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1. Introduction

This project is a component of the SEEP-II program to study shelf-slope exchange in the southern Middle Atlantic bight (MAB). It represents the physical oceanographic portion of the SEEP-II research at Lamont-Doherty Geological Observatory (L-DGO). Since the work consists of two parts: data analysis and theoretical modeling, this report will be divided into two parts to describe the progress of each activity. It covers work performed during the time interval March to December 1990 and is a sequel to the report submitted in February 1990.

2. Data Analysis

2a. Scope of the Analysis

Although L-DGO has long ago completed processing and distributing data from the instruments that L-DGO contributed to the SEEP-II field program, not all of the SEEP-II data is yet available. Therefore, our analysis has been confined to data from the first two deployments in spring and summer of 1988.

In the course of this analysis there have been constant checks for consistency among the data. In several cases there still is uncertainty in the compass reading associated with some of the bottom-mounted acoustic Doppler current meters. Until this is resolved we have concentrated our analysis efforts on that portion of the data set for which we have high confidence.

In last year's report much of the initial data analysis contained contoured sections or individual time series to illustrate specific events and processes in a qualitative sense. We have now gone beyond this stage and can present quantitative results of flux calculations. For this, the availability of vertical profiles of temperature and velocity has proved invaluable and represents a significantly new contribution that the SEEP-II data set makes toward studying coastal processes. We expect the benefits of this data set to become increasingly evident as analysis continues.

To orient the reader of this report we repeat Figures 1 and 2 of the previous report which shows the SEEP-II array and the instrumentation deployed on the northern line. In this report we will concentrate on the results of two lines of investigation. These are the heat and mass flux across the north line. Our results will focus on mooring 3 the only location for which we have complete temperature and velocity profiles for both deployments.

Figure 1: Detail bathymetry in the region of the SEEP-II mooring array. Mooring positions are indicated by circled numbers.

Figure 2: Schematic diagram of the instrumentation on the northern portion of the SEEP-II array.

A continuous record of the fluctuating thermal structure across the northern line is given in Figure 3. Here the bottom temperature at the indicated mooring site is contoured as a function of time. Some of this fluctuation corresponds to the local cross-shelf motion of the water column and some to alongshore advection. For example, in February and early March (YD 45-75) the fluctuations are wind driven upwelling and downwelling events. In April (YD 100-110) a massive wind driven downwelling event pushed shelf water out into the slope region. The mid-shelf temperature minimum during May and early June (YD 130-145 and 155) represent the formation of the cold pool. The reappearance of this cold bottom water in October (YD 275-285) is due to a southward advection of cold pool water. We will attempt a quantitative estimate of the flux resulting from some of these events.

2b. Heat Flux

The heat flux is calculated from the product, $T \cdot u$ where u is the cross-shelf velocity. If we set $T = T_0 + T'$ and $u = u_0 + u'$ then $T \cdot u = T_0 \cdot u_0 + T' \cdot u' + T_0 \cdot u' + u_0 \cdot T'$. We have made a considerable effort to test the robustness of this calculation. If T_0 and u_0 are record means then the total record time average $\langle T \cdot u \rangle = T_0 \cdot u_0 + \langle T' \cdot u' \rangle$ since the last two terms are identically zero. However, for averages over intervals less than the full record length these last two terms will make a contribution. In order to minimize this contribution and to accommodate the seasonal trends in temperature data, the data are fitted to a fourth order polynomial when computing T' and u' .

Since u is very sensitive to the definition of the cross-shelf direction we tested in detail how sensitive the flux calculations are to changes in this direction. In the initial processing of data on the northern line the alongshore axis was set to that of the local isobaths, 30° east of north. For the spring deployment the depth average flow was within 0.5 degrees of this axis. Since this difference is within the compass accuracy the depth average cross-shelf flow is taken to be zero. During the summer deployment the depth average flow was 6° onshore of the isobath axis. This difference could be due to compass error, the procedure used to extrapolate the flow field from the sensors to the surface and bottom of the water column, or indicate that there is a real net cross-shelf flow at the northern line. Until the compass uncertainty is resolved we have defined U by the axis for which the cross-shelf flow is identically zero.

