Modeling the Mean Barotropic Circulation in the Bay of Fundy and Gulf of Maine

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ABSTRACT

Two two-dimensional, nonlinear numerical models are used to study the residual barotropic circulation generated by tides and steady winds in the Bay of Fundy and Gulf of Maine. The first, a multi-grid model, is used to examine the Bay of Fundy with a coarse look at the Gulf of Maine. The circulation in the upper Bay of Fundy is predominantly tidally driven. The model clearly reproduces the major gyres observed at the head of the Bay. Steady wind stresses have some effect on the strength of these currents but little effect on the pattern. The counterclockwise gyre, observed in the body of the Bay of Fundy, is not reproduced in the model. The second model covers the lower Bay of Fundy and the Gulf of Maine with a single fine grid and is used to look at details in the Gulf of Maine. A clockwise circulation around Georges Bank and Nantucket Shoals is clearly indicated from tidal forcing alone, as is a gyre over the shallow part of Browns Bank. Different steady wind-stress fields give rise to variations in current strength and current patterns. The counterclockwise Maine eddy is only found in the model when forced by a steady northeast wind stress in addition to tides.

1. Introduction

It is generally believed that a reasonable delineation has been made of the mean circulation in the Bay of Fundy and Gulf of Maine (Fig. 1). However, the pictures that emerge from various studies show considerable variation from season to season, and are not always consistent. Bigelow's (1927) description of the nontidal summer circulation (Fig. 2) shows a large counterclockwise gyre (the Maine eddy) in the body of the Gulf of Maine, a clockwise gyre over Georges Bank, and clockwise circulation around Nantucket Shoals. Flow into the Gulf is observed between Georges Bank and Nova Scotia, with a compensating outflow south and southwest from Georges Bank. Flow into the Bay of Fundy along the southern side and outflow along the northern side of the Bay are indicated.

The above qualitative circulation pattern has generally been supported by more recent work, but some variations and more details have been noted. Watson (1936) attributed the circulation in the lower Bay of Fundy to density effects. His picture of surface circulation is consistent with that of Bigelow, but he finds the deep waters flowing generally into the Bay. Bumpus (1960) reported wind effects disrupting the Maine eddy and complicated flow patterns around the mouth of the Bay. He also observed seasonal and secular changes in surface drift which he felt could be related principally to variations in river runoff and, to some extent, wind. Bumpus and Lauzier's (1965) analysis of drift bottle data for summer is in broad agreement with Bigelow's analysis. It showed some seasonal variability, notably the absence of the Georges Bank gyre in fall and winter, and the absence of any recognizable pattern in the Gulf in winter. The Bumpus-Lauzier seasonal charts give no clear indication of circulation around Nantucket shoals.

There are observations showing that the mean flow has a significant barotropic component. Lauzier (1967) inferred bottom currents from bottom drifters in the northeast Gulf and in the Bay of Fundy. These are consistent with Bigelow's surface pattern showing flow around southwest Nova Scotia into the Bay and flow out of the Bay on the New Brunswick side but differ in indicating a strong onshore bottom drift around southwest Nova Scotia. Winter current-meter observations of Vermersch et al. (1979) in the southwest Gulf show some variation of mean current strength and direction with depth, but are still consistent with the Maine gyre. Similarly, observations of Butman et al. (1982) show the flow to be clockwise around Georges Bank at all depths, and observations of Ramp et al. (1981) at depths 100 m and below show a mean inflow at all depths on the north side of the Northeast Channel although varying significantly with depth on the south side. Smith's (1983) observations indicated mean and seasonal components of inflow into the Gulf around southwest Nova Scotia in the upper layers but some differences were noted in near-bottom current-meter data. His data also indicated a clockwise gyre around Browns Bank. Godin's (1968) analysis of Bay of Fundy current-meter data indicated some variability in the Bay of
Fundy gyre. It also showed very strong currents in the Minas Channel from which he deduced the existence of eddies near Cape Split. He concluded that a strong outward mean flow observed along Cape Split must be balanced by an inward flow on the opposite side of the channel. Tee (1977) supplemented these observations and found further eddies in the narrows at Cape Split and off Cape Blomidon.

Theoretical studies give indications of some driving mechanisms that could be important to the barotropic mean flow. Csanady (1974) suggests that the Maine eddy and the Georges Bank gyre are driven by a northeast wind stress, but given Saunders' (1977) analysis of wind data it is difficult to see that such a stress is present in a magnitude or a duration that would drive the observed seasonal-mean circulation. Loder (1980) has shown that large tidal flows across the edge of a bank can be rectified to produce
2. The numerical models

The two numerical models used are fairly standard two-dimensional schemes with only minor variations from those in common use for the last ten years or so. Only brief descriptions will be given here. More details are given in the Appendix.

