NOTES AND CORRESPONDENCE

A Model of the Tidally Induced Residual Circulation in the Gulf of Maine and Georges Bank

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ABSTRACT

A three-dimensional nonlinear numerical hydrodynamic model using Legendre polynomials to represent the vertical structure of the horizontal currents has been used to study the tidally induced residual flows in the Gulf of Maine-Georges Bank study region using a 6.25 km square grid. Tidal elevations in terms of the M2 phase and amplitude along the open boundaries are specified using Schwiderski's deep ocean tidal model. The model predicts strong clockwise circulation gyres around Georges Bank and Nantucket Shoals with a weak gyre around Browns Bank. Strong inflow to the Gulf of Maine is predicted near the Southwestern tip of Nova Scotia. These results are in good agreement with recent model predictions of Greenberg.

1. Introduction

Bigelow (1927) first presented a description of the summer mean circulation in the Georges Bank, Gulf of Maine and Bay of Fundy showing a large counterclockwise gyre in the Gulf of Maine, a clockwise gyre over Georges Bank, and clockwise circulation around Nantucket Shoals and Island. An inflow around the southwestern tip of Nova Scotia merges with the southerly side of the Gulf of Maine eddy to form a mean flow into the Bay of Fundy along the Nova Scotian shore which is compensated by an outflow on the north side of the Bay.

Bumpus and Lauzier's (1965) atlas of currents, based on an analysis of drifter data, shows a decided seasonal variation in the circulation pattern with the disappearance of the Georges Bank gyre in the fall and winter and the lack of any recognizable pattern in the Gulf in the winter. The Bumpus and Lauzier patterns are consistent with those of Bigelow for the summer season.

Recent field observations support this basic circulation pattern but continue to delineate the variations and add detail to the picture. Butman et al. (1982), using an extensive analysis of both Eulerian and Lagrangian current observations show a mean (averaged over the entire record length) clockwise circulation gyre around Georges Banks at all depths with a narrow jet-like structure along the northern flank. The water on the southern side of the bank flowed toward the Mid-Atlantic Bight with a portion of the water flowing along the eastern side of Great South Channel and recirculating around the Bank. The satellite tracked drogues, however, showed considerable variability in closure of the gyre. The winter current observations of Vermersch et al. (1979) in the western Gulf of Maine, while showing considerable variability in mean current strength and direction with depth, are consistent with the Gulf of Maine gyre. Ramp et al. (1981) show a mean inflow at depths 100 m and below along the north side of the Northeast channel but varying significantly with depth on the south side. Smith's (1983) data show a small clockwise gyre around Browns Bank and a well defined mean and seasonal inflow into the Gulf of Maine around southwest Nova Scotia.

Recent theoretical work, numerical modeling and analysis of observations have suggested that the tidally induced residual currents could be a major contributor to the observed mean flow patterns. Loder (1980) has illustrated, using an analytic model, that large tidal flows across the northern flank of Georges Bank can be rectified to produce the jet-like feature noted by Butman et al. (1982). Magnell et al. (1980) also noted the jet on the north flank and indicated the importance of the tidal nonlinear interaction and the effects of the density structure as well. Numerical modeling of the residual tidal current patterns by Spaulding et al. (1982a) at 12.5 km resolution and Greenberg (1983) with ~7 km grid have further detailed the potential contributions of the tide to the observed mean circulation pattern.

As part of a larger study to understand the roles of tide, wind, density and offshore pressure gradients in driving the mean current pattern, the goal of this paper is to present the results of the tidal simulations performed to date. This work is intended to supplement Greenberg's (1983) recent numerical modeling study of tidally induced flows showing results computed using another modeling approach, different grid structure and size, and an alternate set of tidal amplitude and phase boundary conditions for the cross- and along-shelf open boundaries.
2. The numerical model

The three-dimensional numerical model employed for the present study follows the development given in Owen (1980), Gordon (1982) and Isaji et al. (1982). Only a brief overview is presented here with concentration on the tidally induced residual flows. While the selection of a three-dimensional hydrodynamic model was not strictly necessary for the present investigation, where only vertically averaged flows are addressed, it was selected in anticipation of the extension of the present study to incorporate wind and density as driving mechanisms for the mean flow.

