On the Continuity of Mean Flow between the Scotian Shelf and the Middle Atlantic Bight*

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ABSTRACT

Oxygen-isotope tracer data combined with results from two linear barotropic coastal models are used to argue that the observed equatorward mean alongshelf flow in the Middle Atlantic Bight is a downstream extension of the mean alongshelf flow over the Scotian Shelf. Qualitative agreement between model results and observations supports the concept that the alongshelf pressure gradient associated with the mean alongshelf flow in the Middle Atlantic Bight has an upstream or downstream and not an offshelf origin. The role of the local large-scale general circulation is apparently to help keep the shelf water on the shelf rather than to drive the shelf mean flow.

1. Introduction

The mean alongshelf flow in the Middle Atlantic Bight (MAB) is toward the southwest from Cape Cod toward Cape Hatteras and is generally opposed to the mean eastward alongshelf wind stress (Fig. 1). This has led to the conclusion that there must be a mean alongshelf pressure gradient associated with the mean alongshelf flow in the MAB (Stommel and Leetmaa, 1972; Csanady, 1976). Csanady (1978) and Beardsley and Winant (1979) have suggested further that local freshwater inflow is inadequate to drive the mean flow and that the alongshelf pressure gradient probably originates in the deep ocean as a boundary effect of the large-scale general circulation. (See Beardsley and Boicourt, 1981, for a review of this argument.) However, Shaw (1982) and Wang (1982) have questioned this driving mechanism by using a barotropic model to show that an alongshelf pressure gradient in the deep ocean does not penetrate shoreward of the shelf break but instead is largely confined to the continental slope region. The extent to which this result is altered by stratification is unclear and will be discussed later.

As an alternative, the mean alongshelf pressure gradient could originate entirely over the shelf (independent of the local deep-ocean circulation) as a result of some upstream or downstream forcing mechanism. For example, a large steady freshwater input poleward of the MAB might produce a strong density gradient leading to a mean alongshelf pressure gradient and, in turn, a mean alongshelf flow. Another possibility is that the Gulf Stream might entrain a sufficient quantity of shelf water near Cape Hatteras (e.g., Ford et al., 1952) to set up a mean alongshelf pressure gradient and an associated mean alongshelf flow. In either case, some continuity of flow from the Scotian Shelf, around or through the Gulf of Maine, onto the southern flank of Georges Bank and into the MAB would be required. While this idea is implicit in the regional circulation models dating back to Bigelow (1933), some direct evidence of flow continuity has only recently been obtained in observations of current, salinity and tracer distributions. Mean alongshelf flows of shelf water have been measured both entering the Gulf of Maine from the Scotian Shelf (Smith, 1983) and leaving the Gulf of Maine along the southern flank of Georges Bank into the MAB (Butman et al., 1982; Beardsley et al., 1985). Further, the estuarine character of the Gulf of Maine, i.e. the predicted deep inflow of saline slope water through the Northeast Channel, has been confirmed by Ramp et al. (1985).

Based on the available current observations, the mean alongshelf volume transport entering the Gulf of Maine appears to be about 0.4 Sv [1 Sv = 10⁶ m³ s⁻¹; about 0.14 Sv from the Scotian Shelf (Smith, 1983) plus 0.26 Sv through the Northeast Channel (Ramp et al., 1985)]. This is nearly equal to the mean alongshelf volume transport of 0.38 Sv observed (within the 120 m isobath) leaving the Gulf of Maine south of Nantucket Island and entering the MAB (Beardsley et al., 1985). Shorter-term measurements have been used to estimate a mean alongshelf transport of about 0.2 Sv within the 100 m isobath in the New York Bight and near Chesapeake Bay (Beardsley et al., 1976). These

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latter estimates would probably increase substantially if taken out to the 120 m isobath, suggesting that the mean alongshelf volume transport from Georges Bank through the MAB is roughly constant. This, combined with the oxygen-isotope and salinity distributions presented in section 2, forms perhaps the most convincing evidence of flow continuity from the Scotian Shelf through the Gulf of Maine and into the MAB.

The observed continuity of shelf water masses throughout this region does not by itself address the dynamics of such an implied mean alongshelf flow. On the contrary, Wang's (1982) model results suggest that the bulk of an alongshelf flow, such as that observed on the southern flank of Georges Bank, would move off the shelf well before reaching the southern end of the MAB, thus placing doubt on the idea of an upstream or downstream driving mechanism for the mean flow in the MAB. Further, it is unclear how the flow from the Scotian Shelf would traverse the Northeast Channel and the Gulf of Maine to reach Georges Bank. The roles, if any, of the large-scale general circulation and of the mean wind stress are also unclear. These dynamical questions are addressed here with the aid of two linear barotropic coastal models. In particular, in section 3 the boundary-layer model of Wang (1982) is reexamined using more realistic bottom friction and different diagnostic examples. Then, in section 4, the steady circulation of the MAB/Gulf of Maine region is explored with a more general barotropic numerical model. The roles of an upstream source, the large-scale general circulation and the mean wind stress are considered. A discussion of model results and observations appears in section 5, followed by a summary in section 6.

2. Oxygen-isotope tracer results

The use of the stable oxygen isotope $^{18}$O as a tracer in coastal waters of the MAB has been described by Fairbanks (1982). Basically, the H$_2$O/H$_2^{18}$O ratio in precipitation is temperature dependent resulting in the relative depletion of H$_2^{18}$O molecules at higher (i.e. colder) latitudes. The strong variation of $^{18}$O with latitude along the east coast of North America allows the use of $^{18}$O to distinguish between MAB river runoff, slope water and shelf water which originated at higher latitudes (i.e. an upstream source). Therefore, because the H$_2$O/H$_2^{16}$O ratio is a physical property of the sea water which is not changed by biological activity, it is an ideal oceanographic tracer in the MAB and can provide clear evidence of the origin of the MAB shelf water.

