A Brief Overview of the Physical Oceanography of the Gulf of Maine

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1.Background and Setting

The Gulf of Maine is a marginal sea that is nearly cut off from the Atlantic Ocean by Georges and Browns Banks. The gulf is well known for its anomalously cold waters, a fact that is primarily attributed to the gulf's location in the lee of the North American continent and to the shoal offshore banks that isolate and insulate the gulf from the warmer waters of the Atlantic (Fig.1). The principal connection between the gulf and the Atlantic is the Northeast Channel, a glacially-scoured valley with a sill depth of about 230 m that cuts across the continental slope and divides the banks. The shallower Great South Channel, with a sill depth of about 70 m, is a comparatively gentle depression in the shelf topography that allows a more limited exchange between waters of the gulf and Nantucket Shoals. Relatively fresh and cool water entering the gulf from the Scotian Shelf and fresh water from rivers contrasts with warm and salty Atlantic slope water that flows in through the Northeast Channel as an intermittent bottom current. The resulting water property distributions and estuarine-like circulation are complicated by rugged topography, tidal mixing, seasonal atmospheric interactions, and the influence of the Earth's rotation.

Inside the gulf, the complex bottom topography defines three major basins - Georges, Jordan and Wilkinson - separated at the 200 m depth by various topographic rises and banks. Jordan and Wilkinson Basins have maximum depths of about 275 m, but Georges Basin, which forms the inner terminus of the Northeast Channel, contains the greatest water depth in the gulf at 379 m. Deep water access to the inner basins is controlled by several sills, notably Truxton Swell on the north side of Georges Basin and Lindenkohl Sill (indicated by the "L" in Fig. 1) on the west side of Georges Basin. These sills evidently play an important role in the seasonal evolution of the circulation by controlling the spreading of denser water from the continental shelf outside the gulf.

For the purposes of this brief overview, it is convenient to divide the discussion into sections dealing with the tidal and non-tidal components of the circulation. However, it should be kept in mind that the tides are an integral part of the total circulation of the gulf-bay-bank system, and that the separation of the tides from the longer period water motions is a somewhat artificial demarcation. For example, the vigorous tidal currents in the eastern gulf are certainly coupled with the nontidal flows through nonlinear interactions, particularly frictional interactions with the complex bottom topography. Ultimately, it is necessary to consider the full spectrum of circulation scales in order to develop an adequate predictive capability for the circulation in the gulf.

The final section of this paper discusses several hypotheses for the seasonal evolution of circulation in the gulf and describes a monitoring program that will hopefully lead to an improved understanding of the circulation and its causes.

2. The Tides

The Gulf of Maine and its adjoining Bay of Fundy are well known for a nearly resonant semi-daily (M2) tidal response to oceanic forcing (Garrett, 1972; Brown, 1984). A co-tidal chart for the M2 constituent is shown in Fig. 2 (cotidal lines redrawn from Brown, 1984). The solid lines show the amplitude (half-range) and the dashed lines show the phase lag of high water relative to the time of the moon's passage of the Greenwich meridian. The near-resonant condition is indicated by the large increase in tidal amplitude from the outer edge of Georges Bank to the head of the Bay of Fundy, and by the relatively constant phase of the sea level oscillations in the inner gulf-bay system, both indications of a standing quarter-wave response. The tidal currents in the upper Bay of Fundy, over the shallowest parts and the inner edge of Georges Bank, and near the western end of the Scotian Shelf frequently exceed 1 m s^{-1}, whereas in the southwestern gulf the tidal currents are more typically a few tens of cm s^{-1}. Tidal height excursions increase eastward and northward in the gulf and at springs can exceed 15 m in the upper reaches of the Bay of Fundy.

Numerical models of the barotropic (vertically integrated) semidiurnal tide in the gulf area reproduce the actual tidal regime in considerable detail (Greenberg, 1983). For example, a recent calculation on a 60 x 40 grid with about 12 km resolution (Fig.3) illustrates the wavelike nature of the M2 tide as it enters and spreads through the gulf (Fig. 4, Brooks, unpublished). The figure shows contours of sea level height at six times during the three hours between a relatively "flat" sea surface condition and high water at the head of the Bay of Fundy. The tidal wave enters the gulf moving southwestward around the tip of Nova Scotia, then spreads into the Bay of Fundy, hugging the Scotian side in the manner of a Kelvin wave. Upon reaching the head of the bay, the wave reflects and moves southwestward along the New Brunswick side, filling the bay. The sea level increase continues to advance toward the western gulf, with high water reaching Cape Cod at about the time of the high at the head of the Bay of Fundy. The ebb cycle generally follows the inverse pattern.