The accuracy with which the depth average flow is defined depends in part on how we extrapolate the velocity through the water column. At mooring 3 the current meters provide direct measurements within 5m of the bottom and 11m of the top. Within the water column the 5m resolution given by acoustic Doppler profile gives for the first time, a reliable velocity profile. We

Figure 3: Bottom temperature across the shelf as a function of time. Vertical axis is the distance in km from mooring 1. Arrows indicate mooring position for the data used. Time scale is yearday with tick marks to indicate the 1st of each calendar month.

have not used a specific model to extrapolate the velocity to the top and bottom of the water column because of the potential errors that can be introduced. For instance, if the surface flow is modeled from the measured wind stress, it would neglect the flow induced by other forces that contribute to the flow measured at the uppermost current meter. As a compromise and to minimize the uncertainty we have simply extrapolated the velocity of the bottom and uppermost current meter to the respective surface when calculating the depth average flow or heat flux.

As mentioned in our previous report, there are significant differences in the flow field at the mid-shelf, mooring 2, and the outer shelf edge, mooring 3. This is illustrated in Figure 4 where we plot low-passed u vs v for February and March (YD 45-95) when the wind driven fluctuations are most vigorous. Note the distinct difference at the bottom of mooring 2 and 3. At 2 the cross-shelf flow (u) is highly correlated with the alongshore flow (v) in an Ekman sense with veering of approximately 18° . At 3 the correlation is much weaker and there is a net flow offshore and no large excursions of flow onshore.

We noted in last year's report that at mooring 2 the cross-shelf isotherm displacements shown in Figure 3 during February and March are in phase and of the same magnitude as the water column displacements. Thus u' and T' are in quadrature (out of phase by 90°) and the eddy heat flux $\langle T' \cdot u' \rangle$ will be zero. This is illustrated in the eddy flux calculations, shown in Figure 5. At mooring 2 the large values of $u' \cdot T'$ produced by the wind driven fluctuations in February average to zero. At mooring 3 there is a much larger and statistically significant non-zero eddy heat flux. A negative value of $T' \cdot u'$ corresponds to a flux of heat onshore, i.e. down gradient. Co-spectra of unfiltered u' and T' (Figure 6) shows that the eddy flux is predominantly at frequencies lower than 1 cycle per day. The cross-spectra (Figure 7) shows that at these low frequencies the phase difference between u' and T' is closer to 90° at mooring 2 than mooring 3. We conclude that the wind driven motions over the mid-shelf result in large quasi-adiabatic fluctuations in the temperature field which result in little net cross-shelf eddy heat flux. A conclusion consistent with SEEP-I results. In contrast at mooring 3, which is at the shelf break slightly seaward of the mean location of the front, the flow and thermal fluctuations are more complex and result in a significant eddy heat flux.

We are able to calculate the cross-front eddy flux in the following way. We use the bottom temperature at mooring 4 and along the northern line to define the orientation of the frontal boundary as a function of time. Using this axis we then calculate the cross-frontal component of the velocity at mooring 3 and use it to calculate a cross-frontal eddy heat flux. The results are

Figure 4: Low-passed cross-shelf (u) vs. alongshelf (v) velocity components for mooring 2 (left) and mooring 3 (right).

Figure 5: Eddy heat flux $T' \cdot u'$ ($^{\circ}\text{C cm/s}$) at the bottom of mooring 2 (top) and mooring 3 (bottom).

Also shown is T' ($^{\circ}\text{C}$), u' (cm/s), and the accumulative average $\overline{T' \cdot u'}(t) = \frac{1}{t} \int_0^t (T' \cdot u') dt$. Record mean heat flux and standard deviation is listed at top.