The depth-averaged equations of continuity and motion as used in the models may be written:

\[ \frac{\partial \zeta}{\partial t} + \frac{\partial}{\partial x} [(h + \zeta)u] + \frac{\partial}{\partial y} [(h + \zeta)v] = 0, \]  

\[ \frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} = f v - g \frac{\partial \zeta}{\partial x} - ku \frac{(u^2 + v^2)^{1/2}}{h + \zeta} + \frac{F_x(x, y)}{\rho(h + \zeta)} + A_h \nabla^2 u, \]  

\[ \frac{\partial v}{\partial t} + u \frac{\partial v}{\partial x} + v \frac{\partial v}{\partial y} = -fu - g \frac{\partial \zeta}{\partial y} - ku \frac{(u^2 + v^2)^{1/2}}{h + \zeta} + \frac{G_x(x, y)}{\rho(h + \zeta)} + A_h \nabla^2 v, \]

where the notation used is as follows:

- \( t \) time
- \( x, y, z \) left-handed Cartesian system, with \( z \) vertically down and the \( x, y \) plane coinciding with the undisturbed surface
- \( h \) depth of the undisturbed water
- \( \zeta \) elevation of the water surface
- \( u, v \) depth-averaged currents in the \( x \) and \( y \) directions
- \( F_x, G_x \) local surface wind-stress components in the \( x \) and \( y \) directions
- \( \rho \) density of sea water (assumed constant)
- \( f \) Coriolis parameter, varied with latitude
- \( g \) acceleration due to gravity
- \( k \) frictional coefficient
- \( A_h \) horizontal eddy viscosity coefficient
- \( \nabla^2 \) horizontal Laplacian operator

Model definition is completed by taking zero flow normal to shore boundaries, and specifying sinusoidal tidal elevations with zero mean along the open boundaries. Reservations about open-boundary specification are considered in a later section.

The first model (Fig. 3) is essentially as described by Greenberg (1976, 1979), and was used to examine the circulation in the Bay of Fundy and to give a coarse look at the Gulf of Maine. There have been two changes from the original model. The advective terms, which were earlier included only in the Minas area, are now used throughout the model. The numerical scheme used (Siedlecki and Wurtele, 1970) in this model can give rise to grid-scale oscillations in the current field. This is seen in some model results as currents changing direction and magnitude regularly at each successive grid square. Although not unstable, this tended to obscure the underlying large-scale features of the mean current pattern. An eddy-viscosity formulation based on that of Schwiderski (1980), which depends on grid size and depth, smoothed the results, partially damping out these oscillations. The addition of this term caused no other
Fig. 3. Grid outline of the Bay of Fundy—Gulf of Maine numerical model. The grid in the upper Bay is only approximate and the grid in Minas Basin is not shown because of presentation limitations.
obvious changes and a recalibration of the model based on the observed tidal data was not necessary.

The second model (Fig. 4) was used to examine in more detail the circulation in the Gulf of Maine. The numerical method employed aspects of the schemes of Flather and Heaps (1975) and Henry (1981) which allowed time-centering of the advective and Coriolis terms and reduced the artificial generation of grid-scale oscillations. The eddy-viscosity term was therefore not included in the formulation of this model. The upper Fundy region is cut off with an open boundary, at which tidal amplitude and mean elevation are prescribed according to values from the first-model results with no applied wind stress. This can lead to some residual transport through the Bay (exaggerated in the diagrams since Eulerian currents are plotted in these shallow waters where the Stokes drift is of order 1 cm s$^{-1}$ headward). This did not unduly influence the results in the Gulf of Maine.

The two models were independently calibrated using tidal-elevation data both onshore and offshore and run for several tidal cycles with a specified wind-stress distribution until the motion was steady from one cycle to the next. Results were then averaged over the last tidal cycle to obtain Eulerian residual flows.

There has been some debate as to whether residual circulation is best modeled as described here or with the equations in a linearized form (see Nihoul and Ronday, 1975). Heaps' (1978) analysis of the two techniques indicated that reasonable results have
been obtained from both methods. However he quoted unpublished work that suggested convergence of the linearized equations could be a problem if friction were to vary spatially as would be necessary in the Bay of Fundy–Gulf of Maine system.

3. Tidal residuals

a. The Bay of Fundy

The tidally induced residual circulation in the upper Bay of Fundy as derived from the first model is shown in Fig. 5. The four gyres indicated by Tee (1976, 1977)—two in the narrows by Cape Split, one off Cape Split and one off Cape Blomidon—are clearly delineated. The magnitudes of the currents are also consistent with the observations of Tee (1977) and Godin (1968) (Fig. 6). The strong gyre in Minas Channel to the west of Cape Split has currents of order 50 cm s\(^{-1}\). The model indicates a strong seaward residual along the northern side of Cape Split, increasing from 20 to 170 cm s\(^{-1}\) from Cape Blomidon to the end of Cape Split. The one current meter central to this flow, but close to shore, indicated a residual current of 78 cm s\(^{-1}\) seaward. Similarly, with the gyre in Minas Basin to the east of Cape Blomidon, the observations and the model are consistent at points of observation with stronger currents on the Cape Blomidon side of the gyre (|u| > 20 cm s\(^{-1}\)) and weaker currents headward (|u| \approx 10 cm s\(^{-1}\)). The model indicates that the strongest currents of this gyre are closer to the narrow entrance to the Minas Basin. Other eddies, which could be similarly generated by strong tidal currents around headlands are also pres-

![Diagram of Bay of Fundy and Minas Basin](image)