By convention, the isotope data are expressed as the fractional difference between the ratio $^{18}$O/16O in the sample to the ratio in standard mean ocean water (SMOW) (Craig, 1961):

$$
\delta^{18}O(\%o) = \frac{^{18}O/16O_{sample} - ^{18}O/16O_{SMOW}}{^{18}O/16O_{SMOW}} \times 1000.
$$

Salinity ($S$) and $\delta^{18}$O then provide two nearly conservative tracers with which to identify the origin of each water sample.

The data presented here are from water samples collected as part of the National Marine Fisheries Service Marine Resources Monitoring, Assessment, and Prediction (MARMAP) Program and were analyzed as described by Fairbanks (1982) at the Lamont-Doherty Geological Observatory. Data along six cross-shelf transects (labeled B–G on Fig. 1) were obtained during Cruise 81-03 of R/V Delaware II from 20 May to 18 June 1981 (Spring 1981). Data along transect A (Fig. 1) were obtained during Cruise 83-04 of R/V Albatross IV from 23 May to 22 June 1983 (Spring 1983). Additional samples along transect G were obtained during Cruise 81-14 of R/V Albatross IV from 16 November to 22 December 1981 (Fall 1981) and during Cruise 82-02 of R/V Albatross IV from 16 February to 25 March 1982 (Winter 1982).

Each shelf water sample is expected to consist of a mixture of slope water (SLW), Scotian Shelf water (SSW) and possibly local river runoff. Slope water is high in salinity ($S \approx 35.5\%o$) and is relatively enriched in $^{18}$O ($\delta^{18}$O $\approx 0.75\%o$). Both the salinity and $\delta^{18}$O of mean SLW may vary slightly on a seasonal or interannual basis ($\pm0.5\%o$ for salinity; $\pm0.3\%o$ for $\delta^{18}$O). Scotian Shelf water will be defined here as the water flowing around Nova Scotia into the Gulf of Maine. Using data from transect G near Nova Scotia (MARMAP station 172 located at 43°01'N, 66°20'W), SSW is defined by the depth-weighted average values of $S \approx 32.35\%o$ and $\delta^{18}$O $\approx -1.09\%o$ in spring 1981. Both salinity and $^{18}$O estimates of SSW changed somewhat to $S \approx 31.92\%o$ and $\delta^{18}$O $\approx -1.26\%o$ in fall 1981, and $S \approx 32.16\%o$ and $\delta^{18}$O $\approx -1.20\%o$ in winter 1982.

A plot of salinity versus $\delta^{18}$O for all samples along a transect typically confirms the expectation of mixing between SLW, SSW and (sometimes) local river runoff. For example, Fig. 2a is such a plot for transect D. The symbols correspond to the water depth at each station. Multiple symbols at each station represent samples from different depths. All of the points fall close to the line connecting SLW and SSW indicating simple binary mixing. Samples from shallower water tend to fall closer to SSW indicating a higher content of SSW. Similar plots using data from transects E, F and G are qualitatively identical. In Fig. 2b, data from transect A show evidence of mixing with a local freshwater source which is characterized by zero salinity and large depletions of $^{18}$O ($\delta^{18}$O $\approx -8$ to $-10\%o$). This local river runoff is mixed with shelf water in the Chesapeake Bay and then is discharged onto the shelf. Though limited to the nearshore region, the freshwater source creates a three-point mixing situation for some samples. Transects B and C are qualitatively identical to A.

The sample salinity and $\delta^{18}$O values have been used to estimate the percentage of each water type present in the mixture. Whenever the mixing was binary (no
obvious local freshwater contribution), the observed 
\((S, \delta^{18}O)\) point was projected perpendicularly onto the 
mixing line connecting SLW and SSW. The percentage 
of SSW is then the ratio between the distance from 
SLW to the projection and the distance from SLW to 
SSW \((\times 100)\). This method was also used along transects 
A, B and C for all points which fell very close to the 
mixing line between SLW and SSW. However, for data 
points along transects which obviously contained some 
local freshwater the procedure was different. First, a 
linear regression of all points which clearly deviated 
from the binary mixing line was used to estimate the 
\(\delta^{18}O\) value of the local freshwater source. Then the 
percentage of each water type was estimated as a three-
point mixing problem in which the mixture is required 
to conserve salt, \(\delta^{18}O\) and mass. It is important to limit 
the three-point mixing approach to those samples 
which obviously contained some local freshwater be-
cause, otherwise, all variability about the binary mixing 
line would be attributed to local freshwater influence 
even when there is no local freshwater present. This is 
unreasonable and can lead to inconsistent results.

The largest error involved in estimating the per-
centage of each water type comes from the estimate of 
SSW. Figure 2a shows the location of SSW for fall 1981 
and winter 1982 computed as for spring 1981. Both 
points fall essentially along the same mixing line as 
does the spring 1981 SSW point and all of the other 
data points. Thus, SSW is a mixture of SLW and a 
very high-latitude (low \(\delta^{18}O\)) water type, the percentage 
of which varies slightly in time. This introduces a 10–
15 percent uncertainty in the absolute percentage 
estimated by the present analysis. However, the error in 
the relative percentages from sample to sample on any 
one transect is smaller, probably less than 5 percent.

The percentage of SSW, computed as described 
above, has been contoured in a vertical section along 
each transect (several of which are displayed in Fig. 3) 
to show the continuity of SSW through the MAB. Transect G (Fig. 3a) cuts across the eastern tip of 
Georges Bank, over the Northeast Channel and onto 
the Scotian Shelf. The definition of SSW was obtained 
from the samples near Nova Scotia, hence the very 
high percentage of SSW at the left of Fig. 3a. The deep 
inflow of primarily SLW through the Northeast Chan-
nel is evident as a decrease in percent of SSW there 
(see section 5 for further discussion). This SLW inflow 
mixes with SSW (the details of which are not considered 
here) such that the water which approaches the south-
ern flank of Georges Bank (either leaving the Gulf of 
Maine or flowing past the Northeast Channel) contains 
70–80 percent SSW. Only the deeper water located off-
shore of the shelf break contains predominantly SLW. 
Transect F (not shown) is very similar to transect G. 
Transect E (Fig. 3b) also shows the SLW influence at 
depth in the Gulf of Maine. The well-mixed water over 
Georges Bank is still 70–80 percent SSW while the SLW 
is still located primarily offshore. Transect D (not
shown) is located within the MAB and still shows shelf water which is 70–80 percent SSW. Again, the SLW contribution is minor except in the deep water offshore of the shelf break.