The tidal currents exhibit a rotational character at most places in the gulf (Brown, 1984), as shown for example by the currents measured at several mooring locations in the gulf and near the inner edge of Georges Bank (Fig. 5, Cook, 1990). Near the bank and in the eastern part of the gulf, the tidal currents generally rotate clockwise, the vector tips forming an elliptical locus. In the western gulf, the tidal current magnitudes are smaller (note the scale change in Fig. 5) and the rotation sense is generally anticlockwise. In areas confined by topography, such as the location of mooring '86A', which was in a valley defining Lindenkohl Sill, the ellipses may become essentially rectilinear oscillations, and the tidal currents lose much of their rotary character. Table 1 lists the principal ellipse parameters for the depth-weighted tidal currents at the sites shown in Fig. 5. The data sets shown in Fig. 5 are from Moody et al. (1984), Marsden (1986), and Gottlieb and Brooks (1986).

Vigorous tidal stirring of the waters in the gulf keeps the water vertically well mixed over the cap of Georges Bank, the western shelf of Nova Scotia, and in most of the Bay of Fundy. In these regions (and in smaller areas along the eastern Maine coast), the turbulent kinetic energy supplied by tidal friction with the bottom is sufficient to overcome the stratification that would result from surface heating or fresh water buoyancy sources. The result is tidally mixed regions separated from stratified regions by surface and subsurface fronts, with the mixed regions revealed by relatively low surface temperatures. By comparing the kinetic energy provided by tidal mixing with the potential energy of buoyancy sources, the locations of the tidal fronts can be predicted with some accuracy (Simpson and Hunter, 1974). A recent calculation of this nature in which the effects of wind mixing were included shows the locations of tidal fronts in the regions just mentioned, and also near some of the eastern Maine coastal bays, near Grand Manan Island, over the cap of Browns Bank, and Nantucket Shoals (Fig. 6, Loder and Greenberg, 1986). One of the important consequences of the tidal stirring is to bring deep dissolved nutrients upward into the surface layers, where the enhanced light can result in higher biological productivity, so that the areas near and inside the tidal fronts shown in Fig. 6 also tend to support high primary and secondary production. For example, the tidally-stirred waters of the eastern gulf and Georges and Browns Banks support one of the world's richest fisheries (Yentsch and Garfield, 1981).

The tidal currents also contribute significantly to a residual (non-tidal) clockwise circulation around Georges and Browns Banks (Loder, 1980), which combines with a seasonal thermohaline-driven clockwise flow around the banks that reaches maximum strength of order 0.5 m/s in late summer (Loder and Wright, 1985). The low-frequency, tidally-rectified flow around the bank is modulated in strength by the fortnightly tidal variation, and over the steep northern slope of the bank the tidal and non-tidal currents are nonlinearly coupled (Magnell, et al., 1980).

3. The Non-tidal Circulation

The non-tidal circulation in the Gulf of Maine is basically an anti-clockwise gyre that is severely distorted by bottom topography and seasonally modulated by interactions with the atmosphere and the waters of the Atlantic continental margin. The first comprehensive description of the circulation was given by Bigelow (1927), based on extensive hydrographic and drift-bottle observations. Bigelow's classic circulation scheme, shown in Fig. 7, consists of two meshed gyres, with anticlockwise currents inside the gulf and a partially closed clockwise circulation around Georges Bank. The southern limb of the gulf gyre coincides with the northern limb of the clockwise flow around the edge of Georges Bank. Later drift-bottle studies (e.g. Bumpus and Lauzier, 1965) suggest that the surface currents intensify in the spring and summer months as stratification increases; the bottle data also indicated that in the winter months the gyre-like circulation becomes less evident and the surface waters inside the gulf drift slowly seaward toward and over the banks. Recent extensive current measurements near Georges Bank (Fig. 8, Butman et al., 1982) tend to support Bigelow's view of at least a partially closed flow around the edge of the bank; in fact, the mean current vectors, which represent conditions at different times of the year over a four-year period, suggest a stronger inflow into Great South Channel (and a greater horizontal shear in that channel) than might be inferred from Bigelow's schematic. The mean current vectors also show the presence of a strong eastward flow along the inner edge of the bank, as Bigelow inferred and as hydrographic data soon to be discussed confirm.

