Sources of Variability in Gulf of Maine Circulation

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Variability in the circulation of coastal oceans must ultimately be driven by changes in the meteorological conditions which force currents in the coastal ocean, and by variability in the waters entering the coastal ocean from elsewhere. If a coastal ocean is to be accurately understood and modeled, the external forcing which drive the largest portions of the circulation variability must be observed adequately. Thus some method of comparing the relative importance of various sources of circulation variability must be developed, so that there is confidence that the most important sources of variability are included in any analysis or modeling. This is done for variability in the time-integrated transport.

The relative importance of various sources of circulation variability in the Gulf of Maine are then quantified, with an emphasis on variability on times longer than tidal or weather-band timescales. It is found that the variability forced by fluctuations in the winds and the inflow from the Scotian Shelf to the Gulf of Maine produce roughly comparable amounts of circulation variability. However, changes in the density structure of the GoM produce changes in time-integrated transport that are an order of magnitude larger, at least in the central Gulf of Maine. The changes in the large-scale density gradients that cause circulation variability are governed by mixing processes in the Gulf and by changes in the water masses entering the Gulf of Maine. Unless the density and transport of the waters entering the Gulf of Maine are routinely observed, numerical models will fail to capture much of the variability in the circulation of the Gulf. An analysis is given of the minimal set of observations needed to allow numerical models of the Gulf of Maine to adequately resolve the true variability in the circulation.

There is an increased desire to understand the sources of interannual variability the ecosystems of coastal oceans, and thus of the variability of the ocean circulation that govern these ecosystems. In order to understand the circulation variability, the relative importance of the sources of this variability must be understood. In the analysis below, the relative importance of various sources of circulation variability in the Gulf of Maine (GoM) is quantified. The analysis will show that many current models fail to capture the majority of the variability in the circulation on the ecologically and societally relevant timescales of a few months and longer. The models do not fail because of any inherent flaw, but because observations are currently unable to adequately constrain and force these models adequately.

The transports driven by the currents in the GoM control many important ecological processes, and the examination of circulation variability below will focus on time and space scales of broad ecological interest. The transport of near surface waters from the northeast GoM to the southwest and then onto Georges Bank move copepods from the gulf where they can successfully reproduce onto the bank where they are food for economically important larval fish [Hannah et al., 1998]. Transport along the coast of Maine can move
harmful algal bloom species along the coast and drive toxic blooms in the western GoM
[Franks and Anderson, 1992]. Many economically important benthic species have planktonic larval stages which are dispersed by the currents. The population dynamics and the retention of these populations is thus strongly influenced by the variability in the currents, integrated over the time the larvae are in the plankton. (Byers and Pringle [2005], Pringle and Wares [2005]). In each of these three examples, what matters is not the instantaneous value of the currents, but the net transport the currents drive when integrated over an ecologically relevant timescale. Thus the variability in these three ecological systems, and many other similar biological systems, will depend on the variability in these integrated transports. It is thus the variability in time integrated and time averaged transports that will be the focus of this paper.

The Gulf of Maine (GoM) is a semi-enclosed sea, bounded to the northeast by Nova Scotia, to the southwest by the Massachusetts coast, and offshore by large shallow banks and shelves, including Georges Bank, Nantucket Shoals, and the Scotian Shelf (figure 1). At depths below 70m, the only connections to the outside ocean are through relatively narrow channels, the North East Channel (NEC) and a mid-shelf channel entering from the Scotian Shelf. In the GoM, there are three large deep basins, ranging from Georges Basin whose depth exceeds 350m and is connected directly to the deep ocean over a 240m sill, and the shallower Jordan’s and Wilkinson basin’s which are partially isolated from the offshore ocean by sills shallower than 200m [Uchupi, 1968]. Because the GoM is a semi-enclosed sea, its dynamics are markedly different from a more open coastline. The influence of alongshore winds are reduced over the basins, where isobaths are closed upon themselves, and the alongshore flows are distorted by the contorted isobaths. At the same time, an estuarine like circulation is developed in which dense water drawn in at depth in the NEC and less dense water entering near the surface from the Scotian Shelf is transformed in the GoM and leaves as intermediate density water via surface and mid-depth pathways onto the mid-Atlantic bight [Brown and Beardsley, 1978].

In the following sections, the variability in the circulation of the GoM will be quantified, along with estimates of the relative importance of the sources of that variability. Attention will be focused on the impact of that variability on transport over the ecologically relevant timescales of several months and longer. Because of this focus on longer time scales, super-inertial phenomena such as tides and internal waves will not be explicitly considered (though tides are included in the numerical model described below).

The circulation in the GoM and its variability is ultimately forced by the flow through
alongshore and offshore boundaries, the temperature and salinity transported into the do-
main by these flows, the winds and by the surface fluxes of heat and water. However, to fully
quantify the variability driven by the boundary forcing, it would be necessary to have ac-
curate information on the variability of the water masses entering the GoM. Unfortunately,
the variability of the temperature and salinity transported through the open boundaries on
the Scotian Shelf (SS) and through the North East Channel (NEC) are poorly constrained.

Thus the examination of the sources of circulation variability will be constrained to
include variability forced by winds, the changes in the volume of water entering the GoM
from the SS, and the variability in the large scale density field in the Gulf of Maine. This
later source of variability will implicitly include the effects of changing temperature and
salinity at the inflows to the GoM, as well as surface buoyancy fluxes, river inputs, and ver-
tical mixing variability. A brief analysis of the origins of the variability in density driven
circulation will then be given, emphasizing the importance of water mass transformation
within the GoM. The volumetric effects of inflow into the NEC (those forced by the trans-
port of a volume of water alone through the NEC, as opposed to those forced by changes in
the temperature and salinity of the inflowing water) will not be explicitly examined because
the NEC inflow is seen below to be forced by the winds and SS inflow, and thus is not an
independent constraint on the system. Variability driven by radius of deformation scale
features in the hydrographic field, such as eddies formed by baroclinic instabilities (e.g.
Vermersch et al. [1979]), will not be included in this analysis, thus making the estimates of
variability presented here a lower limit on the actual variability.

This partitioning of sources of variability into wind, boundary inflow transport, and
hydrographic forced variability introduces some ambiguity. Winds and inflow can alter
the hydrography in the GoM, and care must be taken to avoid double counting sources
of variability. To avoid this, the analysis of the effects of Scotian Shelf inflow will be
limited to the effects of a change in the alongshelf volume flux of water through boundary.
Changes in the T/S property of the inflowing water will be accounted for in the analysis
of hydrographically driven variability. Likewise, the analysis of wind driven variability
will be limited to the subinertial currents driven directly by the winds and by the barotropic
pressure gradients the wind sets up on the weatherband timescale of a few days. The effects
of wind driven mixing and the advection of density by wind-driven flows will be subsumed
into the discussion of hydrographic variability. This partitioning of sources of variability
can be made because the timescales over which the GoM first responds to a change in
winds or inflow occur on a timescale of a few days or less, while the effects of the change
in currents on density occurs on advection timescales. At least for density changes caused by the alongshore advection of density, these advection timescales are much longer, with O(10 cm s\(^{-1}\)) flows and distances of many hundreds of kilometers leading to timescales of weeks to several months. The analysis below will become suspect in locations where density advection modifies the baroclinic flow field on weatherband timescales, e.g. near the coast where a few days of upwelling favorable winds can create an upwelling front.