SEEP II, RCM2254 M2/55/90
9-bin avg., TE. vs. UC

SEEP II, RCM8410 M3/85/90
9-bin avg., TE. vs. UC

cycles/day
Period (days)

Figure 6: Co-spectra of T and u at the bottom of mooring 2 (top) and mooring 3 (bottom).

SEEP11, MOORING2 at 55 M
TE vs. UC, 9-band averaged

SEEP11, MOORING3 at 85 M
TE vs. UC, 9-band averaged

Figure 7: Cross-spectra of T and u at the bottom of mooring 2 (left) and mooring 3 (right).

SEEPH MOORING 3 AT 85 M -- TE & UC MEAN REMOVED
 Temperature Flux Mean -0.746 ± 10.178

Figure 8: Same as Figure 3 (bottom) except heat flux is calculated along the cross-frontal axis.

shown in Figure 8 and are comparable with the cross-shelf eddy heat flux. Higher in the water column the lateral temperature gradients are weaker so the axis of the front is not as well defined.

We now look at the heat flux throughout the water column at mooring 3. In Figure 9 we show $\langle T_0 \rangle \langle u_0 \rangle$ and $\langle T' \cdot u' \rangle$ as a function of depth for the three separate deployments. Note that the scale for $\langle T_0 \cdot u_0 \rangle$ is an order of magnitude larger than that of $\langle T' \cdot u' \rangle$ and the two are comparable only when depth averaged. The vertical structure of $\langle T' \cdot u' \rangle$ changes markedly with season. In the winter it is maximum near the foot of the front in the summer it is greatest just below the seasonal thermocline. The summertime flux values, both the eddy and the total $T \cdot u = T_0 \cdot u_0 + T' \cdot u'$, are greatest in the summer; again consistent with SEEP-I results.

This seasonal change in the vertical structure of $\langle T' \cdot u' \rangle$ is not simply a function of the change in the variance of T' and u' . Although in the summer wind stress diminishes markedly and cross-shelf displacements of the front are not discernable the standard deviation of the cross-shelf velocity σ_u decreases by a factor of 2 at the bottom and approximately 50% in the upper water column. The standard deviation of the temperature σ_T remains the same at the bottom and increases by a factor of 2 at the top and a factor of three in the middle of the water column (25-45m depth). At all depths the seasonal change in the heat flux $\langle T' \cdot u' \rangle$ is greater than the change in the product $\sigma_T \cdot \sigma_u$, so the changes in the eddy heat flux arise from changes in the phase of T' and u' and not just their magnitude.

The time dependence of the vertical structure of $T' \cdot u'$ is shown in Figure 10 where the data are further low passed to make the contours more distinct. As evident in Figure 5, most of the heat flux onto the shelf during the spring occurred during the April event. During the summer the locus of the depth of maximum eddy heat flux rises in the water column then descends in the fall to complete the annual cycle. This is again illustrated in Figure 11 which shows low passed time series of $T' \cdot u'$ at 11m, 47m, and 85m depths and depth averaged for the entire SEEP-II experiment. Note how episodic the heat flux events are. These results demonstrate the necessity of time series data spanning the entire water column in order to study coastal exchange processes.

The heat flux values that we calculate compare favorably with earlier bulk estimates made by R. Fairbanks at L-DGO. Using oxygen isotope data Fairbanks estimated a yearly mean depth average shelf-slope water mass exchange of $0.2 \text{ m}^2/\text{s}$ per unit alongshelf length in the MAB. Applying this to the 100m water column at the shelf break yields an effective exchange velocity $V_d = 0.2 \text{ cm/s}$. Assuming a temperature difference across the front of $\Delta T = 4^\circ\text{C}$ this implies a heat flux $V_d \Delta T \approx -0.8^\circ\text{C cm/s}$. It is this number that compares with our heat flux calculation. The