**Fig. 5.** The model-produced residual currents in the upper Bay of Fundy with only tidal forcing. The size of the vectors is proportional to \(\log_{10}(1 + \text{current speed})\).
ent. With the exception of current-meter data near Cape Enragé (K. T. Tee, personal communication) that agree with model results showing a cross-channel residual, there is little observational data to confirm these current magnitudes or patterns. Interpretation of results in Upper Minas and Upper Chignecto is hindered by poor resolution and by numerical grid-scale oscillations not completely removed by the eddy-diffusion term in the equations. In the main part of the Bay of Fundy, the inward flow close to the Nova Scotia side of the Bay and the broad return flow in the center and on the New Brunswick side, with magnitudes of 3–10 cm s$^{-1}$, are in agreement with Godin’s (1968) analysis.

The model results for the lower Bay of Fundy (Fig. 7) differ in several respects from the established pattern (e.g., Godin, 1968). The accepted picture (e.g., Fig. 8) has a strong flow along the Nova Scotia coast into the Bay of Fundy, outflow along the New Brunswick coast and seaward of Grand Manan Island, and a gyre in the lower part of the Bay between the two flows. On the New Brunswick side the model flow is in the observed direction only northeast of St. John and seaward of Grand Manan Island and in the opposite direction elsewhere. Any counterclockwise gyre in the Lower Bay is obscured in an unrecognizable pattern of current vectors. Lagrangian currents were examined here by the addition of the Stokes drift, associated with time-varying surface elevation, to these Eulerian currents, but only slight differences in pattern were noted. The barotropic effect of the Saint John River, the largest tributary to the Fundy-Maine system, was found not to be significant when the maximum river flow was added to a model simulation. The gyre by Saint John is confirmed in Neu’s (1960) observations, but the discrepancies indicate that the mean flow in the lower Bay is not adequately represented in this model.

b. The Gulf of Maine

The expected clockwise circulations around Georges Bank and Nantucket Shoals are reproduced in the first model (Fig. 9), but the resolution is poor and the strength of the current on the north side of Georges Bank is low compared to the observations of Butman et al. (1982) of 15–30 cm s$^{-1}$, depending on depth, and the theoretical value (23 cm s$^{-1}$) of Loder (1980) based on idealized topography. This suggested that better model resolution was necessary to reproduce these currents. Wright and Loder (personal communication) find values of up to 14 cm s$^{-1}$ with a more realistic topography in Loder’s depth-independent model, but larger values due to frictional effects when depth-dependence is considered. The second model (Fig. 10) was designed to rectify the resolution problem. There is better detail showing a tight gyre on top of the Bank and a larger gyre around the edge. Stronger currents, up to 11 cm s$^{-1}$, are indicated. The circulation around Nantucket Shoals appears to consist of two smaller clockwise cells, one around the southeast end of the shoals and another around Nantucket Island, both embedded in a larger clockwise gyre. Neither model adequately reproduces the Maine eddy; indeed, the currents along the Maine Coast are reversed in the first model. There is little order to the current pattern at the mouth of the Bay of Fundy.

The increased resolution of the second model enables the identification of several additional features that have been noted in observations. Current-meter data of Smith (1983) show the gyre around the cap of Browns Bank and indicate strong residual flows
into the Gulf of Maine close to the Nova Scotia coast. Smith’s depth-averaged annual-mean current of 9 cm s\(^{-1}\) on the north side of Browns Bank compares to the model current at that site of 5 cm s\(^{-1}\) and a peak current in the gyre of 8 cm s\(^{-1}\). Smith’s observations and those of Ramp et al. (1981) support the model results giving a strong inflow on the northeast side of the Northeast Channel. Their values vary depending on location and depth from 2–15 cm s\(^{-1}\) compared to the model current of \(\sim 5\) cm s\(^{-1}\) in that area. However, this is close to the fixed open boundary of the model, and caution is therefore warranted in drawing conclusions. Outflow from the modeled system is noted southwest along the New England shelf at 1–2 cm s\(^{-1}\) and comparable to the observations of Butman et al. (1982) of \(\sim 5\) cm s\(^{-1}\).

4. Steady wind stress

The effects of wind stress were examined coincident with the tides to see how the above circulation pattern might be modified. Saunders’ (1977) seasonal wind-stress values for 1° longitude–latitude squares were used as well as two special uniform-wind cases, a strong northwest and a lighter northeast wind stress. The two models were run for four tidal cycles and results of the last tidal cycle were analyzed. If the motion was not steady from cycle to cycle a further four-cycle run was sufficient to get to steady motion. There were only slight differences in the pattern and magnitude of the residual currents when the additional runs were made.