It will be found that most of the variability relevant to the transport on timescales of months or longer is not captured in most current models, not because of any failures of the models, but because of the inadequacies of our observational schemes which initialize and provide the boundary conditions for our models. A minimal set of ongoing observations needed allow contemporary models to capture the true variability of the gulf will be defined.

1. Methods

1.1 Quantifying Sources of Variability

1.1.1 Comparing two sources of Variability

In order to compare the relative importance of various sources of circulation variability in the ocean, it is necessary to examine how circulation variability of a known magnitude and timescale affects the transport across a section when averaged or integrated over a fixed time interval. If a fluctuating forcing produces a flow with a decorrelation time scale of \(\tau\), and induces a transport whose standard deviation is \(\sigma_V\) across the section, then over a single random fluctuation of the forcing it can be expected to drive a net time-integrated transport which scales as \(\tau \sigma_V\). After \(N\) random fluctuations of this magnitude, the standard deviation of the net, time integrated transport is \(N^{\frac{1}{2}} \sigma_V\) (derived from Bevington and Robinson [1992]). The random fluctuations decorrelate on a timescale \(\tau\), so \(N = T/\tau\), where \(T\) is the time over which the transport is integrated. Thus the standard deviation of the transport integrated over a time \(T\) is

\[
\text{STD of time integrated transport} = (\text{Dispersion}) \times T^{\frac{1}{2}}
\]

where the dispersion is defined as:

\[
\text{Dispersion} = \tau^{\frac{1}{2}} \sigma_V.
\]

Thus for timescales longer than \(\tau\), the relative contribution of two uncorrelated random processes to the standard deviation of the integrated or mean transport through a section
will scale linearly with their dispersion, as defined in (2). On timescales less than each \( \tau \), the relative contribution of each processes would scale linearly with their \( \sigma_v \). Implicit in this discussion is an assumption that the ocean responds linearly to the two sources of variability – e.g. that changes in the ocean caused by one source of variability do not effect the magnitude of the oceans response to the other source of variability.

1.1.2 Relating transport variability to forcing variability

In order to calculate the transport dispersion defined in (2), it is necessary to relate the strength and decorrelation timescale of the ocean transport to the strength and decorrelation timescale of the various forcing mechanisms of interest.

As a simple model of ocean dynamics, consider a transport which responds linearly to some forcing \( F \), with a linear first order friction of strength \( \lambda \):

\[
\frac{\partial U}{\partial t} = F - \lambda U
\]

(3)

This is a good model of the depth-integrated alongshore current forced by an alongshore wind of infinite extent along a straight coast (e.g. Dever [1997]), and is a reasonable model of many forcing processes in the coastal ocean which at first accelerate an along shore or along-isobath flow until some other mechanism retards the flow. In order to understand the decorrelation timescale of \( U \), it is useful to examine the case in which \( F \) is a white noise process, and thus the expected value of the Fourier transform of \( F \), \( \hat{F} \), is a constant independent of the frequency \( \omega \) and of some magnitude \( \hat{F}_0 \). Fourier transforming (3) allows us to write the spectra of the transport \( U \):

\[
\hat{U} = \frac{\hat{F}_0}{i\omega + \lambda}
\]

(4)

From (4), the expected power-spectra of \( U \) can be calculated. From the convolution theorem [Bracewell, 1986], it can be shown that the Fourier transform of the power-spectra is the unnormalized lagged auto-correlation function, allowing the calculation of the auto-correlation function of \( U \) from (4) as a function of a lag \( t_{\text{lag}} \):

\[
\text{Auto-correlation of } U = \exp \left( -\lambda |t_{\text{lag}}| \right)
\]

(5)

from which it can be seen that the decorrelation timescale of this simple model-ocean when forced by white noise forcing scales as \( \lambda^{-1} \). This can be explained by noting that in this simple system \( U \) is a low-pass filtered version of \( F \). On timescales longer than \( \lambda^{-1} \), \( U \) is
proportional to $F$. On timescales less than $\lambda^{-1}$, $U$ is proportional to the time-integral of $F$. If $F$ has a decorrelation timescale greater than $\lambda^{-1}$, and thus has little energy in periods less than $\lambda^{-1}$, $U$ will be roughly proportional to $F$.

From this it can be seen that if the response timescale of the ocean to forcing, $\lambda^{-1}$ is greater than the timescale of the forcing, then the decorrelation timescale of the transport is set by $\lambda^{-1}$, while if $\lambda^{-1}$ is less than the decorrelation timescale of the forcing it is the decorrelation timescale of the forcing which sets the decorrelation timescale of the transport. Similar results can be easily obtained for other related systems – e.g. the decorrelation timescale of surface wind-driven transport in an infinite $f$-plane ocean is that of the wind if the timescale of the wind is greater than the relevant response time, the inertial period, while the decorrelation timescale of the flow is roughly the inertial period if the wind has uniform energy at all frequencies.

1.1.3 Calculating timescales $\tau$ from observations

The decorrelation timescale of a timeseries can be estimated from the integral decorrelation timescale as discussed by Davis [1976]. For a finite length of discrete time data of length greater than $l$ taken at an interval of $\Delta t$, the timescale can be estimated using $l$ lags of the data with

$$\tau = \sum_{n=-l}^{l} \left( l - \frac{|n|}{l} \right) r(n\Delta t) \Delta t$$

where $r(n\Delta t)$ is the lagged auto-correlation of the data lagged by a time $n\Delta t$. The maximum lag $l$ must be greater than the actual decorrelation timescale for this to be valid. Where data is missing, it is not included in the lagged correlation. Because $r(n\Delta t)$ becomes poorly defined as $n$ approaches the record length, $\tau$ is estimated for the value of $l$ which produces the largest $\tau$, and $l$ is limited to half the record length.

1.2 Numerical Modeling

To estimate the variability in GoM transport driven by the fluctuating winds or SS inflow, a numerical model will be used to estimate the transport driven by these fluctuating forcings. Model runs will be used to estimate the time it takes the GoM circulation to reach a nearly steady state circulation after a change in winds or SS inflow (the “response timescale”). The numerical model runs will also be used to estimate the transfer coefficients between
the wind or the SS inflow and the transport in the GoM on timescales longer than the response timescale, and will be used to examine the linearity of the oceanic response to forcing on timescales longer than the response timescale. The GoM is numerically modeled with the Finite Volume Coastal Ocean Model (FVCOM). FVCOM is a free-surface, hydrostatic, primitive-equation numerical model with an unstructured triangle based finite element mesh [Chen et al., 2003]. A description of the model configuration, boundary conditions, model runs, and forcing are given in the appendix.