Away from the banks, inside the gulf, typical non-tidal upper level current speeds are 30-50 cm s^{-1} (Vermersch and Beardsley, 1979; Gottlieb and Brooks, 1986), but deep currents of several times that strength occur over sills and in narrow channels between basins (Brooks, 1990). Mean vectors derived from available current measurements do not conclusively indicate a single anticlockwise gyre, as the Bigelow schematic suggests, but rather the measured currents indicate a more complex spring and summer pattern in which separate cyclonic (anticlockwise) flow patterns develop in Georges and Jordan Basin, as will be shown shortly. In contrast, Wilkinson Basin, which appears to be partly sheltered by the shoal topography of Jeffreys Bank off Penobscot Bay, has a comparatively weak or nonexistent closed circulation, but instead seems to provide a pathway for the export of waters from the inner and eastern gulf toward the Great South Channel and the inner edge of Georges Bank.

In the winter the interior flow weakens in the gulf, and intense surface cooling produces sinking and convective overturning at least to mid-depths in the basins (Brown and Beardsley, 1978). Except for the eastern gulf, where tidal mixing prevails, spring and summer restratification eventually seals off a mid-depth layer known as Maine Intermediate Water (MIW), whose distinguishing characteristic is a pronounced temperature minimum (Hopkins and Garfield, 1979). Thus in the summer months the interior waters of the central and western gulf can be characterized as a three layer "sandwich," with a thin layer of relatively warm and fresh surface water overlying a somewhat saltier but much colder layer of MIW, beneath which is found (in all seasons) the warmer but very salty and therefore dense Maine Bottom Water, which derives much of its nature from the waters of the Atlantic continental slope. In the winter, surface cooling and convection erases much of the contrast between the surface waters and MIW. During the summer and fall months, the MIW is prominently exported via the strong flow along the inner flank of Georges Bank, thence through the Northeast Channel (Hopkins and Garfield, 1979). The MIW also carries a primary source of nutrients that can be mixed upward in the frontal region north of the bank to feed the rich biological population of the bank top (Hopkins and Garfield, 1981).

Maps of surface temperature and salinity show the tendency for late spring (June, 1983) water movements in partially separated gyres in Georges and Jordan Basins (Fig. 9, Brooks, 1985). The temperature (upper panel) clearly shows cooler water moving westward off the Scotian Shelf along the northern side of Georges Basin, while eastward tongues of both salinity and temperature trace the narrow, eastward flow along the inner edge of the bank and define the southern limb of a gyre-like flow in Georges Basin. Cool and fresh water in the eastern Maine coastal current moves southwestward along the inner part of Jordan Basin, and then turns offshore and eastward over the central and southern basin, again suggesting an anticlockwise surface circulation at least partially confined to Jordan Basin. Ten-week mean upper level currents measured at two locations (heavy arrows in Fig. 9) confirm the offshore and eastward water movement in western Jordan Basin and also show a surface water movement toward the South Channel in northern Wilkinson Basin. These apparent circulation patterns, while generally suggesting an overall anticlockwise circulation in the gulf, also indicate considerably greater complexity of the flow than indicated by the Bigelow schematic, at least in the spring and summer months.

As noted earlier, the ingress of salty bottom water from the Atlantic slope (referred to simply as "slope water," or SLW) is important to the seasonal cycle of circulation inside the gulf. For example, in June, 1984 the slope water (here denoted by a lower salinity bound of 34 ppt, following Bigelow, 1927) extended inward from the mouth of the Northeast Channel, where the outer edge of a warm-core ring from the Gulf Stream brushed northeastward across the entrance to the channel (Brooks, 1987). The slope water pooled in the deep part of Georges Basin, where the

34 ppt surface rose to less than 80 m from the surface. The salinity surface contours also indicate that the slope water spread northward across Truxton Swell into the eastern part of Jordan Basin (the evidence for spreading into Jordan Basin is stronger in 1983, corresponding to the maps shown in Fig. 9; see Brooks, 1985). The up-domed pools of slope water, which by virtue of its high salinity is more dense than the surrounding waters, result in dynamic "lows" in the sea-surface topography relative to level isobaric surfaces at depth, so that the sea surface dynamic topography relative to 100 m, for example, shows a typical relief of about 5 cm, with closed contours located over the deep basins (Fig. 10, right panel). The associated geostrophic surface flow relative to 100 m, indicated by the arrowheads on the figure, is consistent with separate anticlockwise gyres in Georges and Jordan Basins with maximum current speeds of order 30 cm/s, a clockwise eddy deflection over the shoal Jeffreys Bank, and a much weaker anticlockwise circulation in the southern part of Wilkinson Basin. Much of the detailed structure of the dynamic topography mimics the distribution of the deep slope water, pointing to the importance of exchange through the Northeast Channel. Adjacent to the Maine coast, the freshening influence of river waters adds to the tendency for southwestward geostrophic flow along the coast. Part of the coastal current eddies over Jeffreys Bank, carrying important nutrient and primary productivity stimuli into the warmer waters of the central and western gulf (Brooks and Townsend, 1989), and part of it turns back toward the east in the offshore limb of the Jordan Basin gyre, as indicated in Fig. 10.

