and when considering natural flows, occurring in large scales (rivers, lakes, oceans), this complexity is evidenced. This book presents conclusions about different aspects of flows in natural water bodies, such as the evolution of plumes, the transport of sediments, air-water mixtures, among others. It contains thirteen chapters, organized in four sections: Tidal and Wave Dynamics: Rivers, Lakes and Reservoirs, Tidal and Wave Dynamics: Seas and Oceans, Tidal and Wave Dynamics: Estuaries and Bays, and Multiphase Phenomena: Air-Water Flows and Sediments. The chapters present conceptual arguments, experimental and numerical results, showing practical applications of the methods and tools of Hydrodynamics.