1.3 Hydrographic Data

The hydrographic data used in the analysis below were taken from the Bedford Institute of Oceanography hydrographic database, and include all data from 1970 to 2003. Data were removed if they extended below the water depth given by the USGS 15' GoM bathymetric product [Roworth and Signell, 1998]. The entire cast of these errant data points were not removed, for most appeared to be from XBT casts that were not terminated upon the impact of the BT with the bottom. All calculations of mean density are from casts which measured both temperature and salinity, and density is calculated from each measurement and then averaged.

2. Wind Driven Variability

The monthly mean winds in the GoM are persistently upwelling favorable as defined by the Maine coast (55°T), but the standard deviation of the winds averaged over a single month is much larger than the climatological mean wind for that month, indicating that in any specific month there is a large probability that the month’s averaged winds will not even be of the same sign as the climatological mean for that month (figure 2, from the NCEP reanalysis of Kalnay et al. [1996]). Variability is largely isotropic and is much stronger in the winter and early spring (figure 2 and Manning and Strout [2001]). The decorrelation times of the wind varies less from season to season than do the winds themselves, with a decorrelation timescale of 2.1 days in the winter and a slightly larger 2.4 days in the summer for the alongshore winds.

The response of the GoM to these winds has been studied extensively, both observationally [Brown, 1998; Noble et al., 1985] and numerically [Greenberg et al., 1997; Naimie, 1996]. The response of the GoM is strongest to winds along the Maine coast, and weaker to winds across the Maine shelf. Upwelling favorable winds (defined here and below with
respect to the Central Maine coast) and offshore winds drive sea-level setdown over the entire GoM, and the pressure gradients so formed drive substantial currents. The currents and pressure response to the winds is rapid, with roughly 80% of the response observed to occur within 36 hours [Brown, 1998], and the response is observed to be very nearly linear to windstress, both with respect to magnitude and direction [Noble et al., 1985]. These results are consistent with those obtained with the numerical model (see appendix). The modeled response of the gulf to winds does not change greatly by season, suggesting that typical changes in the density field do not significantly alter the response of the GoM to winds. Since the timescale of the response of the GoM to the winds is less than the decorrelation timescale of the winds, the timescale of the wind shall be used in (2) to estimate the dispersion of the wind-driven flows.

Consistent with observations, FVCOM and other numerical models of the region find that an alongshore upwelling favorable wind drives an offshore Surface Ekman flux (figure 3), which causes a sea-level setdown that drives a largely barotropic flow along the shelf, through the NEC [Ramp et al., 1985], along the ridge between Wilkinson and Jeffreys basins and to the west of Wilkinson basin, along the Maine coast to Nova Scotia and out the Scotian Shelf. The wind-forced flow is roughly along lines of constant depth, slightly more so in the winter when the water is less stratified (in all seasons, the internal radius of deformation is much less than the dimension of the Gulf and the deep basins). Near the bottom, the wind driven upwelling flow tends to flood the deep basins. The wind-forced depth averaged transport across the section between Georges Basin and Wilkinson Basin is relatively small, as the geostrophic transport across this section is nearly equal and opposite to the Ekman transport across the section.

Similarly, an offshore wind (as defined by the direction of the Maine Coast) drives a surface Ekman transport to its right, to the southwest. There is a divergence of this surface transport when it leaves the coast of Nova Scotia and a convergence when it encounters the north/south trending Massachusetts and New Hampshire coasts. These divergences and convergences are fed by Ekman transports from the Scotian Shelf and by a geostrophic return flow which travels north-westward along the Maine coast.

In table 1, the dispersion of the wind-driven transport is shown for the depth averaged transport across the standard sections. These dispersion results are valid for times greater than the decorrelation timescale of the winds for each season. The alongshore wind driven dispersion is large through the NEC, across the central GoM, along the EMCC and WMCC, and out the Scotian Shelf. This dispersion is much less in the summer than in the winter,
as would be expected from the seasonal change in the variance of the winds (figure 2). The transport dispersion driven by the cross-shelf winds are again as expected from the depth-averaged wind-driven currents shown in figure 2, with the cross-shore winds forcing less transport dispersion than alongshore winds in most sections.

3. Inflow Driven Variability

The transport along the Scotian Shelf (SS) is dominated by alongshore currents driven by cross-shelf density gradients [Loder et al., 2003]. The effect on the circulation in the GoM of the volume of water entering from the SS and driven by cross-shelf density gradients is quantified in this section – the effects of changes in the temperature and salinity transported into the GoM from the SS will be briefly considered in the next section and in the discussion. The effects of SS transport fluctuations impact the entire model domain quickly, within a few days, consistent with the observed coastal trapped wave speeds on the SS of 6.5 m s⁻¹ or faster [Schwing, 1989], while the different water masses entering the SS shelf affect the GoM on a slower advective timescale set by the O(10) cm s⁻¹ mean currents [Smith et al., 2001]. The response of the GoM to changes in SS inflow do not vary significantly in runs made with different monthly mean hydrographies, suggesting that the response is not significantly altered by realistic changes in the gulf’s hydrography.

The variability of the inflow along the SS is quantified from 3 different data sets – the repeated occupations of a cross-shelf section at Halifax, Nova Scotia [Loder et al., 2003], an analysis of this and other hydrographic data gathered by the Bedford Institute of Oceanography (BIO), and current meter data gathered as part of the Canadian Arctic Storms Project (CASP) from December 1985 to April 1986 [Anderson and Smith, 1989].

The analysis of Loder et al. [2003] divides the alongshore transports into “inner” and “outer” portions, the former from the shore to the 244 m isobath in a deep basin offshore, the later from that isobath across an offshore bank to the shelf break at 341 m. Experimentation with FVCOM indicates that little of the transport in the “outer” section enters the GoM, and alterations made to the outer transport by changing the upstream boundary conditions have negligible effects on the GoM circulation. Thus attention will be focused below on the variability of the transport in the “inner” section of Loder et al. [2003], and the effects of changes in inflow through the models open boundary are calibrated by its effect on the Halifax “inner” transport.

The standard deviation of the inner-shelf transport as determined from 1955 to 1970
hydrographic sections with a level of no motion at the bottom is 0.24 Sv for the entire year (data courtesy of C. Hannah, from the analysis of Loder et al. [2003]). The variability is 48% greater in the winter, with a standard deviation of 0.28 Sv for December through April and 0.19 Sv from July to September.