Time Average

Depth Average

$$\langle T_0 \rangle \cdot \langle u_0 \rangle + \langle T' \cdot u' \rangle = \langle T \cdot u \rangle$$

Figure 9: Vertical structure of time average heat flux for the 3 SEEP-II deployments: the eddy flux $\langle T' \cdot u' \rangle$ heavy line and the mean flux $\langle T_0 \rangle \cdot \langle u_0 \rangle$ light line. Also listed are depth average values of the total flux $T \cdot u = \langle T_0 \rangle \cdot \langle u_0 \rangle + \langle T' \cdot u' \rangle$. Note that scale for $\langle T \rangle \cdot \langle u \rangle$ is order of magnitude larger than the scale for $\langle T' \cdot u' \rangle$. Negative values indicate heat flux onshore.

Figure 10: Vertical structure of $T \cdot u'$ (low-pass filtered) as a function of time for the 3 SEEP-II deployments.

SEEPH MOORING3 EDDY FLUX 4th order trend removed T&U
rotations: -0.2 (spring), -6.43 (summer), -8.1 (fall)

Figure 11: Low-passed eddy heat flux at 11m (top), 42m (middle), 85m (bottom), and depth averaged. Units are $^{\circ}\text{C cm/s}$.

SEEP-II data is able to provide now information on the time dependence and the vertical structure of this flux.

It is interesting to compare the magnitude of this heat flux with heat content changes over the shelf. The heat added to the shelf is given by the divergence of the eddy flux. For a value of $u' \cdot T' = 1^\circ\text{C cm/s}$ and assuming that the flux that passes mooring 3 is mixed over the inner shelf we find that

$$\frac{\partial T}{\partial t} = \frac{\partial}{\partial x} (u' \cdot T') \approx 0.2 \times 10^{-5} \text{ }^\circ\text{C/s} \approx 0.17 \times 10^{-1} \text{ }^\circ\text{C/day}.$$

This is equivalent to a surface flux of 40W/m^2 . Since the magnitude of both this temperature change and surface heat flux are small compared to observed values, it is clear that the cross-front heat flux is not the major contribution to the shelf water heat budget.

However these calculations are very important for quantifying shelf-slope exchange. To the extent that nutrients and other dissolved constituents are correlated with temperature, we can use the eddy heat flux to infer the eddy exchange of these constituents, and for these constituents the eddy flux may be the dominant term in their budget calculations. In collaboration with D. Wallace at BNL we are attempting to estimate cross-shelf nutrient fluxes. For example, using the wintertime temperature-nitrate regression a heat flux of 0.6°C cm/ corresponds to a nitrate flux of $0.6 \times 10^{-3} \text{ } \mu\text{mol/cm}^2/\text{s}$. This nutrient input could support an export of $0.1 \text{ gC/m}^2/\text{day}$ which is roughly 10% of the total springtime production. An equivalent summertime calculation will be more difficult since during this time temperature and nitrate are not correlated by a simple regression.

The heat flux calculations are also useful to identify the relevant exchange processes at the shelf edge. From SEEP-I we had concluded that during the winter the front was a barrier to exchange and that exchange rates increased during the summer in the upper portion of the water column. To date our results generally confirm these conclusions with important additions. The location of the mooring used in the SEEP-I calculations is roughly equivalent to SEEP-II mid-shelf mooring 2 where we also find an absence of a net heat flux across the bottom in spite of vigorous wind driven cross-shelf motion. SEEP-II results suggest that the fluxes in the frontal environment may be significantly different. Certainly the flow field such as the u, v correlation is different. The fact that most of the heat flux occurs at time scales greater than one day may be important

when evaluating the possibility that internal wave dissipation in the frontal zone contributes to the heat exchange there. It is clear from our results that most of the eddy heat exchange occurs during a few large-scale events.