Local freshwater influence is not obvious until transect C (Fig. 3c) is reached. Here and in transects B (not shown) and A (Fig. 3d), local river runoff (dashed contours) is evident only in nearshore and shallow waters. The percentages are fairly small reaching a maximum of ~15 percent in transect A. Of greater importance is the high percentage of SSW found over most of the shelf; 70–80 percent in C, 60–70 percent in B and A. This tracer distribution, when combined with the roughly constant mean alongshelf volume transport discussed in section 1, demonstrates that:

(i) most of the shelf water found throughout the MAB flowed into the region from the Scotian Shelf,

(ii) local river runoff within the MAB is concentrated in a relatively narrow nearshore, nearsurface band, and

(iii) the cross-shelf mixing of shelf water with slope water is relatively weak in comparison with downstream advection through the MAB.

Estimates of cross-shelf mass exchange based on the oxygen-isotope tracer data will be presented elsewhere after the complete MARMAP dataset has been analyzed.

3. Boundary layer model for the MAB

While the results of the previous section demonstrate the continuity of flow between the Scotian Shelf and the MAB, the dynamics of such a flow are still unclear. In fact, Wang’s (1982) results suggest that the flow would leave the shelf before reaching the MAB. Therefore, in this section, we use the simple model of Wang to study in more detail the channeling of the upstream inflow onto the continental slope and the effects of more realistic bottom friction.

Consider the steady flow of a homogeneous ocean along an infinite, straight coast with bottom topography $h(x)$ varying only in the offshore ($x$) direction. The alongshelf ($y$) flow is geostrophic, there is no surface wind stress, and bottom friction is parameterized by a linear friction coefficient $r(x)$. The depth-averaged equations of motion are

\begin{align}
-f'v &= -g'x, \\ f'u &= -g'y - rv/h, \\ (uh)_x + (vh)_y &= 0,
\end{align}

where $(u, v)$ are the cross-shelf ($x$) and alongshelf ($y$) depth-averaged velocities, $f$ the Coriolis parameter, $g$ gravitational acceleration, $\xi$ surface elevation, and subscripts denote partial differentiation. Equations (3.1) may be combined into a single equation for the surface elevation:

\begin{equation}
\xi_{xx} + \frac{r}{r} \xi_x + \frac{f h}{r} \xi_y = 0.
\end{equation}

At the coast, the transport normal to the coast must vanish: $uh = 0$ which leads to

\begin{equation}
\xi_y + \frac{r}{r} \xi_x = 0 \quad \text{at} \quad x = 0.
\end{equation}

The surface elevation is required to vanish far from the coast:

\begin{equation}
\xi \to 0 \quad \text{at} \quad x \to \infty.
\end{equation}

After prescribing an alongshelf flow at the upstream boundary, the equivalent surface elevation [$\xi = \xi_0(x)$ at $y = 0$] is computed, so that Eqs. (3.2)–(3.4) define a boundary-value problem which is solvable by standard finite-difference methods.\footnote{Wang (1982) solved (3.2)–(3.4) with constant $r$ in a two-dimensional, finite-difference domain using Gaussian elimination. He was, thus, forced to impose a boundary condition at the downstream boundary ($y = -y_{max}$). Here we use the heat analogy (Csanyi, 1978).}
For the present calculations, the bottom topography (in m) is that used by Wang (1982):

\[
h(x) = \begin{cases} 
20 + 0.001x, & 0 \leq x \leq x_0 \\
160 + 0.05(x - x_0), & x_0 \leq x \leq x_1 
\end{cases}
\tag{3.5}
\]

where \(x_0 = 1.4 \times 10^5\) m is the location of the shelf break and \(x_1 = 1.8 \times 10^5\) m the location of the offshore boundary of the model domain. (The results described below are virtually identical when \(x_1\) is increased.) The bottom friction coefficient is expected to increase in shallow water due to the combined effects of currents and surface gravity waves (Grant and Madsen, 1979). In order to include this effect in a simple way, the bottom friction coefficient is assumed to have the form

\[
r(x) = \frac{a}{h(x) + h_0} + b,
\tag{3.6}
\]

where \(a, b\) and \(h_0\) are arbitrary constants. The present choice of \(a = 300 \text{ cm}^2 \text{ s}^{-1}\), \(b = 0.015 \text{ cm} \text{ s}^{-1}\) and \(h_0 = 2000 \text{ cm}\) represents a compromise among three sources: (i) the Grant and Madsen (1979) model for a surface gravity wave with a 15 s period and a 1 m amplitude over a 10 m thick bottom boundary layer with a 15 cm s\(^{-1}\) rms velocity at the top of the boundary layer; (ii) estimates of bottom friction made by Winant and Beardsley (1979) and Noble et al. (1983) at several shelf locations in the MAB; and (iii) estimates of bottom friction over the continental slope in the MAB made using recent long-term current statistics (Butman, personal communication, 1985) and computed as in Wright and Thompson (1983).