To understand why the flow patterns exhibit the dependence on basin hydrography, it is helpful to examine the vertical structure of the water masses in the gulf. This procedure, in which the temperature, salinity and other data from individual stations are presented as contoured fields on vertical planes or sections, is the oceanographers' equivalent of the physicians' computer-aided tomographic scan, or "CAT" scan. For brevity, only a few examples of vertical slices from June, 1984 will be shown here (the "C" and "D" lines identified in Fig. 10); corroborating sections and plan-view maps from different stations and different years are given for example by Brooks (1985, 1987).

The "C" section begins near Mt. Desert Island at the Maine coast and slices across Jordan and Georges Basins, ending on the northern edge of Georges Bank (Fig. 10). The temperature and salinity sections along this line (Fig. 11) clearly show the "sandwich" structure referred to earlier, with the temperature minimum of MIW prominent across most of the section, except near the northern edge of Georges Basin (stations 12-14) where the warmer slope and shelf waters move westward and weaken the temperature minimum. The very warm (>9 deg. C) and salty (>34.5 ppt) slope water also moves westward and fills the lower part of Georges Basin. The relatively thin surface water layer also extends across most of the section, but with a front-like interruption near station 12 in the boundary region between the Jordan and Georges Basin gyres. Near the Maine coast, the lowered temperatures and salinities indicate the advection of tidally mixed water from the eastern gulf and the influence of coastal rivers.

The tendency to form partially separated gyres in the basins is indicated by the updomed isohalines, which are reasonable proxies for isopycnals for the range of temperatures involved here. A geostrophic interpretation of the salinity section in Fig. 11 is consistent with an upper-level anticlockwise circulation in both basins, as suggested in plan view in the dynamic height field in Fig. 10.

The "D" section crossed the Northeast Channel near its sill (Fig. 10), and the temperature and salinity on this section clearly show the three major water types in the channel (Fig. 12, upper panels). The view is looking out the channel from inside the gulf, with Georges Bank on the right and Browns Bank on the left. Most obvious is the warm and saline slope water, which hugs the northern side of the lower part of the channel as it moves into Georges Basin. Above the slope water much cooler and fresher water moves inward from the Scotian Shelf. Another branch of Scotian Shelf water (SSW) enters the gulf north of Browns Bank, following the edge of the shelf (not seen in Fig. 12, cf. Smith, 1983). As noted by Bigelow (1927) and many investigators since, the low salinity and considerable volume of inflowing Scotian Shelf water significantly influences the seasonal development of the density structure and hence the circulation inside the gulf. Finally, Fig. 12 shows the mid-depth (~75 m) temperature minimum of MIW pressed against the northern edge of Georges Bank as the MIW exits the gulf through the Northeast Channel. The hydrographic fields in the channel for the case shown can be simply characterized as three stacked and interleaved horizontal wedges of water, with the bases of the SLW and SSW wedges abutting Browns Bank and the base of the MIW wedge abutting Georges Bank. The indicated inflows and outflows in the channel are generally typical of spring and summer conditions. However, it is important to note that significant disruptions of this pattern can occur when the shelf-slope front meanders offshore, allowing colder and fresher Labrador waters to influence the inflow (Sutcliffe et al., 1976), or when Gulf Stream rings press against the outer edge of the continental shelf, resulting in a much warmer and saltier shelf water inflow that may temporarily impede or even reverse the usual MIW outflow (Brooks, 1987).

Hydrographic survey data and current measurements from three successive spring-summer seasons were used to construct a schematic description of the inferred circulation, shown in the lower panel of Fig. 12 (Brooks, 1985). The intent was to see if the recent data sets could be combined qualitatively in a way that made consistent sense. The hollow arrows represent the upper layer (MIW and above) water movements, whereas the solid arrows indicate the spreading paths of slope water. It is emphasized that the circulation patterns shown may not be typical of other years or other seasons; however, there are obvious similarities to Bigelow's schematic (Fig. 7) as well as a few differences, and these bear further examination.