Our results also show a clear seasonal progression in the structure and magnitude of the eddy exchange processes. During the summer wind-driven motions of the front diminish but with the formation of the seasonal pycnocline isopycnal surfaces traverse the frontal boundary in the upper portion of the water column. The enhanced heat exchange that we calculate at this depth reinforces our SEEP-I conclusions that these isopycnal surface are important to the shelf-slope exchange. Frontal eddies have been suspected as a primary process in this exchange. However, measurements tracking a single coherent eddy over a two-week period during SEEP-I by Dr. R. Garvine at the University of Delaware did not show any enhanced heat exchange. In contrast, our calculations, although spatially constrained, examine averages over longer time intervals, and these do show enhanced net heat exchange. Whether this arises from the existence of frontal eddies is yet to be determined.

2c. Mass Exchange

During April 1988 there was a major wind driven event in which a strong southward flow (Figure 12) was accompanied by an offshore flow of shelf water. During SEEP-II there is no other event with an equivalent exchange of mass between shelf and slope water regions, so much of our analysis has focused on this event. Since the SEEP-II sediment traps on the slope had a large accumulation during this time (Biscaye, personal communication) the potential for this event to exchange shelf and slope water to export particulate material needs to be carefully analyzed.

Temperature and velocity records throughout the entire SEEP-II array suggest that the April event was coherent over large alongshelf distances and that the cross-shelf isotherm displacements shown in Figure 3 represent local cross-shelf flow and not the advection of an upstream thermal feature. The cross-shelf flow at the bottom of mooring 3 (Figure 13) has an integrated offshore displacement of more than 60 km. From Figure 3 it is clear that this bottom flow does not continue offshore as a continuous feature for then shelf water would be seen far down the slope. Figure 14 is a repeat of Figure 3 but instead uses the temperature at 115m depth at mooring 6. Here we find evidence of shelf water moving out into the slope region at mid depth on an approximately constant density surface.

VC, MOORINGS 2, 3 (DASHED), and 4 (BOLD)
 Tops are VMCM units, Bottoms RCM (2,3) and ADCP (4) bins

Figure 12: Alongshore velocity (v) at mooring 2, 3, and 4. Middle panel shows the depth average flow. Units in cm/s.

Figure 13: Cross-shelf displacement of bottom water (85m depth) at mooring 3 from an integration of u .

SEEPH MOORINGS 1 - 6 at the Bottom

Figure 14: Temperature-time contours calculated from bottom temperature (40m, 60m, 90m, 125m at mooring 1, 2, 3, 5 respectively) and 115m at mooring 6.

This clarifies another observation. In SEEP-I and SEEP-II we find seaward of the 60 m isobath a mean Eulerian flow offshore in the bottom boundary layer. In SEEP-II, along the northern line, this flow is approximately 2 cm/s even down to 400m on the upper slope region. This is only compatible with the mean density structure if in a Lagrangian sense the bottom flow was exported on density surfaces and is not a continuous flow along the bottom.

Analysis of the April event suggests an even more complex flow at the shelf break. The depth average cross-shelf displacement is shown in Figure 15. Note that for mooring 3 the net cross-shelf flux for the entire April event is approximately zero, i.e. as much water moves offshore at the beginning of the event as moves onshore at its conclusion. However, notice that for the bottom most current meter (Figure 13) the flow is offshore with no subsequent onshore flow even though Figure 3 shows that warm slope water temperatures are quickly reestablished.

To examine the bottom flow in more detail we performed the following calculation. We integrate the cross-shelf flow (u) from the bottom up to the depth where the sign of u changes. In Figure 16 we show the resulting mass flux, thickness of the layer over which u is integrated, the average velocity in this layer, and the total flux averaged from the beginning of the time series as a function of time. Since the net depth averaged cross-shelf flow is approximately zero, there must be an equivalent onshore flow elsewhere in the water column. The mean mass exchange from February to June implied by this calculation is $0.46 \text{ m}^2/\text{s}$ or about twice Fairbank's result. However, note that the wind driven fluctuations in February and March yield no net mass exchange along the bottom and that more than half of the mass exchange occurred during the April event.