To illustrate the channeling of an upstream inflow onto the slope region, Wang (1982) chose \(r_0\) to represent a uniform 10 cm s\(^{-1}\) inflow over the entire shelf and slope. He also used a constant bottom friction coefficient of \(r = 0.1 \text{ cm} \text{ s}^{-1}\) [\(a = 0\) and \(b = 0.1 \text{ cm} \text{ s}^{-1}\) in (3.6)] which is larger over most of the shelf (for depths greater than 15 m) than the variable bottom friction values used here. Wang’s large alongshelf inflow over the slope region makes it difficult to determine how much of the shelf flow actually moved out over the slope. Therefore, the flow field has been computed here assuming a uniform upstream inflow of 10 cm s\(^{-1}\) over the shelf only (zero inflow over the slope) using both constant \(r (=0.1 \text{ cm} \text{ s}^{-1}\), Fig. 4a) and variable \(r\) (Fig.
4b). Contours of the alongshelf velocity are shown in a plan view in the upper panel. The corresponding cross-shelf dependence of both the alongshelf velocity and the alongshelf transport at various alongshelf (downstream) locations is shown in the lower panels.

With the large bottom friction used by Wang (1982), the inflow moves toward the shelf break within a relatively short distance (Fig. 4a). The flow then moves onto the slope, but stays close to the shelf break due to the large bottom slope there. The transport shows a peak over the slope which moves offshore from the shelf break with distance downstream. With variable bottom friction (Fig. 4b) the flow is not rapidly channeled onto the slope, but instead largely remains over the shelf for realistic alongshelf distances.

This difference is primarily a result of the smaller bottom friction in the deeper water. Physically, bottom friction allows the flow to cross isobaths (i.e., with no bottom friction, all shelf flow would remain on the shelf). Over some of the shelf, the bottom friction term in (3.1b) becomes large and is balanced by the Coriolis term such that the imposed inflow (−v) requires a seaward velocity component (+u) in the Coriolis term. As the depth increases, the friction term becomes smaller until it can no longer balance the Coriolis term. The flow must then follow the isobaths more closely. This occurs at a shallower depth with smaller bottom friction, and consequently the shelf flow cannot move off the shelf as easily.

A secondary effect of variable bottom friction is to "advect" surface elevation offshore. That is, viewing (3.2) as a steady advection–diffusion equation with unit diffusivity, then the second term in (3.2) is like an advection term with a cross-shelf "velocity" equal to −r_{z} /

Fig. 4. Boundary layer model results with (a) constant r and (b) variable r. In each figure: upper panel shows contours of alongshelf velocity (cm s−1) in plan view; middle panel shows cross-shelf profile of alongshelf velocity at y = 0 (solid), −250 km (dotted), −500 km (dashed); −750 km (dash-dot); lower panel shows cross-shelf profile of alongshelf transport at y = 0 (solid), −250 km (dotted), −500 km (dashed); −750 km (dash-dot). Dashed vertical line denotes location of shelf break.
\( r (r_x < 0 \text{ here}). \) This "advection" is important near the coast where \( r_x/r \) is large, but much less important near the shelf break and over the slope. As a result, there is a convergence of \( \xi \) over the shelf such that the alongshelf velocity may exceed the inflow velocity (e.g., the velocity at \( y = -250 \) km in the middle panel of Fig. 4b).

The effect of the different bottom frictions is emphasized again in Fig. 5 where the percent of the inflow which has remained over the shelf is plotted versus alongshelf distance. More of the inflow tends to remain over the shelf for larger distances with smaller and variable bottom friction (solid curve) than with larger and constant bottom friction (dashed curve). To put the alongshelf coordinate \( (y) \) into perspective, the distance between Georges Bank and Cape Hatteras is \( O(900 \) km), so according to this simple model, only \( \sim 20 \) percent of a uniform inflow at Georges Bank would move off the shelf before reaching the New York Bight \( (\sim 400 \) km) and \( \sim 60 \) percent of the inflow would still be on the shelf at Cape Hatteras. Further, the flow which does leave the shelf would remain close to the shelf break.

4. Middle Atlantic Bight, Gulf of Maine model

The model examined in the previous section was extremely idealized. The results suggest that the idea of an upstream or downstream forcing mechanism for the MAB mean flow may be reasonable, but the details have yet to be resolved. For a more realistic study of the fate of a steady upstream inflow along the Scotian Shelf, a more general barotropic model than that considered in section 3 has been developed.

a. Numerical model

As in section 3, the steady flow of a homogeneous ocean is considered. However, the coastline, bottom topography and bottom friction coefficient are allowed to vary in both \( x \) and \( y \), invalidating the previous assumption of geostrophic alongshelf flow. Equation (3.1a) is thus replaced by

\[
-f_0 = -g \xi_x - ru/h.
\]

The bottom friction coefficient \( r \) is treated as a scalar. A transport streamfunction is defined by

\[
 uh = -\psi_y, \quad vh = \psi_x
\]

and substituted into (4.1, 3.1b, 3.1c) to obtain

\[
 \psi_{xx} + \psi_{yy} - c^x \psi_x - c^y \psi_y = 0,
\]

where

\[
 c^x = \left( \frac{r_x}{r} - \frac{f h_x}{h} - \frac{2 h_x}{h} \right),
\]

\[
 c^y = \left( \frac{r_y}{r} + \frac{f h_y}{h} - \frac{2 h_y}{h} \right).
\]

If (4.3) is viewed as a steady advection–diffusion equation with unit diffusivity, then \( c^x, c^y \) represents a divergent background velocity field which "advects" the streamfunction.

Numerical solutions of (4.3) have been obtained using the upstream differencing method (nonflux form) of Fdieiro and Veronis (1977) which reduces the implicit diffusion associated with standard finite difference schemes. The solution domain and bottom topography (Fig. 6a) represent a realistic but smooth version of the MAB/Gulf of Maine region. A grid with 84 alongshelf grid spaces and up to 40 crossshelf grid spaces is used with \( \Delta x = \Delta y = 12.74 \) km. The shelf break generally occurs at about 100 m depth and a rapid offshore depth increase to about 2000 m followed by a gradual depth increase to 3000 m at the offshore boundary. Bottom friction in the Georges Bank area is dominated by tidal currents which creates somewhat larger estimates of bottom friction coefficients than those used in the boundary layer model of section 3 (Noble et al., 1983). However, for simplicity and consistency with the remainder of the MAB/Gulf of Maine region, the bottom friction coefficient given by (3.6) will be used here with the same constants as in section 3.