Unfortunately, the Loder et al. [2003] data are too widely spaced in time to calculate a decorrelation timescale – so to do so, current meter data from the CASP program of the winter of 1985-1986 was used (data courtesy of P. Smith and is described in Anderson and Smith [1989]). Horizontal density gradients calculated between both the 70m and 100m conductivity/temperature sensors on the 100, 165 and 220m isobath accurately predict the vertical shear of the alongshore currents through the thermal wind relationship on sub-inertial timescales (noted by Anderson and Smith [1989] in the mean, but shown to be true at shorter sub-inertial timescales by the author). The decorrelation timescales is about 10 days for both the cross-shelf “inner” density gradient and the vertical shear in the sub-inertial alongshore currents away from the surface and bottom (assumed to represent a geostrophic interior). Of course, these calculations are only valid for the one winter over which the CASP data exists.

To confirm the validity of the decorrelation time estimate, all hydrographic data from the BIO hydrographic database between 43.8-45°N and 63.8-62.5°W from 1970 to 2003 was used to calculate a time series of cross-shelf density gradients. Any data taken within 2 days of each other were assumed to have been taken simultaneously. The density observations differenced from the mean cross-shore density gradient at depths including 50, 70 and 100m and all data in the two day window were used to compute a mean cross-shelf density gradient anomaly time series at each depth. To robustly estimate a decorrelation time from this unevenly spaced time series, the data was broken into sets of pairs of data separated by a fixed time – e.g. all pairs of data 2-7 days apart, 7-14 days apart, ad infinitum, and the number of pairs both above or below the median density were recorded. If this number was greater than would be expected by chance, as judged by the test for a fair coin toss at the 95% level, the time series was considered auto-correlated at the tested lag. With this test, there is significant correlation for lags of between 2 and 8 days, while there is none on timescales longer, except for a weak indication of a seasonal cycle. This result does not change if the summer or winter data is excluded. Thus the results of the analysis of the CASP data seems robust. Because the decorrelation time of the inflow transport is longer than the timescale of the GoM response to the change in inflow, the former sets the decorrelation timescale used to estimate the dispersion.
The circulation induced in the GoM by a given inflow along the SS shelf is essentially identical in the summer and winter. Figure 4 shows the effect of a decrease in the transport across the Halifax inner-shelf section of 0.24 Sv on the GoM circulation. The anomalous flows forced by a reduction in Scotian Shelf inflow are strikingly similar to the alongshore-wind-driven flows. Flow enters the GoM through the NEC, flows along the ridge between the basins, along the Eastern Maine coast, around Cape Sable and out the Scotian Shelf.

In table 1, the transport dispersion driven by SS variability is seen to be somewhat greater than the wind-driven dispersion in the central and eastern GoM. This is not as much because of the greater strength of the currents the inflows force, but due to the longer timescales of the variability.

4. Density Driven Variability

Near surface geostrophic flows in the GoM are strongly driven by density gradients at depths below 100m (e.g. Brown and Irish [1992]). The deep density structure can be clearly seen to strongly influence the surface circulation in figure 5. As will be shown below, these deep density gradients vary greatly from year to year, and thus are an important source of transport variability. However, changes in the internal density field are not an external source of variability affecting the ocean – they are instead the result of multiple processes that are either hard to quantify (e.g. changes in the density of water entering the GoM) or processes that interact non-linearly (such as the interaction of wind and cooling driven vertical mixing). In this section, the effects of changes in the density field will be quantified, and in latter sections the sources of density gradient variability will be discussed.

Unfortunately, the available density observations are insufficiently dense in space and time to be used to initialize the numerical model adequately, for the model needs a relatively smooth and widespread data coverage to realistically model the circulation. So instead of using the numerical model to estimate transport variability driven by changes in hydrography, the hydrography from the BIO hydrographic database will be used to directly estimate geostrophic transports between the three basins, Georges, Jordan and Wilkinson, and thus to determine the variability of these transports. The numerical model will only be used to check the dynamic consistency of assumptions about the level of no motion. Unfortunately, this method fails between the basins and the shore, where temporal and spatial variation of the density field are on shorter scales than in the basin interior, and so are not reliably resolved by the relatively sparse hydrographic data.
In order to aggregate multiple density observations from multiple space and time points, three areas in the deep basins are chosen where the mean horizontal density gradients are weak, to reduce the aliasing of spatial gradients into a temporal signal. These three areas are shown in figure 6 and are defined by the 200 m isobaths around Wilkinson and Jordan Basin, and by the 200 m isobath of Georges Basin less a region near Georges Bank where there are large spatial gradients associated with tidal mixing [Chen et al., 1995]. The data is then binned by quarter. This quarterly data is included in the analysis if the number of casts with both temperature and salinity data in a bin in a quarter year is equal to or greater than 4. The results presented here change little if the threshold number of points is decreased to 2 or increased to 8.

From these binned averaged densities, the geostrophic transport is calculated assuming a level of no motion at 170m. This depth is about twenty meters shallower than the depths between the basins. This estimate of a level of no motion is roughly consistent with that used in past observations [Brooks and Townsend, 1989; Brown and Irish, 1992], and it is also consistent with the numerical model results. In figure 7, the seasonal cycle of transport across the sections connecting the deep basins of the GoM from the numerical model (with no wind forcing and monthly climatological density) are compared to estimates of geostrophic transport calculated from the density profiles in the numerical model at the endpoints of the sections. The seasonal cycle of the actual transport and the geostrophic estimate of transport agrees well for the transport between Jordan and Georges Basin ($r = 0.82$), and between Jordan and Wilkinson basin ($r = 0.73$). There is poor agreement ($r = -0.15$) in the seasonal cycle of the transport between Wilkinson and Georges basin, though the overall magnitude agrees well. This disagreement is due to the seasonal variation in the SS inflow, which, as can be seen in figure 4, affects flow across this section. This inflow driven transport is barotropic, modifies the local level of no motion as a function of season, and so is not reflected in the estimate of geostrophic transport driven by local density gradients. There is good agreement in the seasonal cycle of transport across the section between Wilkinson and Georges basin if the SS inflow is prevented in the numerical model. Regardless of the SS inflow, however, an anomaly in the density gradient between Wilkinson and Georges basin would drive an anomalous transport across the section connecting the basins.

As can be seen in figure 8, the density differences that drive the transport are at a maximum at depth, and are much reduced or even reversed near the surface, and thus it is the deep density gradients that drive the majority of the transport.
The decorrelation timescale of the geostrophic transport anomaly was calculated from quarterly estimates of the geostrophic transport. The estimates were detrended, and the seasonal cycle removed, for otherwise the seasonal cycle strongly modified the autocorrelations structure. The decorrelation time for the transport between Wilkinson Basin and Georges Basin was between a year and a year and half (with roughly N=40 data points in each correlation). The decorrelation time for the transport across the section connecting Georges Basin and Jordan Basin is about a 3/4 of a year (with N about 20 for each correlation). The decorrelation time for the transport across the section between Jordan basin and Wilkinson basin is difficult to resolve with quarterly data; when the analysis is repeated on bimonthly data the timescale is about four months, but this is dependent on assuming that a single CTD cast in the basin in a monthly period is sufficient to characterize its density. The estimate of the decorrelation timescale in the geostrophic flow between Jordan Basin and Wilkinson basin cannot be considered robust.