The overall impression given by the upper level water movements sketched in Fig. 12 is consistent with a general anticlockwise (cyclonic) gulf gyre, but with significant modifications in the major basins, as noted earlier. The presence of a well defined cyclonic gyre in Georges Basin accounts for the principal differences from Bigelow's schematic. In this regard, it is significant that Bigelow showed a single bottle track curving cyclonically in the eastern part of Georges Basin, so he presumably felt that part of the surface inflow adjacent to Browns Bank was quickly returned to the outflow branch next to Georges Bank. If in Bigelow's schematic one were to substantially enlarge that single cyclonic track in Georges Basin and correspondingly diminish the broad riverlike flow from the central gulf directly into the Bay of Fundy, the resulting pattern would be much more like that suggested in Fig. 12, where the movement toward the Bay of Fundy seems to come from Wilkinson and Jordan Basins but with greater containment of the circulation in the latter. It seems likely that tidally-driven horizontal eddy mixing in the eastern part of Jordan Basin accounted for some of Bigelow's numerous drift bottles that escaped Jordan Basin and moved into the Bay of Fundy. Eddy motions and non-linearities are of course excluded in a geostrophic interpretation of hydrographic data (e.g. Fig. 10), yet these effects become important near the Scotian Shelf, where the tidal action is strong and the interpretation of dynamic height contours as streamlines therefore becomes invalid. It is also possible that the prevailing southwest summer winds over the gulf enhanced movement toward the Bay of Fundy of the surface-drifting bottles, which were drogued to a depth of "a fathom or two at most." Bigelow evidently discounted the wind influence on his drifters, referring to it as "trivial" in comparison with other current sets. Even with the possibility of wind influence on the surface water movements, however, Bigelow found it necessary to invoke a shortened return trip via Jordan Basin for some of the bottles released near the eastern Maine coast, as shown in Fig. 7 by three or four tightly recurving bottle tracks turning offshore near Mt. Desert Island, which resemble the dynamic height contours in Fig. 10. Four of the bottle tracks indicate a closed anticlockwise circulation confined to Jordan Basin, and in this sense the two schematics are quite similar. There are a few other details that differ, such as the clockwise eddy motion apparent "behind" Jeffreys Bank, but overall it is remarkable that Henry Bigelow, with Nansen bottles, surface drifters and a schooner for tools, was able to deduce with such clarity the salient summer circulation features in the Gulf of Maine - Georges Bank region.

The vernal intensification of the circulation inside the gulf is associated with springtime river runoff and inflow from the Scotian Shelf, contrasted with dense Atlantic slope water that enters as a deep current in the Northeast Channel and spreads over sills into the inner basins. The currents at the sills can be surprisingly energetic and variable, as shown for example by current records from a mooring at Lindenkohl Sill, which controls the deep water pathway from Georges to Wilkinson Basin (Figs. 13 and 14; Brooks, 1990). The mooring was located close to the saddle point of the sill, in water 225 m deep. The current meters were positioned at depths of 76, 127 and 212 m, respectively near the MIW core depth, near the lower boundary of the MIW core, and 13 m off the bottom in a valley defining the sill. The upper and mid-depth instruments sampled currents that were relatively free from topographic steering, but at the near-bottom instrument the currents were obviously guided by the topography of the valley, which is about 5 km wide at the 212 m depth. The cross-sill axis (i.e. the valley axis) is oriented roughly northwest-southeast, and these directions are referred to here as "inward" and "outward," respectively. Figure 13 shows that, at the time the mooring was deployed (June, 1986), a slope water ridge extended across Georges Basin (depth contours of the 34 ppt surface shown to the right of the heavy dashed line) and a pool of MIW with minimum temperature <4.8 deg. C occupied southern Wilkinson Basin (contours of minimum MIW temperature shown to the left of the heavy dashed line). The mooring was located in a region that was alternately influenced by both water types as the circulation evolved during the summer.

The current, temperature and conductivity records were smoothed with a low-pass digital filter with a half-power period of 40 hr, which effectively removes the tidal and other short-period motions from the records. For comparison with the currents, a wind stress vector time series was calculated from hourly observations of wind speed and direction recorded at Manana Island, which is located about 10 km off the central Maine coast (Fig. 1).For periods longer than about 2 days the coherent wind stress scales in the Gulf of Maine region are larger than the cross-shelf scale of the gulf (Noble and Butman, 1979), so the wind speed and direction measurements from Manana Island are taken to be representative of those over the interior of the gulf. The filtered current and wind stress records are shown in Fig. 14 for the month of August and the first week of September to illustrate the variability in the longer record, which is discussed in greater detail by Brooks (1990). The vectors are shown as a sequence of three-hourly "sticks," oriented with north at the top of the figure.