The coastal boundary condition of no normal flux is satisfied by making the coast a streamline, \( \psi = \psi_T \). The model is driven by a specified inflow transport at the upstream boundary \( (y = 0) \) and/or at the offshore boundary \( (x = W) \). An upstream inflow is specified as an alongshelf depth-averaged velocity at each grid space which then determines the spatial distribution of \( \psi \) along that open boundary [integrating (4.2) from the coast where \( \psi = \psi_T \)]. At the offshore boundary, an inflow is specified as a pressure gradient along the boundary. Since bottom friction at 3000 m depth is negligible, then (3.1b) becomes \( fu = -g \xi_y \) which, combined with (4.2), yields

\[
\psi_{y} = \frac{fu}{g}.
\]
FIG. 6. MAB/Gulf of Maine model with standard bottom friction and no wind stress. (a) Bottom topography, coordinate system and transect locations. Transport streamlines for the cases of (b) shelf inflow only and (c) combined shelf inflow, deep upstream inflow and deep offshore inflow ($\zeta_y \neq 0$), normalized by the total shelf inflow transport. Thicker, unlabelled contour is 200 m isobath. Inflow velocity structure is shown at upstream boundary.
\[ \psi_y^w = \frac{gh}{f} \zeta_y^w \quad \text{at} \quad x = W, \quad (4.4) \]

where \( \zeta^w \) and \( \psi^w \) are \( \zeta \) and \( \psi \) at \( x = W \). After determining \( \psi^w \) at the upstream edge of the offshore boundary \((x = W, \ y = 0)\) from the specified inflow at \( y = 0 \), then (4.4) may be integrated to determine \( \psi^w \) along the offshore boundary. Note that if \( \zeta_y^w = 0 \) is specified in (4.4), then the offshore boundary becomes a transport streamline, i.e., there is no transport across the offshore boundary. At the downstream open boundary a smoothness condition is imposed, \( \psi_{yy} = 0 \). The effect of this boundary condition is limited to a narrow region near the downstream boundary (\(~10\) grid spaces).

b. Circulation driven by shelf inflow

As a generalization of the results of section 3, the model is first driven solely by an inflow over the Scotian Shelf. Based loosely on the observed mean flow over the Scotian Shelf (Smith, 1983), the inflow velocity is assumed uniform within 50 km of the coast and zero otherwise. The pressure gradient along the offshore boundary is assumed zero (\( \zeta_y^w = 0 \)). Because the model is linear, the inflow velocity may be scaled to give any desired total transport through the model. For the present case, a velocity of \( v = -10 \text{ cm s}^{-1} \) is chosen so that the resulting total transport (0.456 Sv) is roughly equivalent to the observed mean total transport entering the Gulf of Maine around Nova Scotia (\(~0.14\) Sv from Smith, 1983) and through the Northeast Channel (\(~0.26\) Sv from Ramp et al., 1985). This is clearly an oversimplification, but it is the simplest way in which the deep inflow through the Northeast Channel can be included in the present barotropic model.

The resulting transport streamline pattern is shown in Fig. 6b. The streamfunction has been scaled by the total shelf inflow transport, \( \psi_f = 0.456 \text{ Sv} \). Thus, \( \psi = 1 \) at the coast, \( \psi = 0 \) at the offshore boundary and 10 percent of the total transport occurs between any two streamlines. The mean flow generally tends to follow the local isobaths. The nearshore inflow at \( y = 0 \) spreads out somewhat before reaching the Gulf of Maine. About 70 percent of the flow enters the Gulf of Maine directly while 30 percent crosses over the Northeast Channel without really entering the Gulf of Maine. The flow is counterclockwise within the Gulf of Maine. About 10–20 percent of the total transport leaves the Gulf of Maine and enters the MAB via the Great South Channel and/or over Georges Bank. The remainder flows in a narrow jet along the southern side of the Northeast Channel, turns sharply southwestward and flows along the southern flank of Georges Bank toward the MAB. This inflow on the right and outflow on the left of the Northeast Channel (looking toward the Gulf of Maine from the slope region) is consistent with the observations of Ramp et al. (1985). The flow in the MAB roughly follows the local isobaths while spreading offshore due to the influence of bottom friction.

The model alongshelf transports perpendicular to transects H–K (Fig. 6a) and the transect topographies are shown in Fig. 7. The approximate total transport over the shelf (at depths less than 150 m) is indicated by the open bars. At transect K, the downstream shelf transport (0.23 Sv) occurs mainly in a narrow jet near the shelf break and represents about 50 percent of the total inflow transport. Note that the portion of the total inflow which leaves the Gulf of Maine via the Great South Channel region does not show up in transect K. There is evidence (on the left) of the eastward jet along the northern flank of Georges Bank. At transect J, the transport through the Great South Channel (0.09 Sv) is evident on the left while the shelf-break jet transport (0.16 Sv) is still present on the right. The total alongshelf transport over the shelf (Great South Channel plus southern flank of Georges Bank) is \(~55\) percent of the total inflow. Note that roughly 30 percent of the jet in transect K has moved off the shelf before reaching transect J. At transect I, the shelf transport (0.23 Sv) has broadened somewhat and includes \(~50\) percent of the total inflow. The location of transect I is nearly coincident with the location of the Nantucket Shoals Flux Experiment (NSFE) array (Beardsley et al., 1985). The observed mean alongshelf transport during NSFE was 0.38 \pm 0.07 Sv which is somewhat larger than the present model estimate of 0.23 Sv. At transect H, the alongshelf transport (0.17 Sv) is again mostly located near the shelf break and includes \(~37\) percent of the total inflow.