There are two major factors to be considered in examining these mean wind stress cases. The first is whether the response to the mean wind stress is equal to the mean response to the time-dependent wind stress in that season. This paper addresses only that component of the flow that is driven by the mean stress. The second factor is whether the mean elevation along the open boundary of the model can reasonably be held constant for the wind stresses considered. Sandstrom (1980), in analyzing wind data
and shore-based tide-gauge data, has found that sea
level along the coast is correlated with the along-shelf
wind speed but not with the cross-shelf wind. He was
not able to find a correlation of along-shelf sea-surface
slope with any wind stress (Sandstrom, personal com-
munication). The only stresses considered here that
are not predominantly cross-shelf are the summer
wind stress, which is quite low at about 0.25 dyn cm$^{-2}$
from the southwest and the uniform northeast wind
stress of 0.5 dyn cm$^{-2}$. This does not mean that the
open-boundary slope is unimportant; indeed, it is felt
that even a small cross-shelf slope can modify sig-
nificantly inflow to the Gulf along the Scotian Shelf
(work in progress). However, these slopes are ignored
for the first approximation of the effects of the wind
stresses considered here.

The mean summer wind-stress values for each 1°
square from Saunders (1977) are light, about 0.25
dyn cm$^{-2}$, and from the southwest. The wind-stress
patterns from the other seasons are generally from
the northwest and stronger, from 0.25 dyn cm$^{-2}$ in
the fall to 1.0 dyn cm$^{-2}$ in winter. Wind stress had
very little effect in the upper Bay of Fundy, giving
rise to only minor changes in mean current strength.
Thus only results for the Gulf of Maine from the fine-
grid model will be presented here. All the seasonal
wind stresses generated elements of a nearshore cur-
rent opposing the Maine eddy. Except for the oc-
casional intriguing vector of Bigelow (1927) and of
Bumpus and Lauzier (1965), there is little corrobor-
eating evidence for such a current. The current pat-
tern resulting from the application of the summer
wind stress to the tidal model (Fig. 11) gave rise to
few other changes from the pattern produced solely
by tide. The fall, winter and spring wind stresses drive
similar residual-current patterns, with the winter pat-
tern (Fig. 12) differing most from the pure tidal pic-
ture. The Massachusetts Bay counterclockwise gyre
and the north-to-south flow along the outer arm of
Cape Cod appear and intensify from the weaker fall
stress to the stronger winter stress, then decay some-
what with the intermediate-strength spring stress.
Only the winter wind stress is strong enough to dis-
rupt the outer circulation of the Georges Bank gyre,

![Diagram](https://example.com/diagram.png)

**Fig. 8.** Residual current pattern in the Bay of Fundy (from Godin, 1968). The multiple arrows at single sites indicate areas where the residual currents varied in direction.
reversing the inward flow in the Great South Channel. The circulation around Nantucket Shoals is also modified with the opening of the cell around Nantucket Island. Care must be taken in interpreting these changes in the gyres because of their proximity to open boundaries where the elevations are strictly specified. Similarly it is too speculative, though tempting, to draw conclusions on changes in inflow in the Northeast Channel and on the Scotian Shelf.

A steady and uniform wind stress of 0.5 dyn cm$^{-2}$ from the northeast does give rise to part of the Maine eddy driving a strong counterclockwise flow along the northwest and southwest boundaries of the Gulf (Fig. 13) as predicted by Csanady (1974) and it contributes to the Massachusetts Bay gyre. However, as mentioned earlier, Saunders' (1977) wind-stress analysis shows that mean seasonal stresses are not from the northeast direction, so wind is not thought to provide a principal driving force of this eddy. To examine the effects of a strong persistent wind, the model was run with a northwesterly wind stress of 2 dyn cm$^{-2}$. The results were similar to the winter picture, only more intense. The Georges Bank gyre was strongly perturbed as were both cells of the circulation around Nantucket Shoals.

5. Discussion

The model currents are in agreement in many instances with the observations available. Previous work is useful in identifying some but not all processes important in generating these currents. The
Fig. 10. The high-resolution model-produced residual currents in the Gulf of Maine with only tidal forcing. The circles indicate points where the terms of the momentum equation were analyzed in Table 1. The size of the vectors is proportional to $\log_{10}(1 + \text{current speed})$. 
Fig. 11. The model-produced residual currents driven by tidal forcing and the summer wind-stress values derived by Saunders (1977) for each 1° longitude–latitude square. Stresses are predominantly from the southwest and are about 0.25 dyn cm$^{-2}$ in strength. The size of the vectors is proportional to $\log_{10}(1 + \text{current speed})$. 
Fig. 12. As in Fig. 11 except for the winter wind-stress values derived by Saunders (1977) for each 1° square. Stresses are predominantly from the northwest and are generally from 0.8 to 1.3 dyn cm⁻² in strength.
FIG. 13. The model-produced residual currents driven by tidal forcing and a uniform wind stress of 0.5 dyn cm⁻² from the northeast. The size of the vectors is proportional to log₁₀[1 + current speed].
results, however, should be viewed with the modeling uncertainties in mind.