The current records show energetic fluctuations in the period range of days to months, with speeds ~ 50 cm s^{-1} sustained for periods of several days. The most striking features are the bottom-intensified, pulse-like bursts of inward water movement over the sill from Georges Basin into Lindenkohl Basin, with inflow speeds > 100 cm s^{-1} recorded 13 m above the bottom in late August. The inflow pulses are steered by the channel topography, so that the strongest flow vectors are generally transverse to the sill (approximately toward 310 °). Compensatory outflow events are not noted, so the flow over the sill indicates a net inflow and not an oscillatory exchange between the basins. Such strong nontidal currents in the gulf are surprising (cf. Vermersch and Beardsley, 1979) and they indicate active deep water exchange between Georges, Lindenkohl and Wilkinson basins.

The deep inflow pulses were correlated with alongshore (northeastward) wind stress pulses during late June to early July and again in late August. A detailed examination of the wind-to-current coherence is given by Brooks (1990). The correlated inflow events occur with periods in the range of days to weeks, similar to current fluctuations noted in the Northeast Channel (Ramp et al., 1985). Over Lindenkohl Sill, the deep inflow pulses typically began when the wind stress vector turned in the alongshore direction toward the eastern end of the Maine coastline. When the wind stress vector rotated offshore, the inflow abruptly ceased. This relationship suggests a coastal upwelling mechanism in which alongshore (northeastward) wind stress produces an offshore surface Ekman transport that lowers the sea level in the inner part of the gulf relative to offshore. The ocean response to the resulting inward pressure gradient is complicated by many factors, but in the presence of weak density stratification the barotropic pressure gradient arising from sea level tilt will be only partially relieved at depth, so that the deep inward currents over Lindenkohl Sill can respond quickly to northeastward (or the relaxation of southwestward) wind stress before the stratification increases in the summer months. On the other hand, in the presence of sufficiently strong stratification, mass adjustments can completely compensate the pressure gradient at depth, so that the deep currents can gradually become decoupled from the wind as seasonal stratification intensifies.

The wind response mechanism just outlined of course requires examination in greater detail with better observations and with three-dimensional numerical models. Barotropic models that include wind forcing illustrate the complexities present even without the density stratification (Wright, et al., 1986). However, the density-modified upwelling scenario just described is supported by the fact that the visual correlation between the wind stress and the deep current over Lindenkohl Sill weakened after the first week of July and essentially vanished by the second week of August (Fig. 14), when relatively dense Atlantic slope water moved westward from Georges Basin to the mooring location (Brooks, 1990). During this time, the upper and mid-level currents turned southward and accelerated, consistent with geostrophic flow around the western end of the advancing slope water ridge in Georges Basin. After about August 21, the slope water influence decreased at the mooring site and wind-current coupling resumed, but with a nearly out-of-phase relationship, which suggests that the simple upwelling mechanism was modified by a more complicated response involving the residual density stratification remaining when the slope water moved away from the mooring site.

While there is much more to be said about the Lindenkohl and other current records, the most important point to be made here is that the non-tidal currents in the gulf can be surprisingly strong and variable, reflecting active exchange between the basins. Almost certainly these exchanges, which involve the movements of the major water masses and to some extent the influence of the surface winds, play a central role in the seasonal evolution of the circulation patterns in the gulf and adjacent to Georges Bank.

4. Strategies for the Future

As noted previously, the contrast between buoyant coastal waters and the denser slope water provides a potential energy source for the seasonal baroclinic circulation in the gulf. Thus it is of interest to examine the relationships between river, shelf and slope water inflow into the gulf (Fig. 15). The upper curve is a smoothed representation of monthly 10-year averages of runoff from the Merrimack River, which is characteristic of the total river flux into the gulf (Meade, 1971). The central curve is an annual regression on a 6.5 year record of alonshore current over the Scotian Shelf, with negative values indicating inflow into the gulf (Smith, 1989). The lower curve is a semiannual and annual regression on transport below 75 m in the Northeast Channel, calculated from two-year current records (Ramp et al, 1985). The annual maximum inflow of Scotian Shelf water occurs in February, before the April maximum runoff from rivers inside the gulf, and before seasonal restratification can inhibit mixing with MIW, which reaches its lowest salinity in March (Hopkins and Garfield, 1979). The deep inflow through the Northeast Channel occurs episodically, but with a smoothed annual and semi-annual cycle that has a minimum near zero in the spring, a rapid increase in June, a maximum in August, and a secondary maximum in December.