The structure of the model currents is compared to the observed mean currents from the NSFE (Beardsley et al., 1985) in Fig. 8 (a and c). The currents are plotted such that alongshelf and equatorward is down while onshelf is to the left. The observed currents are based on one-year records taken at the locations of the solid dots in the bottom panel. The depth of each instrument is in parentheses beside its vector. The numerical model is barotropic so only a single vector is drawn. The modeled and observed current structures compare favorably, both containing a region of high velocity located near mid- to outer shelf. There also tends to be an offshore component of flow over the shelf in both the observed and modeled currents. The model current magnitudes are considerably smaller than the observed currents which is expected since the modeled transport is also reduced from the observed transport.

Overall, these results suggest that, while some of the Scotian Shelf inflow remains on the shelf through the MAB, this inflow alone is not adequate to account entirely for the observed mean flow.

c. Role of the deep ocean

The possibility of the local large-scale general circulation directly driving the observed shelf mean flow
has been questioned (Shaw, 1982; Wang, 1982). However, its effect on the shelf flow resulting from an upstream shelf inflow (section 4b) has not been explored. There are two means by which the large-scale general circulation can be included in the present model. First, a deep inflow transport (seaward of the shelf break) can be imposed at the upstream boundary. Second, a deep inflow transport can be imposed through the offshore boundary by specifying a nonzero pressure gradient along that boundary ($\xi_y^{*} \neq 0$).

The observed depth-averaged mean flow seaward of the shelf break (out to the 3000 m isobath) is generally equatorward as part of the recirculating cyclonic gyre described by Hogg (1983). A crude estimate of the along isobath transport across 70°W longitude and shoreward of 3000 m is 7–8 Sv (based on Fig. 2 from Hogg). A similar estimate can be made at the approximate location of the model upstream boundary using more recent observations (Hogg, personal communication, 1985) which suggests a deep transport of about 3–4 Sv into the model domain. This suggests an apparent inflow of about 4 Sv through the offshore boundary between the upstream boundary and the Nantucket Shoals region (~70°W). Using (4.4), this inflow is equivalent to a sea surface slope of $\xi_y^{*} \approx 0.2 \times 10^{-7}$.

To model these deep inflows, the upstream velocity was specified as in section 4b, but now with an additional deep velocity of $-3$ cm s$^{-1}$ at the offshore edge of the upstream boundary which decreases linearly to zero 76 km shoreward of the offshore boundary. The equivalent total deep upstream inflow transport is 3.5 Sv. In addition, the pressure gradient along the entire offshore boundary is fixed at $\xi_y^{*} = 0.2 \times 10^{-7}$ which imposes a total inflow of 7 Sv uniformly distributed along the offshore boundary.

The resulting transport streamline pattern is shown in Fig. 6c. The streamfunction has been scaled by the total shelf inflow transport, $\psi = 0.456$ Sv. Thus, $\psi = 1$ at the coast and 10 percent of the shelf inflow transport occurs between any two solid streamlines. This scaling allows direct comparison of streamlines between Figs. 6b and 6c. The flow pattern in Fig. 6c is qualitatively similar to that in Fig. 6b especially in the Gulf of Maine/Georges Bank region. The major difference is that the shelf flow ($0 < \psi < 1$) remains almost entirely shoreward of the 200 m isobath throughout the MAB. Thus, while the deep inflow does not generally move onto the shelf, it apparently prevents the shelf transport from moving off the shelf.

Physically, the imposed upstream inflow corresponds to a positive pressure gradient in the alongshelf ($y$) direction. Since the bottom friction term in (3.1b) is negligible in the deep ocean, this pressure gradient must induce a shoreward geostrophic velocity component ($-u$). Thus, the streamlines move toward the shelfbreak where bottom friction becomes important and where the shelf flow has a seaward velocity component. Together these forces balance with the result that the shelf
flow does not leave the shelf and the deep flow does not move onto the shelf.

This deep-ocean influence is shown explicitly in Fig. 7 by the shaded bars which represent the additional shelf inflow transport which remains on the shelf due to the deep-ocean inflow transport. Note that the shaded bars do not generally represent deep water which has moved onto the shelf, but instead they represent shelf water which did not leave the shelf because of the deep-ocean effect. The deep-ocean influence increases with distance away from the source simply because the amount of shelf water which would have left the shelf increases with $-y$ also. The model transports over the shelf are now: 0.29 Sv at transect K; 0.11 Sv through the Great South Channel and 0.25 Sv over the shelf at transect J; 0.59 Sv at transect I; 0.59 Sv at transect H. Very little shelf water leaves the shelf. In fact, some deep water has moved onto the shelf at transects I and H. The model transport across transect I (0.59 Sv) is now considerably larger than the observed NSFE mean transport (0.38 Sv).

The model current structure at transect I still compares favorably with the observed mean currents from the NSFE [Fig. 8 (b and c)]. The deep-ocean inflow transport has not greatly changed the shelf currents except to increase their magnitude. The model currents near the shelf break and over the slope, however, are much larger with a strong onshore component. (These large model currents near the shelf break account for the excess model transport over the shelf.) Nevertheless, the present results suggest that while the large-scale general circulation may not drive the shelf mean flow, it may still play an important role in the MAB steady circulation by keeping the shelf flow on the shelf.

d. Sensitivity tests

The present numerical model could obviously be adjusted so as to produce the observed NSFE mean transport at transect I. However, in this study the real interest is in qualitatively determining the dominant dynamical balances rather than reproducing observations. Thus, the sensitivity of the model to changes in the dynamics must be examined to ensure that the above qualitative results and ideas are robust and not fortuitous. With this in mind, the flow simulation of section 4c is taken to be the standard case, and four variations from this standard have been computed: 1) the upstream shelf inflow was removed; 2) the offshore pressure gradient was removed ($\xi_y = 0$); 3) bottom friction was doubled ($a = 600 \text{ cm}^2 \text{ s}^{-1}, b = 0.030 \text{ cm} \text{ s}^{-1}$); and 4) a spatially uniform wind stress of $\tau_u = 0.5 \text{ dyn cm}^{-2}$ was applied in the $+x$ direction. The first two tests allow the relative importance of each feature to be ascertained. Doubled bottom friction essentially encompasses the highest estimates of bottom friction in the MAB made by Noble et al. (1983). The wind stress is roughly equivalent in both magnitude and direction to the observed mean wind stress (Beardsley and Boicourt, 1981). To apply the constant wind stress, the term $\tau_u/\rho h$ must be added to the right-hand side of (4.1) where $\rho$ is the water density. This results in the term $\tau_u/\rho h$ being added to the right-hand side of (4.3).