There are three model limitations that will be reiterated here. First, it can be asked what value a barotropic model has in a system where components of the flow are driven by baroclinic forces. However, model results, observations and theories point to significant barotropic components in the residual flows. The second reservation is over model resolution. It is always tempting to draw conclusions at the smallest scales resolved. Thus the model gyres in Chignecto Bay, which have dimensions only slightly greater than the grid size, need independent corroborations. The finer grid of the second model did give a better picture of the Georges Bank gyre but also indicated tidally driven gyres off the tip of Cape Cod. The latter are so poorly resolved that they should only be thought of as evidence for some sort of tidally driven residual. The irregular assortment of vectors for the entrance to the Bay of Fundy is thought to result from a resolution problem. An independent model (K. T. Tee and D. Lefaivre, personal communication), with the same resolution but a completely different numerical scheme, gives the same sort of results. The third uncertainty pertains to the traditional question about how to define open boundary conditions. In all model runs considered here, the elevation on the boundary is strictly specified to follow a sinusoidal tidal variation with zero mean elevation. In spite of simple theory and observations (Sandstrom, 1980) suggesting that only an along-shore wind would influence this boundary, we have to worry that the real world and less-than-perfect numerical schemes may not concur. The northeast wind stress could give rise to a considerable set-up along the coast, inconsistent with clamped boundaries. The extent of penetration of effects due to this clamping are assumed small enough still to obtain useful qualitative results and are the subject of further study. Besides worries about the theoretical problems, the practical problem of stability has required that the elevation field at corner points adjacent to the boundary be smoothed and the advective terms not be included in the momentum equation at all points adjacent to this boundary. Thus, even though the model indicates that the observed inflow (Ramp et al., 1981; Smith, 1983) into Northeast Channel could be tidally driven, conclusions should be reserved at least until a tidal process is identified that would generate these currents.

Much of the tidally driven circulation seems to be attributable to processes that have already been identified as being significant in this area. Loder (1980) has shown that the cross-isobath tidal flows are important in generating the circulation around Georges Bank. The similarities in topography and observed currents around Nantucket Shoals and Browns Bank suggest that the same process could be important there. Smith (1982) has found experimental evidence for tidally rectified flow on the north side of Browns Bank and off Cape Sable. Tee (1976) has identified inertial effects that dominate processes in Minas Basin where large tidal currents move around headlands creating strong residual currents and several gyres. Similar conditions can be associated with other circulation patterns at the head of the Bay, the Saint John gyre, the southwest Nova Scotia currents, and the inadequately resolved currents off the tip of Cape Cod.

The terms of the momentum equations [(2), (3)] have been averaged over one tidal cycle (Table 1) for various points in the fine-grid model where residual flows line up with the topography and, conveniently, also with the axes of the model. The terms are defined:

\[
\begin{align*}
\frac{\partial u}{\partial x} & = \frac{1}{\tau} \int u \frac{\partial u}{\partial x} dt \\
\frac{\partial v}{\partial y} & = \frac{1}{\tau} \int v \frac{\partial u}{\partial y} dt \\
\text{Coriolis} & = \frac{1}{\tau} \int f v dt \\
\text{Pressure-gradient} & = \frac{1}{\tau} \int \frac{\partial \xi}{\partial x} dt \\
\text{Friction} & = \frac{1}{\tau} \int \frac{k u(u^2 + v^2)^{1/2}}{h + \xi} dt \\
\frac{\partial w}{\partial z} & = \frac{1}{\tau} \int \frac{k v(u^2 + v^2)^{1/2}}{h + \xi} dt.
\end{align*}
\]
Table 1. The time average ($\times 10^3$) of different terms in the momentum equations (cgs) over one tidal cycle at different locations in the high-resolution Gulf of Maine model, for the case of tidal forcing only. The axes are oriented with ($x$, $u$) positive southeast offshore and ($y$, $v$) positive northeast along the shelf. The points examined are circled in Fig. 10.

<table>
<thead>
<tr>
<th>Region</th>
<th>Momentum equation for ($^*$)</th>
<th>$\frac{\partial u}{\partial x}$</th>
<th>$\frac{\partial v}{\partial y}$</th>
<th>Coriolis force</th>
<th>Pressure gradient</th>
<th>Friction</th>
</tr>
</thead>
<tbody>
<tr>
<td>Georges Bank</td>
<td>$u$</td>
<td>-0.94</td>
<td>0.09</td>
<td>0.69</td>
<td>0.20</td>
<td>-0.05</td>
</tr>
<tr>
<td></td>
<td>$v$</td>
<td>0.17</td>
<td>0.04</td>
<td>-0.11</td>
<td>0.17</td>
<td>-0.30</td>
</tr>
<tr>
<td>Nantucket Shoals</td>
<td>$u$</td>
<td>-0.17</td>
<td>0.59</td>
<td>-0.13</td>
<td>0.28</td>
<td>-0.57</td>
</tr>
<tr>
<td></td>
<td>$v$</td>
<td>-0.16</td>
<td>1.54</td>
<td>-1.00</td>
<td>-0.68</td>
<td>0.31</td>
</tr>
<tr>
<td>Browns Bank</td>
<td>$u$</td>
<td>-0.151</td>
<td>-0.066</td>
<td>0.128</td>
<td>0.090</td>
<td>-0.003</td>
</tr>
<tr>
<td></td>
<td>$v$</td>
<td>-0.21</td>
<td>-0.18</td>
<td>0.10</td>
<td>0.34</td>
<td>-0.05</td>
</tr>
<tr>
<td>South Nova Scotia Coast</td>
<td>$u$</td>
<td>1.95</td>
<td>-0.07</td>
<td>-0.74</td>
<td>-1.06</td>
<td>-0.09</td>
</tr>
<tr>
<td></td>
<td>$v$</td>
<td>-1.20</td>
<td>-0.19</td>
<td>0.09</td>
<td>0.78</td>
<td>0.52</td>
</tr>
<tr>
<td>Northeast Channel</td>
<td>$u$</td>
<td>-0.122</td>
<td>-0.043</td>
<td>0.020</td>
<td>0.114</td>
<td>0.032</td>
</tr>
<tr>
<td></td>
<td>$v$</td>
<td>0.03</td>
<td>-0.01</td>
<td>0.26</td>
<td>-0.28</td>
<td>0.00</td>
</tr>
</tbody>
</table>

Here $\tau$ is the tidal period and the other notation is as defined earlier.