More details of the vertical structure of the cross-shelf flow during the April event are shown in Figure 17. Here the cross-shelf displacement (the integral of u) for five depths is plotted as a function of time. At the beginning of the April event surface flow is onshore and the bottom is offshore as expected for a downwelling event. However return flow is in the interior of the water column and not the bottom. The mid-depth return flow begins not with the reversal of the alongshore flow from the south to north but when the southward flow begins to diminish. This sequence is illustrated by the profiles in Figure 18 which show the cross-shore displacement (integrated u) for different time intervals. Water motions at the bottom are rectified, it moves offshore but not back onshore an equivalent distance. The onshore return flow is predominantly at mid-depth.

SEEP II, MOORINGS 2, 3 (DASHED) AND
Progressive Displacement, Barotropic Mean

Figure 15: Depth average cross-shelf displacement of the water column (integration of u).

Figure 16: Bottom mass flux (see text for explanation) at mooring 3.

Figure 17: Cross-shelf water displacement (integration of u) for various depths at mooring 3.

Figure 18: Vertical structure at mooring 3 of the horizontal displacement (integral of u) for different time periods associated with the April event. Positive values are offshore displacements. Tick mark denotes the displacement for $u = 5$ cm/s. Alongshore velocity also denoted.

We have begun to investigate with Creighton Wirick at BNL how the transmissometer and fluorometer data are consistent with the observed flow pattern. Indeed chlorophyll a is seen to move off the shelf at the bottom and out to mooring 6 in the interior of the water column. Subsequently the chlorophyll a is found in non-shelf water suggesting mixing in the frontal region.

These results are particularly interesting in light of recent numerical results by Glen Gawarkiewicz and Dave Chapman at WHOI. When stratification is added to their frictional model they get convergence at the foot of the shelf-slope water front. The Lagrangian trajectories show bottom shelf water moving offshore leaving the bottom boundary layer in a complex 3-dimensional flow pattern. In SEEP-II there is evidence that shelf water leaves the bottom boundary layer when it moves out into the slope region. Comparing the smaller cross-shelf displacement at 115m depth at mooring 6 (Figure 19) with the displacement at mooring 3 (Figure 13) during the April event suggests a convergence in the frontal zone. The correspondence between these model predictions and the SEEP-II observations is an exciting development that will be pursued in our subsequent analysis.

Our analysis to date suggests that the circulation at the shelf break contributes a significant exchange and mixing of shelf and slope water. Most of the mixing and exchange appears to have occurred during the single event in April. It remains to be seen whether this event is qualitatively different or just stronger than other wind driven events. During the summer shelf-slope exchange occurs higher in the water column. It is thought to be the result of sub-meso scale frontal eddies but whether this is consistent with the observed flow patterns has yet to be established.

3. Modelling

In last year's report, we have discussed some preliminary results from a model that examines the interaction of a separated Gulf Stream with coastal boundary in the Middle Atlantic Bight (MAB). In the past year, we have considerably expanded the scope of that model to improve its application to the observation and to further our physical understanding. Specifically, we have a) considered the case of a backward jet, or a jet approaching convergent coastal geometry, as is the case just south of the Cape Hatteras; b) incorporated topography which modifies the flat-bottom solutions in important ways; and c) included bottom friction which allows us address the shelf circulation as driven by the Gulf Stream. The following discussion provides a self-contained and more complete account of our modelling effort.

SEEP II, MOORING 6 PVD, 115M

Figure 19: Cross-shelf displacement (integral of u) at 115m depth at mooring 6.