The standard deviation of the transport is, as seen in table 2, higher in the winter (quarter 1) than in the summer. The dispersion of the transport, as shown in table 1, is correspondingly greater in the winter than the summer, but in either case, the dispersion of the transport driven by density changes is much greater than that driven by other sources of variability. Thus on timescales of several months and longer, the transport variability driven by changes in hydrography dominates the other sources of variability. The implications of this finding to the predictability and modeling of the GoM flows is discussed below.

5. Discussion

The interannual variability of the density differences between the deep basins of the GoM are shown above to cause the large majority of the variability in the time integrated transport in the central GoM on longer timescales. But these density differences are, of course, not a phenomena external to the GoM. They are forced by the surface fluxes acting upon and the inflows into the GoM. In the following sections, the sources of these density differences will be briefly examined, and from this understanding recommendations of the observations needed to understand and model the circulation in the central GoM will be made.

5.1 Origin of large-scale density differences in the GoM

The mean circulation in the GoM is counter-clockwise, with water entering from the NEC and SS flowing from Georges Basin to Jordan Basin and then to Wilkinson Basin and then
out of the GoM either through the Great South Channel, the NEC, or over the north flank of Georges Bank \cite{Brown and Beardsley, 1978; Hopkins and Garfield, 1979; Smith et al., 2001}; The density in the basins at depths below 50 meters decreases as the water flows around the basins, with the lowest deep densities in Wilkinson basin (figure 8, Smith et al. [2001]). This decrease in density is associated with deep freshening and cooling, as the warm salty water entering the GoM through the NEC at depth is mixed during winter mixing events with cooler and fresher surface waters from the SS and from estuarine outflows \cite{Brown and Beardsley, 1978; Smith et al., 2001}.

The climatological cycles of density, temperature and salinity suggest that both winter fluxes of heat and momentum are important contributors to vertical mixing. Xue et al. [2000] found that much of the circulation in the GoM is driven by buoyancy fluxes, especially around the perimeter of the GoM. However, the water mass alterations experienced by the water moving through the central GoM are not consistent with mixing driven by surface cooling alone. When cooling or evaporation drives vertical mixing, the entire mixed water column becomes denser. When vertical mixing is driven by surface forced mechanical turbulence, deeper waters become less dense, and surface waters denser, as the water column is homogenized.

In a monthly climatology (figure 9) of the density of Wilkinson Basin, the waters below 100m are freshest, coolest and least dense at the end of the winter. Furthermore, as the deep waters circulate counterclockwise through the GoM, they get steadily less dense. Thus mechanical mixing, presumably driven at least in part by winter storms, must be an important contributor to the deep mixing. This is not to say surface buoyancy loss is unimportant: the surface waters are coldest at the end of the winter, and this contributes significantly to their densification. (A vertically homogeneous water column is not seen in the climatology to the full depth of the apparent effect of winter-time mixing because the climatology is an average over years with deep mixing and years without. The later years will contribute an average stratification to the deep waters in the climatology.)

The other basins show similar but weaker seasonal cycles of density as a function of depth, but their climatological stratification remains much stronger. This suggests that deep winter-time mixing is much less common in the other basins than in Wilkinson basin, which agrees well with prior analysis \cite{Brown and Beardsley, 1978; Hopkins and Garfield, 1979}.

A qualitative estimate of the relative importance of surface cooling and wind-driven mixing to the variability in the density gradients in the central gulf can be found by comparing the density and density differences in and between Wilkinson Basin and Georges
Basin to the surface fluxes of heat and momentum. (Jordan Basin is excluded from this analysis because its relatively poor data coverage reduces the power of the statistical analysis there). Comparisons were made between the density and density differences averaged from May to July and the integrated surface heat fluxes and the integrated absolute value of windstress for the preceding January through March. The time lag between the fluxes and the density was introduced in order to capture the full effect of the mixing (e.g. January densities do not incorporate the effects of March winds) and to take advantage of the greater oceanic data coverage in the summer. The strength of the wind mixing was estimated from the integrated absolute value of the windstress from the NCEP reanalysis [Kalnay et al., 1996]. Using the integral of stress raised to the three-halves power (e.g. $u^{3/2}$) made little difference. The mixing effects of buoyancy fluxes were estimated from the integrated surface heat fluxes from the NCEP reanalysis as well. This product is known to over-estimate latent and sensible heat fluxes [Renfrew et al., 2002]. However, comparison of the fluxes with in-situ estimates over the period of February-April 1995 (Beardsley et al. [2003], data courtesy of R.C. Beardsley) find that the correlation between the total heat flux and NCEP fluxes, both low passed with a three-day filter, is 0.95. This indicates that the NCEP fluxes capture the variability in the surface buoyancy fluxes very well.

The density in Wilkinson Basin is significantly positively correlated with winter cooling from 30 to 170m depth with a peak correlation of about 0.7 between 60 and 150 m (the surface 30 meters has been warmed by the summer sun). The density in Georges Basin is more weakly correlated to the integrated cooling, and the significant correlation between cooling and density is limited to the top 90 meters of the water-column, again suggesting limited deep mixing in this basin. Thus the density difference between the two basins is negatively correlated to the net cooling (peak correlation of $r = 0.6$ from 150 to 50 m), with greater cooling reducing the density difference between the two basins. The geostrophic transport across the section connecting the two basins, as calculated above, is significantly correlated to the net cooling with more cooling resulting in less transport ($r = 0.6$. All correlations greater than $r = 0.42$ are significant at $P < 0.05$ with 21 degrees of freedom from 21 years in which both basins had at least 4 casts within the basin in the months of interest.)

A similar pattern is seen between the integrated magnitude of the windstress, but with weaker correlations. The density in Wilkinson basin is weakly but positively correlated to the strength of the integrated winter wind magnitudes from 30 to 150m depth, while the positive correlation in Georges basin is limited to the surface 80 meters. The density
differences between the basins is only weakly correlated to the integrated wind from 110 to 200 m, and then with only marginal significance \((r = 0.43)\). The geostrophic transport is not significantly correlated to the strength of wind mixing \((r = 0.25)\).

The correlations of cooling and integrated windstress magnitude with between basin density differences are consistent with a model of deep mixing in Wilkinson basin setting up deep density contrasts with Georges basin which then drive cross-gulf transports. Winter time vector-mean alongshore and cross-shore winds are not correlated to the geostrophic transports, suggesting that mean wind-driven transports are not important drivers of inter-basin density differences. Nor are the density or density differences between basins correlated to the peak wind magnitudes in each winter.