In the first test, the transport streamline pattern (not shown) shows the deep flow being inhibited by the slope and not penetrating very far onto the shelf as expected. In tests (2)–(4), the transport streamline patterns (not shown) are qualitatively (and nearly quantitatively) identical to the standard case result (Fig. 6c). A more detailed comparison is presented in Fig. 9 where currents along transect I are plotted. The crosses mark the ends of the corresponding vectors from the standard case (Fig. 8b). The removal of the upstream shelf inflow (Fig. 9a) greatly reduces the shelf currents yet hardly changes the deep currents. This shows the "insulating" effect of the slope. The shelf currents are not negligible, however, as might have been expected. The removal of the deep-ocean pressure gradient (Fig. 9b) has almost no effect on the currents at transect I indicating that the deep upstream inflow is far more important. It should also be noted that the deep geostrophic velocity component induced by the imposed pressure gradient is always shoreward regardless of the presence of the offshore boundary. This suggests that the use of a solid
Fig. 9. Sensitivity of MAB/Gulf of Maine model currents along transect I in the standard case (Fig. 8b) to: (a) shelf inflow removed, (b) deep offshore inflow removed (\(\tau^{*} = 0\)), (c) doubled bottom friction, and (d) spatially uniform wind stress, \(\tau^{*} = 0.5 \text{ dyn cm}^{-2}\). Crosses show location of arrowheads for the standard case (Fig. 8b). Shelf break is indicated by fat arrows.

The length of each bar represents the percent of the total shelf inflow shoreward of the 100 m isobath which would not be on the shelf if the respective mechanism were removed. The effect of both deep inflows (upstream and offshore) increases with distance from the upstream boundary. The deep upstream inflow effect is much larger than the offshore inflow effect (consistent with Fig. 9). The shelf inflow effect decreases with distance from the upstream boundary.

Figure 10b shows the effects of bottom friction and wind stress at various alongshelf locations. Increased bottom friction lowers the percentage of shelf transport everywhere and has an accumulating effect in the downstream direction. The wind stress effect is small and decreases with the downstream direction. Its greatest effect is in the Georges Bank region where the wind acts over a larger area and generates more alongshelf Ekman transport.

The overall conclusion from these tests is that none of them significantly changes the model results of the offshore wall probably has little effect on the results. Doubled bottom friction (Fig. 9c) reduces the velocities by about 20 percent at midshelf depths with almost no change in direction. The reduction in magnitude is small in shallow water, but there is an additional frictionally induced offshore velocity component. In deep water, the effect is negligible. The uniform wind stress (Fig. 9d) has little effect on current magnitudes, but it induces an offshore component of velocity everywhere with greatest effect in shallow water where \(\tau^{*}/\rho h\) is largest. This is because the imposed uniform wind stress has a significant alongshelf component at transect I.

The effects of these changes on shelf transport at various alongshelf locations are presented in Fig. 10. In both panels, the abscissa is the grid space along the offshore boundary with zero being the downstream end, while the ordinate is the percentage of the shelf inflow transport which is shoreward of the 100 m isobath. Figure 10a shows the relative influence of the deep-ocean pressure gradient \(\tau^{*}\) (solid), the deep upstream inflow (shaded), and the shelf upstream inflow (open).

Fig. 10. Influence of various model components at different alongshelf locations in the MAB/Gulf of Maine model. Ordinate is percentage of shelf inflow transport which is still within the 100 m isobath. Abscissa is grid number in the y direction (some reference locations are labeled). (a) Relative influence of shelf inflow (open bars), deep upstream inflow (shaded bars) and deep offshore inflow (solid bars). (b) Effect of doubled bottom friction (dotes) and uniform wind stress, \(\tau^{*} = 0.5 \text{ dyn cm}^{-2}\) (diamond) on the transport for the standard case (crosses).
standard case (section 4c) or the comparisons with observations, suggesting that the qualitative results concerning the relative importance of the upstream shelf and deep inflows in maintaining the shelf mean flow are quite robust.

5. Discussion

Although the results presented here are fairly self-consistent, several aspects of the present study deserve further discussion. For example, the absence of stratification and mixing processes in the models could be a serious shortcoming. There are known regions of strong stratification (e.g., the shelf/slope front, the northern flank of Georges Bank, the Gulf of Maine in summer) and seasonal variations in stratification, but their influence on the mean flow is unclear. Stratification could, for example, allow a deep-ocean alongshelf pressure gradient to penetrate onto the shelf and drive shelf currents. Some supporting theoretical evidence for this process has been presented recently by Csanady (1985). Further, flow over the relatively steep slope region is expected to be quite complicated in the presence of stratification. In addition, tidal rectification, which may be an important contributor to the mean circulation around Georges and Browns Banks, has been neglected. The model flow along the northern flank of Georges Bank (Fig. 6b) would certainly be strengthened by including tidal rectification. Without more complicated models with which to compare, it can only be hoped that the present model has isolated the dominant physics of the mean flow along the southern flank of Georges Bank and the MAB.

At first glance, it might seem peculiar that there has been no mention of whether or not the model alongshelf pressure gradient is sufficient to drive the mean flow or how it compares with that inferred from data (Scott and Csanady, 1976). However, this is a somewhat meaningless question because the existence of a mean flow of the proper order of magnitude in the numerical model (without wind driving) implies the existence of an alongshelf pressure gradient of the proper order of magnitude as well. This follows from the fact that the dynamical balance along a streamline must be the along-streamline pressure gradient balancing the along-streamline bottom friction because there can be no flow across streamlines. And, if the streamlines roughly follow the local isobaths, then the alongshelf (along-isobath) pressure gradient must be sufficient to drive the alongshelf (along-isobath) flow. This also suggests that estimates of an alongshelf pressure gradient which are based solely on current observations do not provide evidence for the origin of the alongshelf pressure gradient.

