THREE-DIMENSIONAL CIRCULATION MODEL DRIVEN BY WIND, DENSITY, AND TIDAL FORCE FOR ECOSYSTEM ANALYSIS OF COASTAL SEAS

1. Introduction

Coastal seas are receiving a great deal of attention due to the increasing use of their resources. Escalating demands for coastal development have directed both governments and industries to investigate the basic mechanisms that govern water circulation in coastal seas. Knowledge of this circulation is useful for environmental management and conservation of marine ecosystems.

The movement of water in a coastal sea is driven by such factors as freshwater discharge from rivers, tidal excitation, intrusion of warm and highly saline ocean water at the mouth of bays, heat transfer, and wind stress at the surface of the water. The ecosystem in Tokyo Bay is strongly influenced by these complex factors. Thus, it is essential to analyze the water circulation there. To simulate this system, we used a 3-dimensional numerical circulation model developed originally by Blumberg and Mellor (1983, 1987), which is able to deal with stratified flows occurring on a time scale of 30 days and spatial scales of 1 to 100 km. The model was extended by Blumberg and Goodrich (1990) to include river segments as a domain of calculation up to the point at which tidal effects are minimal. This is a major difference between the two models, and the latter model is more favorable for simulating circulation in a bay with significant freshwater impact from rivers.

2. Model Description

Here we provide a relatively detailed description of a numerical circulation model called the Princeton Ocean Model (POM; Blumberg and Mellor, 1983). The model belongs to that class of models in which model realism is an important goal and mesoscale phenomena are addressed, that is, activities that are 1 to 100 km long and; 30-day time scales are commonly observed in estuaries and coastal waters (Beardsley and Boicourt, 1981). It is envisioned that the model ultimately will be used as part of a coastal sea-forecasting program. The model is 3-dimensional, incorporating a turbulence closure model to provide realistic parameterization of the vertical mixing processes. The prognostic variables are the three components of velocity, temperature, salinity, turbulence kinetic energy, and macroscale turbulence . The momentum equations are nonlinear and incorporate a variable Coriolis parameter. Prognostic equations governing the thermodynamic quantities, temperature and salinity account for water mass variations brought about by highly time-dependent coastal upwelling processes as well as horizontal advective processes. Free surface elevation is also calculated prognostically, with only some sacrifice in computational time so that tides and storm surge events can also be simulated. This is accomplished by use of a mode-splitting technique whereby the volume transport and vertical velocity shear are solved separately. Other variables include density, vertical eddy viscosity, and vertical eddy diffusivity. The model also accommodates realistic coastline geometry and bottom topography.

The model's performance has been tested in a variety of applications (Blumberg and Mellor, 1979a, b, 1980, 1981a, b, 1983; Blumberg, 1997), including simulation of the tides in Chesapeake Bay, simulation of coastal circulation off Long Island, New York, and a computation of the general circulation in the Middle Atlantic and South Atlantic bights and in the Gulf of Mexico. The grid spacings have ranged from 1 to 50 km in these applications.

2.1 Governing Equations

The equations that form the basis of the circulation model describe the velocity and surface elevation fields, and the salinity and temperature fields. Two simplifying approximations are used (Bryan, 1969); first, it is assumed that the weight of the fluid identically balances the pressure (hydrostatic assumption), and second, density differences are neglected unless the differences are multiplied by gravity (Boussinesq approximation).

Consider a system of orthogonal Cartesian coordinates with x increasing eastward, y increasing northward, and z increasing vertically upwards. The free surface is located at $z = \eta(x, y, t)$ and the bottom is at $z = -H(x,y)$. If U and V are east and northward horizontal velocity components, respectively, and W is the vertical velocity, the continuity equation is

$$\frac{\partial U}{\partial x}+\frac{\partial V}{\partial y}+\frac{\partial W}{\partial z}=0 \tag{1}$$