The relative unimportance of interannual changes in integrated windstress magnitude to the interannual changes in horizontal density gradients, despite the importance of mechanical mixing to the evolution of the mean vertical density structure described above, can be perhaps explained by the relative lack of variability in the time-integrated windstress magnitude. The ratio of the standard deviation to the mean integrated windstress magnitude is about 6%, while the same ratio for the integrated cooling is 28%. Thus while the wind-mixing is important, its relative constancy means it does not contribute as significantly to year to year changes in the vertical mixing of the water column or to the variation in the horizontal density gradients. Also, much of the mechanical mixing may be caused by the strong tides in the GoM, and especially in Jordan basin, and this will have no variability from year to year [Brooks and Townsend, 1989].

Neither the variation in the winds nor the surface cooling explain a great deal of the variance in the density difference between the basins, or in the geostrophic transport these density differences drive. At most, year to year variation in cooling explains about 36% of the variability in the geostrophic transport between Georges Basin and Wilkinson Basin. The rest of the variability is likely to arise from two sources. First, time variation in the density of the waters flowing into the GoM will translate into density differences between the basins until water with the new density is able to flow into all of the basins. Second, a change in the stratification of the waters entering the GoM through the NEC or the SS, or a change in the density differences between these two inflows, could modulate the stratification in the basins, and thus ability of the winds or cooling to mix the waters.
5.2 What must be observed to explain GoM circulation variability?

The majority of the variability in the central GoM circulation averaged over seasonal to yearly timescales is driven by the fluctuation of density gradients in the GoM. To successfully model this circulation and its fluctuations, the evolution of these gradients must be captured by the model or imposed by the data used to constrain the model solution. Two approaches exist, neither of which are mutually exclusive.

A model able to run for the timescale of water residence in the GoM could be forced by the observed surface heat and buoyancy fluxes and by observations of the density of the water entering through the NEC and the SS. If it could accurately model the mixing process in the GoM, the model would accurately capture the interannual variability in the transports.

In the absence of detailed knowledge of the inflows, or in the absence of confidence in the ability of the model and forcing to capture mixing processes accurately, some form of data assimilation must be used to constrain the large scale density field in the model.

Recent efforts at modeling the GoM have assimilated sea surface temperature (SST) by nudging the model surface temperature toward observations (Chen et al. [2005b], HJ Xue pers. com.). However, a brief consideration of the thermal structure of the GoM suggests that this may not be useful in constraining the extent of vertical mixing in the winter. In most oceans, temperature decreases with depth, so excess deepening of the ML will overcool the surface, while insufficient deepening of the mixed layer will leave the surface warmer than it should be. Nudging the surface temperature toward the observed value will cool the surface if the ML is too shallow, increasing mixing, and will warm the surface if the ML is too deep, inhibiting mixing. Thus nudging will tend to correct the model errors in an ocean with a more typical vertical temperature distribution. However, in the GoM, temperature increases with depth below the seasonal thermocline. Over deepening the mixed layer will make the surface anomalously warm, while a too shallow mixed layer will leave the surface colder than it should be. Nudging the SST to the observed SST will thus tend to cause the model to mix less if the mixed layer is too shallow, and mix more if the mixed layer is too deep. Thus the incorporation of SST into the model with nudging will not tend to correct any problem the model might have with vertical mixing in the winter. It is unlikely that any more advanced data assimilation scheme would tend to do better, for in the absence of surface salinity data it is not possible to unambiguously ascribe any error in vertical mixing to insufficient surface cooling, erroneous deep temperature/salinity structure, or the model incorrectly parameterizing mixing. No assimilation scheme can
remove ambiguity from a poorly observed system.

What must be observed is the density in the basins at the deeper depths where the density gradients that drive the geostrophic transport exist. The necessary frequency of sampling is set by the decorrelation timescales of the density in the basin. As shown in table 3, the decorrelation timescale of density after the removal of the seasonal cycle ranges from 0.27 years in the deep Georges Basin to half a year or more in the mid-depths of the basins (see Ramp et al. [1985] on rapid change in the T/S properties of water entering through the NEC). Thus quarterly monitoring of the density in the deep basins could begin to provide the data needed to constrain the modeled circulation of the central GoM through the assimilation of the deep density structure in the basins.

6. Conclusions

The variability in the time-integrated transport on timescales of a few months and larger in the GoM is driven largely by changes in the density structure (at least in the interior of the GoM), and to a lesser extent by the SS inflow, and finally by the winds. However, neither the information needed to accurately capture the SS inflow variability, nor the information needed to capture the variation in the internal density structure of the GoM have been made sufficiently often to meaningfully constrain the models. Thus our models will fail to inform us of the nature of the true variability of the GoM circulation, no matter how perfect the models themselves are.

If the year to year variation in the circulation of the GoM is to be understood and quantified, the ocean must be routinely monitored. The volume of water entering the SS must be recorded on timescales short compared to the variability of the inflow (about 10 days). More importantly, the density of the waters entering the GoM at depth and near the surface, or the density in the GoM itself, must be monitored at least once a quarter year to constrain the circulation in the central GoM.

The scarcity of hydrographic data prevented similar calculations from being made for the Maine coastal currents, but it also seems likely that much of the variability in their transport is a function of changes in the density field that are poorly constrained in present models (e.g. Brooks [1994], Hetland and Signell [2005], Pettigrew et al. [2005], Xue et al. [2000]). Thus there is, at present, a mismatch between what we know and observe and what we must know and observe if we are to understand the variations in GoM circulation. No further refinements in our numerical models will be as useful as an effort
to improve the observations that drive the models. Efforts have begun to measure the relevant data, for example the moorings monitoring the NEC and SS inflow maintained by the Gulf of Maine Ocean Observing System (http://www.gomoos.org/), and the routine salinity/temperature measurements made by the National Marine Fisheries Service [Taylor and Bascunan, 2000]. These must be maintained if we are to understand the year to year variation in the GoM circulation and its effects on the ecosystem of the Gulf.

Appendix

The GoM is numerically modeled with the Finite Volume Coastal Ocean Model (FVCOM). FVCOM is a free-surface, hydrostatic, primitive-equation numerical model with an unstructured triangle based finite element mesh [Chen et al., 2003]. The model was configured to the GoM as described by Chen et al. [2005a] and Chen et al. [2005b]. The model mesh, shown in figure 1, has a resolution of about 8 km in the central GoM, reducing to 1.4 to 4 km on the steep portions of the northeast flank of Georges Bank and in coastal regions where the bathymetry has the shortest length scales. The depth of the slope and ocean outside of the GoM is truncated to 300m to reduce the computational demands on the model, and cross-comparison with full depth models indicates that this does not effect the results presented below [Chen et al., 2005b], though it does reduce the magnitude of the flows along the slope by removing the deep density gradients that drive them. The bathymetry is not truncated within the GoM.

The model is initialized at each month with a climatological monthly mean density field, described by Chen et al. [2005a]. Because of the increase in hydrographic data density at later dates, this climatological density field is heavily biased toward the past decade, with 68% of the data from 1990 to the present.