The three-dimensional conservation equations for water mass and momentum suitable for limited shelf waters in Cartesian coordinates may be written:

Conservation of water mass,
\[
\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0, \tag{1}
\]

Conservation of momentum,
\[
\frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} + w \frac{\partial u}{\partial z} - f u = -\frac{1}{\rho} \frac{\partial p}{\partial x} \\
+ \frac{\partial}{\partial z} \left[ N \frac{\partial u}{\partial z} + N_h \left[ 2 \frac{\partial^2 u}{\partial x^2} + \frac{\partial}{\partial y} \left( \frac{\partial u}{\partial y} + \frac{\partial v}{\partial x} \right) \right] \right], \tag{2}
\]
\[
\frac{\partial v}{\partial t} + u \frac{\partial v}{\partial x} + v \frac{\partial v}{\partial y} + w \frac{\partial v}{\partial z} + f u = -\frac{1}{\rho} \frac{\partial p}{\partial y} \\
+ \frac{\partial}{\partial z} \left[ N \frac{\partial v}{\partial z} + N_h \left[ 2 \frac{\partial^2 v}{\partial y^2} + \frac{\partial}{\partial x} \left( \frac{\partial u}{\partial y} + \frac{\partial v}{\partial x} \right) \right] \right], \tag{3}
\]
\[
\frac{\partial w}{\partial t} + u \frac{\partial w}{\partial x} + v \frac{\partial w}{\partial y} + w \frac{\partial w}{\partial z} = -\frac{\partial \rho g}{\partial z}, \tag{4}
\]

where the following notation has been used:

- \(x, y, z\): Cartesian coordinate system with \(x\) and \(y\) measured in the horizontal plane and \(z\) measured vertically upward from mean sea level.
- \(u, v, w\): components of the current in the \(x, y, z\) directions.
- \(p\): pressure.
- \(\rho, \bar{\rho}\): density, depth-averaged density.
- \(N_h, N\): horizontal and vertical eddy viscosity.
- \(f\): Coriolis parameter (2 \(\Omega\) \(\sin\phi\)), assumed constant, where \(\Omega\) is the angular speed of the earth's rotation and \(\phi\) the latitude angle.
- \(g\): acceleration due to earth's gravity.

These equations are then subjected to boundary conditions to complete the model description. At land boundaries the normal component of velocity is set to zero while at the open boundaries the sea surface elevation is specified as a series of sine waves each with its own amplitude and phase.

At the sea surface \((z = \eta)\), the applied stress is
\[
\rho N \left( \frac{\partial u}{\partial z}, \frac{\partial v}{\partial z} \right) = (\tau_{ux}, \tau_{wy}),
\]
where \((\tau_{ux}, \tau_{wy})\) are the components of the surface wind stress in the \(x\) and \(y\) directions. For the present case, the wind stresses are set at zero. The kinematic boundary condition at the surface is given as
\[
w = \frac{\partial \eta}{\partial t} + u \frac{\partial \eta}{\partial x} + v \frac{\partial \eta}{\partial y}.
\]

At the sea bed a quadratic approximation is employed
\[
\rho N \left( \frac{\partial u}{\partial z}, \frac{\partial v}{\partial z} \right) = (\tau_{bx}, \tau_{by}) = C_f \rho (u_b, v_b)(u_b^2 + v_b^2)^{1/2},
\]
where \((\tau_{bx}, \tau_{by})\) are the \(x\) and \(y\) components of the bottom shear stress, the subscript \(b\) denotes evaluation of the parameter at the sea bed and \(C_f\) is a bottom friction coefficient.

Anticipating the use of a weighted residual method, in which vertical variations are represented in terms of a set of basis functions, it is desirable to introduce a new set of independent variables that transforms both the surface and the bottom onto coordinate surfaces. We define this transformation of coordinates \((x, y, z)\) to \((X, Y, \sigma, T)\) by
\[
x = X, \quad y = Y, \quad \sigma = 1 - 2 \left( \frac{\eta - z}{h + \eta} \right), \quad t = T.
\]

As a result, the local depth range of the water column \((-h, \eta)\) transforms to the interval \((-1, 1)\). A detailed presentation of the transformed equations can be found in Owen (1980) and Gordon (1982).

The numerical solution methodology follows that of Davies (1977a,b) and Owen (1980). The vertical variations in horizontal velocity are described by an expansion of Legendre polynomials. The resulting equations are then solved by a Galerkin weighted residual method in the vertical and by an explicit finite difference algorithm in the horizontal.