The Reynolds momentum equations are

$$\frac{\partial U}{\partial t}+\frac{\partial U^2}{\partial x}+\frac{\partial UV}{\partial y}+\frac{\partial UW}{\partial z}-fV=-\frac{1}{\rho_0}\frac{\partial P}{\partial x}+\frac{\partial\left(-\overline{uw}\right)}{\partial z}+F_U \tag{2}$$

$$\frac{\partial V}{\partial t}+\frac{\partial UV}{\partial x}+\frac{\partial V^2}{\partial y}+\frac{\partial VW}{\partial z}+fU=-\frac{1}{\rho_0}\frac{\partial P}{\partial y}+\frac{\partial\left(-\overline{vw}\right)}{\partial z}+F_V \tag{3}$$

$$\rho g=-\frac{\partial P}{\partial z} \tag{4}$$

with ρ_0 the reference density, ρ the in situ density, g the gravitational acceleration, and P the pressure. A latitudinal variation of the Coriolis parameter f is introduced by use of the plane approximation.

The pressure at depth z can be obtained by integrating the vertical component of the equation of motion, (4), from z to the free surface η, and is

$$P(x,y,z,t)=P_{atm}+g\rho_0\eta+g\int_z^0\rho(x,y,z',t)dz' \tag{5}$$

Henceforth, atmospheric pressure, P_{atm}, is assumed constant.
The conservation equations for temperature and salinity may be written as

$$\frac{\partial T}{\partial t}+\frac{\partial UT}{\partial x}+\frac{\partial VT}{\partial y}+\frac{\partial WT}{\partial z}=\frac{\partial\left(-\overline{wT}\right)}{\partial z}+F_T \tag{6}$$

$$\frac{\partial S}{\partial t}+\frac{\partial US}{\partial x}+\frac{\partial VS}{\partial y}+\frac{\partial WS}{\partial z}=\frac{\partial\left(-\overline{ws}\right)}{\partial z}+F_S \tag{7}$$

where T is temperature and S is salinity. Using these two variables, density is computed according to an equation of state of the form

$$\rho=\rho(T,S) \tag{8}$$

given by Fofonoff (1962). The potential density ρ is the density evaluated as a function of potential temperature and salinity but at atmospheric pressure; it provides accurate density information to calculate horizontal baroclinic gradients that enter into the pressure gradient terms, and the vertical stability of the water column, which enters into the turbulence closure model even in deep water when pressure effects become important.

All motion induced by small-scale processes not directly resolved by the model grid (subgrid scale) are parameterized in terms of horizontal mixing processes. The terms F_x, F_y, F_T and F_s found in (2), (3), (6), and (7) represent these unresolved processes and in analogy to molecular diffusion can be written as

$$F_U = \frac{\partial}{\partial x}\left(2A_M \frac{\partial U}{\partial x}\right) + \frac{\partial}{\partial y}\left[A_M\left(\frac{\partial U}{\partial y} + \frac{\partial V}{\partial x}\right)\right] \tag{9a}$$

$$F_V = \frac{\partial}{\partial y}\left(2A_M \frac{\partial V}{\partial y}\right) + \frac{\partial}{\partial x}\left[A_M\left(\frac{\partial U}{\partial y} + \frac{\partial V}{\partial x}\right)\right] \tag{9b}$$

and

$$F_{T,S} = \frac{\partial}{\partial x}\left[A_H \frac{\partial (T,S)}{\partial x}\right] + \frac{\partial}{\partial y}\left[A_H \frac{\partial (T,S)}{\partial y}\right] \tag{10}$$

One should note that F_x and F_y are invariant to coordinate rotation. Although these horizontal diffusive terms are meant to parameterize subgrid scale processes, in practice the horizontal diffusivities, A_M and A_H, are usually required to dampen small-scale computational noise. The diffusivities are chosen so that they do not produce excessive smoothing of real features. The relatively fine vertical resolution used in the applications results in a reduced need for horizontal diffusion, because horizontal advection followed by vertical mixing effectively acts like horizontal diffusion in a real physical sense. A_M and A_H are calculated according to Smagorinski (1963).