3.1 The Motivation

Observations show that Gulf Stream interacts with shelf and slope water in a rather complicated and varied fashion after it leaves the continent. From satellite images, one clearly sees the entrainment of ambient water by the Gulf Stream. This entrainment has been suggested as the main exit route of the southward flowing shelf water near Cape Hatteras and is corroborated by the presence of low salinity water of shelf origin (the "Ford" water) along the northern edge of the Gulf Stream. In addition to this entrainment, satellite images also show occasional peeling off of the Gulf Stream water from its northern edge not unlike the shingling observed in the Southern Atlantic Bight (SAB). Hydrographic and current meter data in SEEP-II have documented episodes of intrusion of Gulf Stream water, apparently narrow, onto the outer shelf, in support of the satellite observations. Recently, Churchill (1990) has identified an additional mode of circulation by which a vast expanse of Gulf Stream water, but with little kinetic energy, resides over the slope sea, extending to well north of the mean path of the Gulf Stream.

It is obvious that to interpret these observation, one needs some dynamical understanding of the Gulf Stream interaction with its ambient water confined by a coastal boundary, an understanding however is totally lacking for want of proper dynamical models. Specifically, we want to address the following questions: What causes these disparate circulation patterns? What is the underlying physical balance? And how do they depend on the property of the Gulf Stream and/or the geometry and topography of the coastal boundary? It is the intention to address these questions and gain some dynamical understanding of the problem that we undertake this modelling effort.

The model configuration is schematically shown in Figure 20 in which the x-axis is aligned with the maximum velocity and the coastal boundary has the form of $y \approx x^\alpha$ with α being an external parameter. For the following discussion, we shall only be concerned with the case of positive α so that the coastal geometry is divergent for a forward jet. As plotted in Figure 20, the coastline changes from being concave to convex as α changes from less to greater than one. To facilitate physical understanding, we shall first consider the case of a flat bottom and forward jet, and then discuss its extension to include the case of backward jet, non-flat bottom and bottom friction.

Figure 20: The model configuration with x -axis aligned with the maximum velocity and a coastline varies as x^α . Cases for three different values of α are plotted.

3.2 Flat Bottom and Forward Jet

For a minimal model, we shall consider the vorticity balance between advection and lateral diffusion, in which case, the only external parameter, besides that associated with the coastal geometry, is the Reynolds number, essentially a measure of the relative importance of the advective to diffusive effects. As a good approximation, we assume that the streamwise scale is much greater than the spanwise scale so that similarity solution can be obtained which provides an analytical representation of the possible flow fields. Since typically, Reynolds number is much greater than one, we shall only be concerned with this case in the following discussion, although solutions can be obtained for small Reynolds number as well.

As plotted in Figure 21, the flow fields differ qualitatively for different coastal geometry and in particular, the values of $\alpha = 1/2$ and 1 mark the boundary of different dynamical regimes. For α smaller than $1/2$ (Figure 21a), the jet is broad and extends to outside the coastal boundary layer where significant shear occurs, being pressed against the coast by an accelerating flow. When α is greater than $1/2$, the diffusive effect is no longer confined to the coastal boundary layer but permeates through the interior, resulting in a sharpened jet, accompanied by reversed flow shoreward of the jet. For a concave coastline (Figure 21b), the reversed flow is entrained into the jet, while for a convex coastline (Figure 21c), the reversed flow is fed from the jet (detrainment).

To infer the flow field in the MAB, one notes that the coastline exhibits different similarity form depending on the scale of interest (Figure 22). On a smaller scale in the vicinity of the Cape Hatteras, the coastline curves convexly, but when the MAB is viewed as a whole, it curves concavely. As a result, the model solution suggests possible detrainment of the Gulf Stream water near Cape Hatteras, occurring over smaller scales, consistent with observations. On the other hand, over the larger scale, if the coastline is sufficiently divergent, one expects the Gulf Stream to entrain the ambient water, resulting in a circulation that resembles the Slope Sea Gyre. Thus through this simple model, we have isolated the effect of coastline geometry in controlling the circulation and shown that it may cause the contrasting entraining and detraining circulation observed in the MAB.