A space staggered grid scheme in the horizontal plane is used to define the study area. Sea surface elevation and vertical velocity are specified in the center of each cell while the horizontal velocities are given on the cell faces. The \(u\) and \(v\) velocities are defined on the cell faces normal to the \(x\) and \(y\) directions, respectively.

To allow time steps larger than the Courant-Friedrichs-Lewy (CFL) condition, which relates the maximum time step to the shallow water wave celerity, a “split-mode” or “two-mode” formulation was used (Simons, 1975; Sheng et al., 1978; Gordon, 1982). In the split-mode models, the free-surface elevation is treated separately from the internal, three-dimensional flow variables. The free-surface elevation and vertically integrated velocities are calculated using the vertically integrated equations of motion (external mode) for which the CFL limit must be met. The vertical structure of the horizontal components of the current then may
be calculated in such a manner that the effects of the surface gravity waves are separated from the three-dimensional equations of motion (internal mode). Therefore, surface gravity waves no longer limit the internal mode calculations and much longer time steps are possible.

The external mode equations are approximated by a forward in time, centered in space (FTCS) finite difference scheme. The internal mode is solved using a forward in time technique with the vertical diffusive terms centered in time, to ease the vertical time step restriction from the standard FTCS procedure. A detailed presentation of the finite difference approximations to the governing equations, a simplified stability analysis of the approximation, and methods to control time splitting inherent in split-mode computations are given in Gordon (1982). Gordon also has undertaken an extensive test of the computer code based on this approach, (Owen, 1980) comparing model predictions to analytic solutions for a standing wave, wind-induced flow, and damped tidal flow in a rectangular basin and density-driven flow in an open channel. Additional tests were performed to study the computational characteristics of the model including sensitivity to the number of polynomials employed, constant and variable vertical eddy viscosities (linear, parabolic, arbitrary shaped), and the time-splitting ratio (internal-to-external time steps).

3. Application to Georges Bank—Gulf of Maine

The first step in the application of the hydrodynamic model to the study area (Fig. 1) is the decision on how the region is to be discretized and includes selection of the grid size and orientation, and the location of the model boundaries. Aligning the x-axis of the model (J-direction, positive to the northwest) in the longshelf direction (56 deg clockwise from true North) and the y-axis (J-direction, positive toward the coast) in the cross-shelf direction, a model domain having square grids of 6.25 km on a side was selected for the Georges Bank—Gulf of Maine region (Fig. 2). Location of the outer boundaries was based on a review of other modeling studies of the region (Greenberg, 1977; Beardsley and Haidvogel, 1981), availability of tidal constituent data, and dynamic considerations. The basic strategy was an attempt to minimize the influence of boundary condition specification on the model-predicted flows. To this end, the cross-shelf boundaries were located in regions of smoothly varying long-shelf bathymetry.
while the long-shelf boundary was located approximately 100 km seaward of the shelf break. Depth data required as input to the model was extracted from the digitized bathymetry dataset from the NOAA National Geophysical Solar, Geophysical, and Terrestrial Data Center; the National Ocean Survey navigational charts; or from Greenberg's fine-scaled (7 km) grid (Greenberg, 1983). A false bottom of 370 m depth was used beyond the shelf break.

Selection of the bottom friction factor was based on comparing a series of model predictions of tidal range and phase for varying $C_f$ with observations for a series of tidal measurement stations throughout the study area. A friction factor, $C_f$, equal to 0.003 provided the minimum global error.

The $M_2$ component is dominant throughout the region composing the major portion of the tidal energy (Moody and Butman, 1980). Boundary conditions describing the amplitude and phase of the $M_2$ component were generally obtained from Schwiderski’s (1979, 1980) (also see Schwiderski and Szeto, 1979) world ocean tidal model. Comparison of Schwiderski’s tidal model predictions along the present model’s open boundaries with available observations showed very good agreement except on the southern New England cross-shelf boundary. In this region the degenerate amphidrome centered on the eastern tip of Long Island (Petrillo, 1982) was not adequately represented in Schwiderski’s one-degree square grid structure. Note that using Schwiderski’s model to specify the open boundary conditions allowed the study region to be better matched with tidal conditions in the North Atlantic, as opposed to the common practice of linearly interpolating between and extrapolating from known tidal measurement stations. (Greenberg, 1977, 1979; Spaulding and Gordon, 1982; Spaulding and Beaufort, 1982; Leendertse and Liu, 1979; Leendertse et al., 1973).