There is an obvious inverse correlation between the fresh water input to the gulf and the deep slope water inflow in the channel, which reaches its minimum essentially at the same time as the maximum river runoff. This relationship is consistent with the idea that an accumulation of fresh water inside the gulf in the unstratified season adds an adverse barotropic pressure "head" that inhibits deep slope water inflow, a suggestion advanced by Hopkins and Garfield (1979). The effect of the barotropic pressure gradient would be to shift the zero line to the right in Fig. 16, which shows the integrated dynamic height difference relative to the surface for various station pairs in June, 1986. For example, the Northeast Channel - Georges Basin (NEC-GB) curve suggests outflow in the channel in the upper layers, with maximum outflow at about the MIW depth, and inflow below the MIW depth. The addition of an adverse offshore barotropic pressure gradient would shift the zero axis to the right, which would enhance the outflow tendency for MIW and retard the inflow of slope water. Similar arguments can be made for the other station combinations in Fig. 16. This mechanism may account for the rapid decrease of slope water inflow in the Northeast Channel that seems to begin in late winter, at about the time of maximum Scotian Shelf water inflow of about 2.8 x 10^5 m^3s^{-1} (Smith, 1989). The river flux may be less important due to its relatively small volume (< 1% of total gulf volume).

The mechanism just described leads to the hypothesis that in "wet" years of extreme precipitation and vernal run-off, inflow from local rivers and especially from the Scotian Shelf (with freshening influence from the Gulf of St. Lawrence) the offshore barotropic pressure gradient force should delay or retard the deep water inflow in the Northeast Channel. Conversely, in relatively dry years the offshore pressure "head" should be weaker and the deep water injection from the Atlantic should occur earlier. Recent data suggest, with some ambiguity, that this relationship holds; e.g. slope water reached Jordan Basin earlier in the "dry" year of 1982 than it did in the relatively "wet" year of 1984, when the Laurentian runoff was at extremely high levels (Brooks, 1985; Smith, 1989). Thus it appears that the climatology of the entire eastern New England states and the Maritime Provinces may partially control the seasonal development of the circulation inside the gulf and along the northern edge of Georges Bank. Because of the nearly one year lag between run0ff in the Laurentian Basin and the arrival of the freshened shelf water pulse at the western end of Nova Scotia, there is also the possibility of a climatalogical feedback mechanism, in which two successive wet years could produce an extreme delay of slope water entry and circulation spin-up in the second year, where as in alternating wet-dry years, the effects of local river runoff in the gulf would tend to cancel the buoyancy signal from the Scotian Shelf water.

It furthermore seems likely that the surface winds, by virtue of the upwelling mechanism described earlier, can modulate the longer term thermohaline factors controlling the indraft and spreading of slope water, so that the timing of the annual vernal spin-up of the circulation will also depend on the complexities of atmospheric forcing. In spite of the complications, however, we now have available a working hypothesis that can be tested with long term monitoring of the currents at critical locations, coupled with the application of fully three-dimensional models capable of reproducing the baroclinic circulation as well as the tides.

A monitoring program for the Gulf of Maine. In broad terms, a practical goal for a useful monitoring program is to provide a long term data set adequate to calibrate, initialize and verify a suitable three-dimensional numerical model that can be used to predict the non-tidal aspects of the circulation in the gulf. The corresponding principal scientific objective would be to understand the factors which control and determine the seasonal evolution of the circulation in the gulf, in its connecting channels, and adjacent to Georges Bank, ultimately leading to an ability to predict the salient features of the baroclinic circulation in the region.

Such a program will require, as a minimum, a multi-year, international commitment to an extensive field and modeling program that should contain at least the following principal elements:

1. Telemetered current, temperature and salinity measurements from moorings located at critical sills, over the Scotian Shelf and in the deep basins,

2. A fast-response capability to carry out shipboard hydrographic and drifter surveys at times of scientific drama,

3. Regular release of drogued, satellite-tracked or LORAN drifters,

4. A coordinated program of satellite-based remote sensing of sea surface temperature and perhaps color,

5. A ship-of-opportunity program to obtain repeated data transects from interesting sections (e.g. Portland-Halifax),

6. A community effort to develop an adequate three-dimensional numerical model of the gulf-bank region, with the model code available to all cooperating investigators, and perhaps most important,

7. International cooperation and good will between agencies, scientists and fishermen.

5. References.

Bigelow, H. B., 1927: Physical oceanography of the Gulf of Maine. *Fish. Bull. 40*, 511-1027.

Brooks, D.A., 1985: Vernal Circulation in the Gulf of Maine. *J Geophys. Res., 90 (C5)*, 4687-4705.

__________,1987: The influence of warm-core rings on slope water entering the Gulf of Maine. *J. Geopys. Res., 92 (C8)*, 8183-8196.

__________,1990: Currents at Lindenkohl Sill in the southern Gulf of Maine. *J. Geophys. Res.*, 95, 22173-22192.

__________ and D. W. Townsend, 1989: Variability of the coastal current and nutrient pathways in the Gulf of Maine. *J. Mar. Res., 47* , 303-321.