$$A_{M,H} = c\Delta^2 \frac{1}{2}\left[\left(\frac{\partial U_i}{\partial x_j} + \frac{\partial U_j}{\partial x_i}\right)^2\right]^{\frac{1}{2}} \tag{11}$$

The Reynolds stress and turbulent heat and salt fluxes, wT and $\overline{ws}$, are evaluated using the level 2 1/2 closure model of Mellor and Yamada (1982) wherein

$$-\left(\overline{uw}, \overline{vw}\right) = K_M \frac{\partial}{\partial Z}(U,V) \tag{12}$$

$$-\left(\overline{wT}, \overline{ws}\right) = K_H \frac{\partial}{\partial Z}(T,S) \tag{13}$$

and the eddy viscosity and diffusivities, K_M and K_H, are given by

$$\left(K_M, K_H\right) = lq\left(S_M, S_H\right) \tag{14}$$

Here, l is the turbulence macroscale and $q^2 = u_i^2$ is twice the turbulence kinetic energy. The stability functions, S_M and S_H, are given in Mellor and Yamada (1982). The level 2 1/2 closure model adds two more prognostic equations to the model, describing the evolution of q^2 and $q^2 l$.

The set of equations (1) to (14) is then transformed using the following bottom and free-surface σ-coordinates

$$x^* = x,\ y^* = y,\ \sigma = \frac{z-\eta}{D},\ t^* = t \tag{15}$$

$$D = H + \eta$$

$$\sigma = 0 \text{ at } z = \eta,\ \sigma = -1 \text{ at } z = -H \tag{16}$$

where η is free surface elevation and H is the depth.

2.2 Boundary Conditions

The boundary conditions at the free surface, $z = \eta(x, y)$, are

$$\rho_0 K_M\left(\frac{\partial U}{\partial z}, \frac{\partial V}{\partial z}\right) = \left(\tau_{0x}, \tau_{0y}\right) \tag{17}$$

$$\rho_0 K_H\left(\frac{\partial T}{\partial z}, \frac{\partial S}{\partial z}\right) = \left(\dot{H}, \dot{S}\right) \tag{18}$$

$$q^2 = B_1^{2/3} u_{\tau S}^2 \tag{19}$$

$$q^2 l = 0 \tag{20}$$

$$W = U\frac{\partial \eta}{\partial x} + V\frac{\partial \eta}{\partial y} + \frac{\partial \eta}{\partial t} \tag{21}$$

where (τ_{0x}, τ_{0y}) is the surface wind-stress vector with friction velocity, $u_{\tau s}$, the magnitude of the vector. It is unlikely that the mixing length is reduced to zero at a surface containing wind-induced waves as given by (17). Error is incurred in the near-surface layer of thickness the order of the wave height. This is an area where further improvement in the model is necessary. The quantity $B_1^{2/3}$ is an empirical constant (6.51) arising from the turbulence closure relationship.

The net heat flux through the water surface is $\dot{H}$, and $S = S(0)[E - P]/\rho_0$ is surface mass flux rate, where $(E - P)$ is the net evaporation-precipitation of freshwater and $S(0)$ is the surface salinity. The net heat flux is given as follows:

$$\dot{H} = (\phi_s - \phi_{sr}) + (\phi_a - \phi_{ar}) - \phi_{br} - \phi_e - \phi_c \tag{22}$$

where ϕ_s (incident solar radiation; short wave) is measured directly by pyrheliometer. When direct measurement is not available, the following empirical formulae are used:

$$\phi_{sn} = \phi_s - \phi_{sr} = 0.94\phi_{sc}\left(1 - 0.65C^2\right) = \text{net incident solar radiation} \tag{23}$$

$$\phi_{an} = 5.9 \times 10^{-3}\left(\frac{e}{T_a}\right)^{\frac{1}{7}} T_a^{\ 4}\left(1 - 0.17C^2\right) = \text{net atmospheric radiation} \tag{24}$$

$$\phi_{br} = 5.9 \times 10^{-3}\left(T_a + 273\right)^4 = \text{back radiation from the water surface} \tag{25}$$