Figure 3 displays the model predicted tidal current ellipses for the study region. The shallow water on Georges Bank causes high tidal velocities (compare 5 cm $\text{s}^{-1}$ tidal currents in deep water to 50–80 cm $\text{s}^{-1}$ on the Bank) while the damped co-oscillating tide in the Bay of Fundy causes the large tidal currents at the entrance to the Bay. Further comparison of the model predicted tidal ellipses with those observed by Moody and Butman (1980) on Georges Bank shows good agreement in terms of magnitude and shape. (Fig. 4).

4. Tidal residuals

The tidally induced residual flows for the study area were calculated by integrating the results of the tidal simulations over thirty tidal cycles. This length of integration time was required in order to assure that the residual currents had achieved a steady state, defined as less than 0.1% change in residual speed from one tidal cycle to the next.

The residual flows resulting from this simulation are shown in Fig. 5 in terms of velocity vectors and clearly illustrate clockwise circulation gyres around Georges Bank, Nantucket Shoals, and Browns Bank. The Georges Bank region shows a tight gyre around the top of the bank with the shape conforming closely to the sharp depth gradients (Fig. 1) caused by Little Georges Bank and the series of lineal shoals on the northwestern part of the bank (Cultivator and Georges Shoals). This small gyre is included within a larger
Fig. 3. Model predicted tidal current ellipses at selected locations. Arrows denote direction of rotation.

Fig. 4. Comparison of model predicted tidal current ellipses (with arrows) on Georges Bank with observations of Moody and Butman (1980).
gyre displaying strong currents along the northern flank of the bank and weaker return currents on the seaward side. Comparison of the present model depth averaged predictions for current strength on the north side of Georges Bank (10–25 cm s⁻¹) are in good agreement with observations of Butman et al. (1982) [15–30 cm s⁻¹] and the theoretical value (23 cm s⁻¹) of Loder (1980) based on an idealized geometry, but higher than Greenberg’s model results (11 cm s⁻¹). The circulation around Nantucket Shoals is strikingly similar to Greenberg’s (1983) results and consists of two interconnected clockwise gyres, one around the southeastern end of the shoals and another around Nantucket Island, embedded in a larger clockwise gyre.

The model predicted clockwise gyre around Browns Bank is in agreement with current meter data of Smith (1983). Smith shows annual mean depth average currents of 9 cm s⁻¹ on the north side of the bank while the present model predicts 6 cm s⁻¹ at the same site. The model is also in qualitative agreement with Smith’s data and Greenberg’s model showing strong residual flows into the Gulf of Maine close to the Nova Scotia coast.

The model results indicate no Gulf of Maine eddy structure but show some northeastward directed currents along the northern Maine coast. In the entrance to the Bay of Fundy two gyres, a clockwise gyre near the east of Grand Manan Island adjacent to a clockwise gyre due south of the Island. The currents are particularly strong where the gyres connect but the pattern is confusing at best.

The unusual shore directed residual current pattern along the cross-shelf boundary south of Long Island is likely induced either by an improper tidal height boundary condition or the zero-mean cross-shelf elevation gradient employed. Specification of the tidal phase and amplitude in this region is hampered by the complex tidal wave behavior and the lack of good quality offshore tidal height observations. The present model predictions, while consistent with those of Greenberg (1983), do not show the relatively large but confused residual currents along the shelf break observed in Greenberg’s model.

5. Conclusions

The present model, although using a different grid structure/size, different boundary conditions and an alternate formulation and solution of the conservation of mass and momentum equations predicts tidally induced residual current patterns very similar to those of Greenberg (1983). The strong jet-like feature along the north flank of the bank found in Magnell et al. (1980) analysis of current observations is readily reproduced here and represents the strong nonlinear tidal interactions caused by the rapid variation in bottom topography present in the region. The model also predicts a complex structure to the Georges Bank gyre with the details of the pattern controlled by the sharp depth gradients defining the bank boundaries and the shallow northwesterly oriented shoal structure on the top of the bank.

Comparison of the present model predictions to the observed mean patterns shows that the Gulf of Maine gyre is not generated by the tide. Recent simulations, not reported here, suggest that the inclusion of density forcing is necessary for the development of the gyre. This result is in general agreement with Greenberg’s conclusions.

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