The oxygen-isotope data have been treated essentially in a synoptic sense with little discussion of temporal variations on monthly and seasonal time scales. The qualitative picture is probably not altered much by these variations. For example, oxygen-isotope data along transect G during fall 1981 and during winter 1982 (not shown) are qualitatively identical to Fig. 3a. The percent SSW over the Northeast Channel is less in fall 1981, consistent with larger observed inflow of deep water in fall and winter (Ramp et al., 1985), but the other contours and percentages of SSW over Georges Bank are almost unchanged. On the other hand, the definition of SSW changes seasonally (Fig. 2a) which means that during spring 1981, each water sample along the MAB came from a slightly different source of SSW depending on the travel time between the Scotian Shelf and the specific location in the MAB. Thus, comparing all spring 1981 MAB samples to the spring 1981 definition of SSW is inaccurate. And, of course, comparing transect A in spring 1983 with the spring 1981 version of SSW is equally inaccurate. As more oxygen-isotope data become available, these temporal errors may be eliminated to obtain a more quantitatively accurate version of Fig. 3 or its equivalent.

The high percentage of SSW within the Northeast Channel (Fig. 3a) may seem surprising. However, it appears to be consistent with the salinity observations of Ramp et al. (1985) who often found salinities of 34.5 to 34.9% in the deep inflow. Because of the binary nature of the mixing process there, either salinity or δ18O alone can be used to estimate the percentage of SSW. Thus, the observed deep inflow apparently contains roughly 20–30 percent SSW (using SLW = 35.5%, SSW = 32.35%) by the present definition, which is consistent with the range of estimates of percent SSW in the deeper parts of the Northeast Channel based on the present oxygen-isotope data. One possible explanation is that the mixing of shelf and slope water may have occurred upstream over the Scotian Shelf and upper slope and was advected into the Northeast Channel. This scenario is supported by (i) mean flow estimates which suggest that the water which enters the Gulf of Maine directly around Nova Scotia comes from relatively nearshore along the Scotian Shelf (Smith, 1983; Drinkwater et al., 1979), and (ii) estimates of mixing over the Scotian Shelf which suggest that the nearshore waters are not mixed appreciably with slope water whereas the outer shelf water appears to mix strongly with slope water even at substantial depths (Houghton et al., 1978). Thus, it may be this mixture of shelf and slope water which flows into the Northeast Channel at depth and accounts for the high percentage of SSW there. However, a full understanding of the details of the mixing and flow in this region must await future studies.

6. Summary

Some evidence has been presented for the idea that the observed mean flow in the MAB is a downstream extension of the observed mean flow on the Scotian
Shelf. The evidence may be summarized in the following points:

- The oxygen-isotope tracer data clearly indicate that the shelf water found in the MAB is largely of Scotian Shelf origin. For the data presented, the MAB water consisted of about 70–80 percent Scotian Shelf water. There was only slight mixing with the deeper slope water and little influence of local river runoff.
- The boundary-layer model of the MAB suggests that an alongshelf flow might remain mostly on the shelf and near the shelf break even after traveling a distance equivalent to the distance from Georges Bank to the New York Bight.
- The results from the MAB/Gulf of Maine steady circulation model, driven by an inflow from the Scotian Shelf and a fairly realistic deep inflow, compare reasonably well with several aspects of observed currents and transports. Further, the results seem to be fairly insensitive to changes in bottom friction, to an imposed wind stress, or to an imposed deep-ocean pressure gradient.

Based on these results and the roughly constant observed mean alongshelf transport through the MAB, the mean flow in the MAB is apparently not driven by an alongshelf pressure gradient induced by the local large-scale general circulation! Instead, the alongshelf pressure gradient originates either upstream or downstream of the MAB/Gulf of Maine region, and the role of the local large-scale general circulation is to help prevent the shelf water from leaving the shelf. Thus, the shelf behaves almost like a two-dimensional channel with relatively little exchange occurring across the shelf edge.

It is important to note that the ultimate source of the upstream inflow and the true forcing mechanism for the MAB mean flow have not been identified. In fact, the present results provide no way of determining whether the forcing mechanism originates upstream, downstream, or both. For example, the numerical model in section 4 could be driven by an outflow specified at the downstream boundary instead of the specified inflows. If an outflow is specified only over the shelf, then about 10–20 percent of the resulting transport in the interior occurs over the shelf, while the remainder is drawn onto the shelf from the deep-ocean within 15 km of the downstream boundary. Thus, a shelf outflow alone is not sufficient to drive the mean alongshelf flow. However, if the specified outflow includes the deep-ocean outflow equivalent to that resulting from the standard case (section 4c), then the interior flow is nearly identical to the standard case. Although this result supports the importance of the deep-ocean circulation, it does not identify the origin of the alongshelf pressure gradient. The real point here is that the driving mechanism is not local or offshore but rather upstream or downstream or both.

The mixing lines implied in Fig. 2 suggest that the ultimate upstream source is probably linked to a freshwater source located well upstream from the Scotian Shelf (δ 18O ≈ −20‰) which is consistent with the results of Fairbanks (1982). Such large depletions of 18O in freshwater sources are found along the coasts of the northern Labrador Sea, Baffin Bay and Greenland. Furthermore, deep Labrador Sea water apparently has an 18O content similar to slope water near the MAB (Fairbanks, 1982) which again seems to rule out the deep ocean as an upstream driving mechanism. However, more 18O measurements and modeling will be necessary to identify the ultimate source and to determine the true driving mechanism(s).

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REFERENCES

Drinkwater, K., B. Petrie and W. H. Sutcliffe, Jr., 1979: Seasonal