$$\phi_c = \left(0.000308 + 0.000185W_z\right)\rho\left(e_s - e_z\right)\left(2493 - 2.26T_s\right) \times 10^3 = \text{evaporative heat flux} \tag{26}$$

$$\phi_c = 269.1\left(0.000308 + 0.000185W_z\right)\rho\left(T_s - T_z\right) = \text{conductive heat flux} \tag{27}$$

where ϕ_{sc} is clear sky solar radiation, ϕ_{sr} is reflected solar radiation, ϕ_a is incident atmospheric radiation, c is the fraction of the sky covered by cloud, e_s is saturated vapor pressure at the temperature of the water surface, e_z is vapor pressure at height z, T_s is water surface temperature (°C), W_z is wind speed at height z, and ρ is the density of the water.

On the side walls and bottom of a bay, the normal gradients of T and S are zero such that there are no advective or diffusive heat and salt fluxes across these boundaries. At the lower boundary (b),

$$\rho_0 K_M \left(\frac{\partial U}{\partial z}, \frac{\partial V}{\partial z} \right) = \left(\tau_{bx}, \tau_{by} \right) \tag{28}$$

$$q^2 = B_1^{2/3} u_{\tau b}^2 \tag{29}$$

$$q^2 l = 0 \tag{30}$$

$$W_b = -U_b \frac{\partial H}{\partial x} - V_b \frac{\partial H}{\partial y} \tag{31}$$

where $H(x, y)$ is the bottom topography and $u_{\tau b}$ is the friction velocity associated with the bottom frictional stress (τ_{bx}, τ_{by}). The bottom stress is determined by matching velocities with the logarithmic law at the wall. Specifically,

$$\left(\tau_{bx}, \tau_{by} \right) = \rho_0 C_D \left(U_b^{\,2} + V_b^{\,2} \right)^{1/2} \left(U_b, V_b \right) \tag{32}$$

with value of the drag coefficient C_D given by

$$C_D \equiv \kappa^2 \left(\ln(H + z_b) / z_0 \right)^{-2} \tag{33}$$

where z_b, u_b and v_b are the grid point and corresponding velocities in the grid point nearest the bottom, respectively, and κ is the von Karman constant. The final result of (32) and (33) in conjunction with turbulence-closure derived K_M is that the calculations will yield

$$U = \left(\tau_{bx} / \kappa u_{\tau b} \right) \ln(z / z_0) \tag{34}$$

$$V = \left(\tau_{by} / \kappa u_{\tau b} \right) \ln(z / z_0) \tag{35}$$

in the lower boundary region if enough resolution is provided. In those instances where the bottom boundary layer is not well resolved, it is more appropriate to specify $C_D = 0.0025$. The actual algorithm is to set C_D to the larger of the two values given by (32) and to 0.0025. The parameter z_o depends on the local bottom roughness; in the absence of specific information, $z_o = 1$ cm is used, as suggested by Weatherly and Martin (1978).

Open lateral boundary conditions are problematic because the environment exterior to the relevant domain must be parameterized. Two types of open boundaries exist, inflow and outflow. Temperature and salinity are prescribed from data at an outflow boundary, wherein at outflow boundaries

$$\frac{\partial}{\partial t}(T, S) + U_n \frac{\partial}{\partial n}(T, S) = 0 \tag{36}$$

is solved, where the subscript n is the coordinate normal to the boundary. Turbulence kinetic energy and the macroscale quantity ($q^2 l$) are calculated with sufficient accuracy at the boundaries by neglecting advection in comparison with the values of other terms in the respective equations.

The open lateral-velocity boundary conditions in some of the applications are computed by using the available hydrographic data in conjunction with a simplified diagnostic model. This type of model uses only geostrophic plus Ekman dynamics and therefore solves a simplified form of the full equations of motion. It does not require a velocity at a reference level but only along a single transect crossing f/H contours. A detailed description of this model can be found in Kantha et al. (1982). While the normal component of velocity is specified, a free-slip condition is used for the tangential component.