On the northern side of Georges Bank the averaged momentum balances closely approximate those in Loder's (1980) idealized model of tidal rectification. In the $u$-momentum equation $\frac{\partial u}{\partial x}$ of the advective terms is balanced by the Coriolis force and the pressure gradient. In the $v$-equation $\frac{\partial v}{\partial x}$ is partly balanced by stress, but results from the numerical model indicate that the pressure gradient and Coriolis force can also be important. Other locations where strong tidal currents run transverse to large bathymetry changes, such as Nantucket Shoals and near the Nova Scotian southern coast, have similar balances. Around Browns Bank, where the situation is analogous, but the residual flows turn in following the depth contours, both parts of the advective terms are important. The residual flow found in the model in the Northeast Channel exhibits a simple balance between $\frac{\partial u}{\partial x}$ and the pressure gradient in the $u$-equation and a balance between the Coriolis and pressure gradient terms in the $v$-equation.

Neither tides nor the seasonal mean winds appear to complete the generation of the Maine eddy or the gyre in the lower Bay of Fundy, although the tidally produced residual flow north of Georges Bank does form that part of the eddy. The Maine eddy has usually been attributed to the baroclinic effects of the freshwater runoff from New Brunswick and Maine. The same process might contribute to the gyre observed in the lower Bay of Fundy, but because of the existing problem with resolution, tidal or wind forcing cannot be ruled out as generating mechanisms in this case.

Wind stresses can have significant effects in the Gulf of Maine. The Massachusetts Bay counterclockwise gyre is not produced by tidal currents. Northwest and northeast wind stresses do generate the gyre while the southwest summer wind stress tends to reverse the gyre flow. The agreement of these model results with the observations of EGK (1976) is excellent. Observations of Schlitz and Allen (1981) show that the model would be hard pressed to resolve spatial variations in the alongshore currents off the outer arm of Cape Cod, but do support the model results showing increasing southward flow with stronger wind stresses from the northwest. The stronger northwest wind stresses also can have a disruptive influence on the major gyres in the Gulf of Maine.

Much importance has been attached to the gyres and their influence on the biology of the area. Attempts have been made to associate weaker year classes of fish species with the strength and duration of offshore winds (Chase, 1955). Other hypotheses rely on the mean circulation to provide a "retention mechanism" to help keep larvae in place (Iles and Sinclair, 1982). The complete integrated picture is yet to emerge. Any stabilizing forces inherent in the biology or provided by gyres have to be balanced against dispersive forces such as seasonal and transient winds, the influence of large features from outside the area, and even the large tidal excursion's important in generating the gyres.

6. Concluding remarks

1) The models presented here confirm that tidal forcing plays a major role in the generation of mean currents in the Gulf of Maine and Bay of Fundy.

2) The mean seasonal wind stresses or persistent steady uniform winds can modify the tidally forced mean circulation and give rise to other patterns such as the circulation in Massachusetts Bay with stronger wind forcing giving rise to more noticeable effects.

3) The Maine eddy is not completely generated by tides or seasonal winds, although tidally produced
residuals are consistent with that part of it north of Georges Bank. More than one process might be necessary to complete the eddy. Two possible candidates are the baroclinic effects of freshwater runoff and the effects of an inflow from the Nova Scotia shelf, neither of which has been investigated here.

4) Work is in progress with J. W. Loder, P. C. Smith and D. G. Wright to make further studies of large- and small-scale processes and delineate the limitations of the barotropic model. This will include an investigation of the influences of the open boundaries on the interior circulation and an examination of how the mean balances of the momentum equations change when different forcing is applied.

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APPENDIX

Details of the Computational Methods

The equations for both models were discretized in a semi-implicit manner on a Richardson lattice (Fig. A1). The continuity equation was the same in each case:

\[
\frac{u_{i,j}(t + \Delta t) - u_{i,j}(t)}{\Delta t} + u_{i,j}(t) \left[ \frac{u_{i+1,j}(t) - u_{i-1,j}(t)}{2\Delta x} \right] + r_{i,j}(t) \left[ \frac{u_{i,j-1}(t) - u_{i,j+1}(t)}{2\Delta x} \right] = f r_{i,j}(t) - g \left[ \frac{\xi_{i+1,j}(t + \Delta t) - \xi_{i,j}(t + \Delta t)}{\Delta x} \right] - k u_{i,j}(t + \Delta t) \left[ \frac{u^2(t) + r^2(t)}{2 \Delta x} \right] d_{i,j}(t) + A_h \left[ u_{i,j-1}(t) + u_{i,j+1}(t) + u_{i-1,j}(t) + u_{i+1,j}(t) - 4u_{i,j}(t) \right] + \frac{F_{i,j}}{\rho d_{i,j}(t)},
\]