The wind forcing of the GoM is coherent over the model domain. On the 2-14 day timescales over which the GoM responds most strongly and coherently to wind forcing, the vector correlation coefficient between the National Center for Environmental Prediction (NCEP) winds [Kalnay et al., 1996] in the central Gulf of Maine and on the model boundaries is greater than 0.8 everywhere, and the magnitude of the correlated part of the winds varies relatively little over the model domain (c.f. Noble et al. [1985], Brown [1998] on the GoM wind response, and Chen et al. [2005b] and Manning and Strout [2001] on the extent and effects of spatially varying winds in the GoM). Thus the model will be forced by a uniform wind whose magnitude is based on the NCEP reanalysis winds at 42.8°N and
67.5°W, roughly in the center of Georges Basin. The NCEP winds compare well to observations (Manning and Strout [2001] and comparison to R. Beardsley in-situ flux estimates on the South Flank of Georges Bank. (c.f. figure 2.)). Uniform winds have also been used in most prior modeling ([Greenberg et al., 1997; Hannah et al., 2001; Naimie et al., 1994], but for exceptions see [Brown, 1998; Chen et al., 2005b]).

The winds force circulation not only in the GoM and SS system, but, because of the long length scales of the winds, they also force circulation outside of the model domain which can affect the model solutions through the open boundary. These effects are most important on the Scotian Shelf, for it is the boundary from which coastal trapped waves can enter the domain. In order to capture these effects, the free surface at the SS open boundary was modified to reflect the coastal setup/setdown driven by alongshore winds as observed by Schwing [1989] as part of the Canadian Arctic Storms Project (CASP). The setup was calculated from their observations, interpolated to their kilometer 300, which is the location of the SS open boundary in FVCOM. The cross-Scotian Shelf windstress was found to have a negligible impact on the coastal sea level, so its effects on the elevation on the open boundary was neglected. Schwing [1989] calculated that a 1 Pa wind along the Scotian Shelf (i.e. from 68°T) would lead to a 50 cm rise in sea level at the coast at the location of the model open boundary. (From Schwing’s (1989) “L” model, which implicitly includes the effects of the winds on the SS and the winds to the northeast which are correlated with these winds). Unfortunately, the observations of wind-driven setup/setdown from CASP do not greatly constrain the cross-shelf structure of the sea level signal. This structure determines the transport driven across the boundary by the winds. As can be seen in figure 1, the bathymetry at the boundary slopes to a maximum depth about a third of the way across the shelf, and then rises onto Emerald bank before reaching the shelf break and slope. Any cross-shelf gradient in sea-level imposed on the bank drives a circulation which in the model joins the coastal transport, though in reality it is likely that little surface pressure gradient would be wind-forced on such a bank [Brink, 1983]. Any cross-shelf gradient imposed on the open boundary offshore of the bank joins a slope circulation which has little impact on the GoM. We choose to impose the free surface cross-shelf gradient associated with the along SS wind linearly from the coast to the deepest point of the offshore basin, with a uniform free surface offshore of this point. This is roughly consistent with the observations of sea-surface pressure change across the shelf at Halifax as observed by Schwing [1989]. Spreading the wind forced cross-shelf sea-surface gradient at the boundary farther offshore would tend to decrease the impact of alongshore winds on the GoM by moving.
more of the wind-driven circulation onto the slope. Whether this is appropriate must be determined by further modeling or observation.

For the model runs used to estimate the transfer coefficients between wind and transport in the GoM, model was spun-up with no winds for 10 tidal cycles (similar to Naimie [1996] and Hannah et al. [2001]), and then the winds were applied at half strength for half an inertial period and then full strength for 3.6 more days. This stepped forcing nearly eliminates inertial oscillations forced by the onset of the winds. The duration of forcing is within the times over which Noble et al. [1985] and Brown [1998] found the maximum coherence between winds and currents. The transfer function between winds and currents is only slightly changed if these results are calculated with a wind stress 50% greater or smaller, and is also only weakly affected if the winds are run at full strength for 2 or 7 days.

Besides the winds, transports through the SS boundary can be forced by cross-shelf density gradient along the open boundary [Anderson and Smith, 1989; Schwing, 1989]. The alongshelf transport across the SS boundary forced by cross-shelf density gradients depends on the choice of a level of no motion. The nearly universal practice on the SS has been to assume that the cross-shelf pressure gradient caused by the hydrography, and thus the alongshelf geostrophic flow, is zero at the bottom. Thus the level of no motion is the water depth [Han and Loder, 2003; Hannah et al., 1998; Hannah et al., 2001; Hetland and Signell, 2005; Naimie, 1996]. The alongshelf transport across the open boundary is calculated by the model and is consistent with thermal wind. This, in turn, is consistent with observations of the vertical shear of alongshelf currents on the SS [Anderson and Smith, 1989].

The variability in the Scotian Shelf inflow to the model driven by changes in the crossshelf density gradient will be simulated by varying the sea-surface pressure along the boundary. On timescales less than that needed to advect density from the boundary to the interior of the model domain, this essentially barotropic forcing has been confirmed in model runs to produce the same effect on the interior of the model as changing the crossshelf density gradient on the boundary of the model. The longer timescale changes in the density field on the Scotian Shelf and in the GoM driven by changes in the inflow are considered elsewhere, as part of the examination of variability driven by changes in the density field. Changing the boundary inflow changes the flow field of the model within a day and a half. As with the winds, models run with modified boundary conditions are run with the unaltered boundary conditions for 10 tidal cycles and then the altered boundary conditions is imposed over two steps a half-inertial period apart, and then held steady for 3.6 days.
The transfer function between the altered SS inflow and the circulation in the model are computed from the last tidal cycle of this period. The results are little changed if the period of the altered flow is increased to the decorrelation period defined below for the SS inflow variability.

The open boundary conditions and the data used to forced the $M_2$ tides, and the details of the open boundary conditions on the Mid-Atlantic Bight and the offshore ocean, can be found in Chen et al. [2005a].
References


Table 1: Dispersion as calculated from (2) for the sections defined in figure 1. Starred estimates are unreliable due to poor estimates of decorrelation timescale. Decorrelation timescale in days, dispersion in $10^6 \text{m}^3 \text{s}^{-3/2}$.