3.3 Backward Jet

The model, as formulated, can be used to examine the case when the offshore jet is backward, or the flow is approaching a converging coastline, as is the case just south of Cape Hatteras. For α greater than $1/2$ (Figure 23b and c), the jet is accelerating downstream and, as

Figure 21: Streamlines for a flat-bottom ocean and forward jet, for three different coastal geometry.

Figure 22: A schematic of Gulf Stream induced circulation in the MAB based on model results.

Figure 23: Same as Figure 21, but for a backward jet.

expected from previous discussions, the jet is broad and extends to near the coast where significant shear occurs. But when α is smaller than $1/2$ (Figure 23a), although the jet is decelerating, one does not recover the dynamical regime of oscillatory flow in the interior as is the case for a forward jet. But instead, one obtains a new dynamical regime in which the flow remains unidirectional with a friction boundary layer pressed against the jet itself by an offshore advection of the vorticity. This is a boundary layer of large vorticity gradient and hence is less visible than the coastal boundary layer where large velocity shear occurs. It nevertheless is dynamically important in that it allows an inviscid interior and hence more predictable flow field.

But since the flow remains unidirectional regardless of the value of α , the derived flow does not compare well with that observed just south of Cape Hatteras where there are pronounced shingling accompanied by flow reversals. One suspects that topography may be an important missing element of the model. Furthermore, as Gulf Stream separates, it enters the deeper water, how would the stretching of the vortex tube affect the resulting circulation? These considerations lead us to expand the original model to include the topography.

3.4 Non-flat Bottom

With finite topography and hence vortex stretching in the vorticity balance, one may obtain similarity solution only for a particular coastal geometry, namely $\alpha = 1/3$. But being concave, it is nevertheless representative of the MAB as a whole and also applies to the coastal geometry just south of the Cape Hatteras. It thus allows us to address the questions raised above.

Some qualitative effects of topography can be easily inferred which can be corroborated by analytical solutions. As Gulf Stream enters the deeper water, the vortex stretching tends to induce positive vorticity. This would sharpen the jet and hence favor the flow reversal farther inshore. We have plotted in Figure 24 a schematic of the similarity solution which employs the same coastal geometry and Reynolds number as that of Figure 21a but with isobaths as indicated by thin dashed lines. The solution shows that even if the coastal geometry is only slightly divergent, so long as there is vortex stretching, one would produce flow reversals. It suggests that the Slope Sea Gyre depicted in Figure 22 is indeed the norm of the circulation because of the added effect of topography and does not depend strongly on the degree of concaveness of the coastline as deduced from the flat-bottom ocean.

Since in a linear regime, cases of forward and backward jet are symmetric, one expects similar modification of the flow field as the jet shoals. That is, the jet is sharpened and flow

Figure 24: Same as Figure 21a, but with isobaths indicated by thin dashed lines.

Figure 25: Same as Figure 23a, but with isobaths indicated by thin dashed lines.

reverses inshore as shown in Figure 25. Because of the accompanying detrainment, one then produces shingle-like features resembling observation in the SAB.

The above discussion suggests that topography is important in reproducing some observed features. But without bottom friction which is important over the shallow shelf, the coastal boundary referred to so far should be identified with the shelf break and the vanishing flow along such boundary is justified by its implicit spin down by the bottom drag. To extend the model domain to include the shelf, one needs to incorporate the bottom friction as the next logical extension of the model.

3.5 Bottom Friction

With the inclusion of a bottom drag, one is further constrained in the choice of external parameters. In this case, it is the downstream variation of the bottom depth, which must vary as $x^{1/3}$. But since it corresponds to a deepening bottom in the positive x-direction, it is representative of a typical topography associated with continental boundaries. As the spanwise distribution of the depth is not constrained, we can then incorporate a shelf topography which would then allow us to address the shelf circulation as driven by the Gulf Stream. This is the task we are presently engaged in and hence cannot yet report on the results, but we anticipate important implication to be deduced regarding the driving of the shelf flow.

END

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