(A4)

\[
\frac{v_{i,j}(t + \Delta t) - v_{i,j}(t)}{\Delta t} + s_{i,j}(t + \Delta t) \left[ \frac{v_{i+1,j}(t) - v_{i-1,j}(t)}{2\Delta x} \right] + v_{i,j}(t) \left[ \frac{v_{i,j-1}(t) - v_{i,j+1}(t)}{2\Delta x} \right] = -f s_{i,j}(t + \Delta t) - g \left[ \frac{\xi_{i,j+1}(t + \Delta t) - \xi_{i,j}(t + \Delta t)}{\Delta x} \right] - k v_{i,j}(t + \Delta t) \left[ \frac{v^2(t) + \xi^2(t)}{2 \Delta x} \right] e_{i,j}(t) + A_h \left[ v_{i-1,j}(t) + v_{i+1,j}(t) + v_{i,j-1}(t) + v_{i,j+1}(t) - 4v_{i,j}(t) \right] + \frac{G_{i,j}}{\rho e_{i,j}(t)},
\]

(A5)

\[
\frac{u_{i,j}(t + \Delta t) - u_{i,j}(t)}{\Delta t} + u_{i,j}(t) \left[ \frac{u_{i+1,j}(t + \Delta t) - u_{i-1,j}(t + \Delta t)}{2\Delta x} \right] + r_{i,j}(t) \left[ \frac{u_{i,j-1}(t + \Delta t) - u_{i,j+1}(t + \Delta t)}{2\Delta x} \right] = f r_{i,j}(t + \Delta t) - g \left[ \frac{\xi_{i+1,j}(t + \Delta t) - \xi_{i,j}(t + \Delta t)}{\Delta x} \right] - k u_{i,j}(t + \Delta t) \left[ \frac{u^2(t) + r^2(t)}{2 \Delta x} \right] d_{i,j}(t) + A_h \left[ \frac{u_{i,j+1}(t + \Delta t) + u_{i,j-1}(t + \Delta t) + u_{i+1,j}(t + \Delta t) + u_{i-1,j}(t + \Delta t) - 4u_{i,j}(t + \Delta t)}{\Delta x^2} \right] + \frac{F_{i,j}}{\rho d_{i,j}(t)},
\]

(A6)
where
\[ r_{i,j}(t) = \frac{1}{4}[v_{i,j}(t) + v_{i,j-1}(t) + v_{i+1,j}(t) + v_{i+1,j-1}(t)] \]  
(A8)
\[ s_{i,j}(t) = \frac{1}{4}[u_{i,j}(t) + u_{i-1,j}(t) + u_{i-1,j+1}(t) + u_{i,j+1}(t)] \]  
(A9)
\[ A_h \text{ is in the form used by Schwiderski (1980):} \]
\[ A_h = \frac{1}{2}a\Delta x \frac{d_i(t)}{v_{i,j}} \quad \text{for} \quad u_{i,j} \]
\[ A_h = \frac{1}{2}a\Delta x \frac{e_i(t)}{v_{i,j}} \quad \text{for} \quad v_{i,j} \]

We take \( a \), the reduced eddy coefficient, to be 1.64 \( \times 10^{-3} \) s\(^{-1}\). This formulation of \( A_h \) resulted in values from 3.5 \( \times 10^{7} \) cm\(^2\) s\(^{-1}\) in the deeper areas of the coarse grid to 1.4 \( \times 10^{8} \) cm\(^2\) s\(^{-1}\) in the shallower areas of the fine-grid region. The addition of this eddy viscosity to the model did not change the tides or qualitatively change the residual circulation. Indeed the circulation pattern is in qualitative agreement with the second model, where eddy viscosity is not included in the formulation. The residual currents, however, were slightly smaller, and grid-scale oscillation of currents was noticeably reduced, particularly in the deeper areas.

A value of \( k = 2.4 \times 10^{-3} \) was used for the friction coefficient in the coarse grid and \( k = 2.3 \times 10^{-3} \) was used in the Bay of Fundy. Changes in \( k \) larger than \( 1 \times 10^{-2} \) did affect the tidal calibration, but did not qualitatively change the residual pattern.