<table>
<thead>
<tr>
<th></th>
<th>$\tau$</th>
<th>CpS-G</th>
<th>G-J</th>
<th>EMCC</th>
<th>J-W</th>
<th>WMCC</th>
<th>W-G</th>
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<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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</tr>
<tr>
<td>March</td>
<td>2.10</td>
<td>161.3</td>
<td>75.3</td>
<td>236.6</td>
<td>103.9</td>
<td>132.8</td>
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<tr>
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<td>2.50</td>
<td>53.1</td>
<td>41.6</td>
<td>94.7</td>
<td>46.5</td>
<td>48.2</td>
<td>5.0</td>
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<tr>
<td><strong>Cross-shore Winds</strong></td>
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<tr>
<td>March</td>
<td>1.60</td>
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<td>78.0</td>
<td>22.3</td>
<td>34.6</td>
<td>56.9</td>
<td>43.4</td>
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<tr>
<td>August</td>
<td>1.50</td>
<td>33.2</td>
<td>25.2</td>
<td>8.0</td>
<td>10.9</td>
<td>18.9</td>
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<tr>
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<td>10.00</td>
<td>144.6</td>
<td>44.8</td>
<td>99.8</td>
<td>78.7</td>
<td>21.1</td>
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<tr>
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<td>72.3</td>
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<tr>
<td>Winter</td>
<td>(see text)</td>
<td>738</td>
<td>700*</td>
<td></td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>Summer</td>
<td>(see text)</td>
<td>616</td>
<td>428*</td>
<td></td>
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Table 2: The mean and standard deviation of the geostrophic transport across the sections connecting the basins, and the number of years which contributed to each estimate. Units are in $\text{m}^3\text{s}^{-1}$. The means do not sum to zero for slightly different years contribute to each estimate, according to the availability of data in each basin and in each year. Each yearly estimate of the transport is independent, or nearly so, so the standard error in the mean should be approximately the standard deviation of the transport divided by the square root of the number of years which contribute to the data. Summer is defined here as July through October, and Winter as January through April.

<table>
<thead>
<tr>
<th></th>
<th>Georges-Jordan</th>
<th>Jordan-Wilkinson</th>
<th>Wilkinson-Georges</th>
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<tr>
<td><strong>Mean</strong></td>
<td>$10^5\text{m}^3\text{s}^{-1}$</td>
<td>$10^5\text{m}^3\text{s}^{-1}$</td>
<td>$10^5\text{m}^3\text{s}^{-1}$</td>
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<tr>
<td>summer</td>
<td>-0.94</td>
<td>-3.28</td>
<td>4.26</td>
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<tr>
<td>winter</td>
<td>-0.99</td>
<td>-1.56</td>
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<table>
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<th><strong>Standard Deviation</strong></th>
<th>$10^5\text{m}^3\text{s}^{-1}$</th>
<th>$10^5\text{m}^3\text{s}^{-1}$</th>
<th>$10^5\text{m}^3\text{s}^{-1}$</th>
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<tr>
<td>summer</td>
<td>1.27</td>
<td>1.32</td>
<td>1.46</td>
</tr>
<tr>
<td>winter</td>
<td>1.52</td>
<td>2.16</td>
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<th>Jordan Basin</th>
<th>Wilkinson Basin</th>
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<tr>
<td>summer</td>
<td>22</td>
<td>20</td>
<td>22</td>
</tr>
<tr>
<td>winter</td>
<td>13</td>
<td>11</td>
<td>17</td>
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Table 3: Decorrelation timescale in years for density in the deep basins in the GoM as a function of depth. The seasonal cycle was removed before making the calculation.

<table>
<thead>
<tr>
<th>Depth (meters)</th>
<th>Georges Basin (years)</th>
<th>Jordan Basin (years)</th>
<th>Wilkinson Basin (years)</th>
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<tbody>
<tr>
<td>10</td>
<td>0.62</td>
<td>0.57</td>
<td>0.31</td>
</tr>
<tr>
<td>50</td>
<td>0.50</td>
<td>0.72</td>
<td>0.57</td>
</tr>
<tr>
<td>100</td>
<td>0.43</td>
<td>0.54</td>
<td>0.89</td>
</tr>
<tr>
<td>150</td>
<td>0.30</td>
<td>0.54</td>
<td>0.89</td>
</tr>
<tr>
<td>200</td>
<td>0.27</td>
<td>0.56</td>
<td>0.75</td>
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Figure 1: Top) The domain of the FVCOM model, including the the 300, 200, 100 and 50m isobaths. Indicated are the sections across which transport is calculated, and the major basins and other features of the Gulf of Maine and Scotian Shelf.
Figure 2: Monthly windstress statistics from the 1970-2003 NCEP reanalysis and from Manning and Strout [2001]. A) Monthly mean winds in the alongshore and cross-shore direction, as defined by the coast of Maine. B) The standard deviation of the NCEP winds. C) Decorrelation timescales of windstress by month.
Figure 3: The oceanic response to 10 m s\(^{-1}\) alongshore (left) and cross-shore (right) winds at (top) 7 m, (middle) 100 m, and (bottom) the depth-integrated response. Alongshore and cross-shore are defined with respect to the Maine coast; the response is found from the difference between model runs with no winds and model runs in which the winds had been blowing long enough that the circulation has become nearly steady. Arrows are not shown for currents of less than 0.75 cm s\(^{-1}\) and transports per unit length of less than 2 m\(^2\) s\(^{-1}\). Wind converted to stress following Large and Pond [1981].
Figure 4: Anomalous depth-integrated velocities forced by Scotian Shelf inflow for the month of March, for an anomalous reduction of Scotian Shelf inflow of 0.24 Sv, the standard deviation of the Inner Shelf Halifax transport from Loder et al. [2003] for the winter months.
Figure 5: Climatological depth of 29.67 $\sigma$ isobath overlain by 20m de-tided currents for the month of March, from FVCOM with no mean winds and no anomalous SS inflow. Both the currents and the isopycnal depth calculated with the climatological density described by Chen et al. [2005a].
Figure 6: Mean density at 150m for (left) July-October and (right) January-April for the years 1970-2003, using all data from the BIO database. The thick black line around the south and west portions of Georges Basin mark the boundary of the area included in the density averages. Elsewhere along the perimeter of Georges Basin, the 200m isobath is the limit of the averaging area, as it is in Wilkinson and Jordan basin. The 300, 200, 100 and 50m isobaths are contoured.
Figure 7: Transport from the numerical model for the sections between the deep basins, and the equivalent geostrophic transport calculated from the density difference in the model between the basins, and assuming a level of no motion of 170 m. Top) Transport between Wilkinson and Jordan basin. Middle) Transport between Wilkinson and Georges Basin. Bottom) Transport between Georges and Jordan Basin.
Figure 8: Mean hydrography of the deep basins, averaged with all data from January to April for years 1970-2003. Left) Density in the deep basins. Middle) Density difference between two basins. Right) Horizontal integral of geostrophic velocity between two basins, with level of no motion marked by asterisk at 170 m. Integrated transports and density differences do not sum exactly to zero because means are calculated over slightly different years, due to different missing years in the data for each basin.
Figure 9: The monthly climatology of density, temperature and salinity in Wilkinson Basin from the BIO hydrographic database, including all data from 1970-2003. Density and salinity inversions in the upper 100m in February are a result of changes in vertical sampling resolution between the beginning and end of the data period.