The second model (fine grid—Gulf of Maine) formulation of the momentum equation employed a scheme that reduces computer memory and computation-time requirements by centering in time the advective-term derivatives and the Coriolis term. This necessitated sweeping the grid in one direction for odd time-steps and in the opposite direction for even time-steps. The discretization for odd time-steps is as follows:

\[
\frac{v_{i,j}(t + \Delta t) - v_{i,j}(t)}{\Delta t} + s_{i,j}(t + \Delta t)\left[\frac{v_{i,j+1}(t + \Delta t) - v_{i,j-1}(t + \Delta t)}{2\Delta x}\right] + v_{i,j}(t)\left[\frac{v_{i,j-1}(t + \Delta t) - v_{i,j+1}(t + \Delta t)}{2\Delta x}\right] = -f s_{i,j}(t + \Delta t) - g \left[\frac{\xi_{i,j}(t + \Delta t) - \xi_{i,j+1}(t + \Delta t)}{\Delta x}\right] - k v_{i,j}(t + \Delta t)\left[\frac{s_{i,j}^2(t) + v_{i,j}^2(t)}{e_i(t)}\right]^{1/2} + \frac{A_h}{\Delta x^2}\left[\frac{v_{i,j+1}^2(t + \Delta t) + v_{i,j+1}(t + \Delta t) + v_{i,j+1}(t + \Delta t) - 4v_{i,j}^2(t + \Delta t)}{\rho e_i(t)}\right] + \frac{G_{i,j}}{\rho e_i(t)}, \quad (A7)
\]

\[
\frac{u_{i,j}(t + \Delta t) - u_{i,j}(t)}{\Delta t} = -\frac{u_{i,j}(t)}{2\Delta x} \times [u_{i+1,j}(t) + u_{i,j}(t + \Delta t) - u_{i-1,j}(t + \Delta t)] \\
- \frac{1}{4\Delta x} \left[\{v_{i,j-1}(t + \Delta t) + v_{i+1,j-1}(t + \Delta t)\right] \\
- [u_{i,j-1}(t + \Delta t) - u_{i,j}(t + \Delta t)] \\
+ [v_{i,j}(t) + v_{i+1,j}(t)][u_{i,j}(t) - u_{i,j+1}(t)] \\
+ f r_{i,j} - k u_{i,j}(t + \Delta t)\left[\frac{u_{i,j}^2(t) + r_{i,j}^2}{d_i(t)}\right]^{1/2} \\
- \frac{g}{\Delta x} \left[\frac{\xi_{i,j+1}(t + \Delta t) - \xi_{i,j+1}(t + \Delta t)}{\Delta x}\right] + \frac{F_{i,j}}{\rho d_i(t)}, \quad (A10)
\]

\[
\frac{v_{i,j}(t + \Delta t) - v_{i,j}(t)}{\Delta t} = -\frac{1}{4\Delta x} \left[\{u_{i,j}(t + \Delta t) + u_{i,j+1}(t)\right] \\
\times [v_{i+1,j}(t) - v_{i,j}(t)] \quad [u_{i,j}(t + \Delta t) + u_{i,j+1}(t)] \\
\times [v_{i,j}(t + \Delta t) - v_{i,j}(t + \Delta t)] \\
\times [v_{i,j}(t + \Delta t) - v_{i,j+1}(t + \Delta t)] \\
+ f s_{i,j} - k v_{i,j}(t + \Delta t)\left[\frac{s_{i,j}^2(t) + v_{i,j}^2(t)}{e_i(t)}\right]^{1/2} \\
- \frac{g}{\Delta x} \left[\frac{\xi_{i,j}(t + \Delta t) - \xi_{i,j+1}(t + \Delta t)}{\Delta x}\right] + \frac{G_{i,j}}{\rho e_i(t)}, \quad (A11)
\]
where

\[ r_{i,j} = \frac{1}{4} [v_{i,j-1} (t + \Delta t) + v_{i+1,j-1} (t + \Delta t) + v_{i,j} (t) + v_{i+1,j} (t) ] \]

\[ s_{i,j} = \frac{1}{4} [u_{i-1,j} (t + \Delta t) + u_{i,j} (t + \Delta t) + u_{i-1,j+1} (t) + u_{i,j+1} (t) ] . \]

The updated values, \( \xi(t + \Delta t) \), are obtained using the continuity equation. If we assume that the topmost current-component row is the \( u \) component, then Eq. (A10) is applied to that row of \( u \)'s. Eq. (A11) is then applied to the following row of \( v \)'s ( \( j \) is the same according to the Richardson lattice discretization shown in Fig. A1) followed by applying (A10) to the next row of \( u \)'s, and so on. In this way the grid is swept left to right, top to bottom. A similar discretization applies to the even time steps.

A value of \( k = 2.7 \times 10^{-3} \) was used for the frictional coefficient. As in the first model, small variations in the friction parameter did affect the calibration but did not change the residual current pattern qualitatively.

Along the open boundary, in both models, the elevation was specified to be a tidal sinusoid with zero mean. Amplitudes and phases were determined from observations. The current components defined between these elevation points were extrapolated by setting them equal to the value found immediately to the interior. The advective terms were excluded from the momentum equations when they would have included such an extrapolated current in their calculation. This effectively removed the advective terms from the momentum equation for calculations of the first \( u \) and \( v \) interior to the open boundary. Interior elevation points adjacent to two open boundary grid squares were smoothed using the simple filter:

\[ \tilde{\xi}_{i,j} = \frac{1}{9} (\xi_{i-1,j} + \xi_{i+1,j} + \xi_{i,j-1} + \xi_{i,j+1} + 4 \xi_{i,j}). \]

REFERENCES


EG&G, 1976: Forecasting power plant effects on the coastal zone. EG&G Environmental Consultants, Waltham, MA, 187 pp and Appendix.