Brown, W. S., 1984: A comparison of Georges Bank, Gulf of Maine and New England Shelf tidal dynamics. *J. Phys. Oceanogr.*, 145-167.

__________and R. C. Beardsley, 1978: Winter circulation in the western Gulf of Maine: Part 1. Cooling and water mass formation. *J. Phys. Oceanogr.,8*, 265-277.

Bumpus, D.F. and L.M. Lauzier, 1965: Surface circulation on the continental shelf of eastern North America between Newfoundland and Florida. *Ser. Atlas of the Mar. Environ., Folio 7,* Am. Geogr. Soc.

Butman, B, R.C. Beardsley, B. Magnell, D. Frye, J.A. VErmersch, R. Schlitz, R. Limburner, W.R. Wright, and M.A. Noble. Recent observations of the mean circulation on Georges Bank. J. Phys. Oceanogr., 12, 569-591.

Cook, M.S., 1990: An analysis of tidal currents in the Gulf of Maine. MS thesis, Texas A&M Univsersity, College Station, August, 1990, 75 pp.

Garrett, C. ,1972: Tidal resonance in the Bay of Fundy and Gulf of Maine. *Nature, 238*, 441-443.

Gottlieb, E and D.A. Brooks,1986: Current meter and atmospheric data from the Gulf of Maine: 1982-1985. Texas A&M Univ., Technical Report No. 86-3-T, 80 pp.

Greenberg, D. A., 1983: Modeling the mean barotropic circulation in the Bay of Fundy and the Gulf of Maine. *J. Phys. Oceanogr., 13*, 886-904.

Hopkins, T.S., and N. Garfield, 1979: Gulf of Maine Intermediate Water. *J. Mar. Res., 37,* 103-139.

Hopkins, T.S., and N. Garfield, 1981: Physical origins of Georges Bank water. *J. Mar. Res., 39*, 465-500.

Loder, J. W., 1980: Topographic rectification of tidal currents on the sides of Georges Bank. *J. Phys. Oceanogr., 10*, 1399-1416.

Loder, J. W. and D.A. Greenberg, 1986: Predicted positions of tidal fronts in the Gulf of Maine region. *Cont. Shelf. Res., 6(3)*, 397-414.

Loder, J.W. and D.G. Wright, 1985: Tidal rectification and frontal circulation on the sides of Georges Bank. *J. Mar. Res.,43*, 581-604.

Magnell, B.A., S. L. Spiegel, R.I Scarlett and J. B. Andrews, 1980: The relationship of tidal and low-frequency currents on the north slope of Georges Bank. *J. Phys. Oceanogr., 10*, 1200-1212.

Marsden, R.F., 1986: The internal tide on Georges Bank. *J. Mar. Res., 44*, 35-50.

Meade, R.H., 1971: The Coastal Environment of New England. New England River Basins Commission, Boston, 47 pp.

Moody, J.A., R.C. Beardsley, W.S. Brown, P. Daifuku, D.A. Mayer, H.O. Mofjeld, B.Pietrie, S. Ramp, P. Smith and W.R. Wright, 1984: Atlas of tidal elevation and current observations on the northeast American continental shelf and slope. U.S. Geolog. Surv. Bull. 1611, U.S. GPO, 122 pp.

Noble, M. and B. Butman, 1979: Low-frequency wind-induced sea level oscillations along the east coast of North America. *J. Geophys. Res., C(6),* 3227-3236.

Ramp, S.R., R.J. Schlitz, and W.R. Wright, 1985: The deep flow through the Northeast Channel, Gulf of Maine. *J. Phys. Oceanogr., 15*, 1790-1808.

Simpson, J.H. and J.R. Hunter, 1974: Fronts in the Irish Sea. *Nature, 250*, London, 404-406.

Smith, P.C., 1989: Seasonal and interannual variability off southwest Nova Scotia. *Can J. Fish. Aquat. Sci. 46* , 4-20.

Sutcliffe, W.H.Jr., R.H. Loucks, and K.F. Drinkwater, 1976: Coastal circulation and physical oceanography of the Scotian Shelf and Gulf of Maine. *J. Fish. Res. Board. Can., 33*, 98-115.

Vermersch J. A. and R. C. Beardsley, 1979: Winter circulation in the western Gulf of Maine: Part 2. Current and pressure observations. *J. Phys. Oceanogr., 9*, 768-784.

Wright, D.G., D.A. Greenberg, J.W. Loder and P.C. Smith, 1986: The steady-state barotropic response of the Gulf of Maine and adjacent regions to surface wind stress. *J. Phys. Oceanogr., 16*, 947-966.

Yentsch, C.S. and N. Garfield, III, 1981: Principal areas of vertical mixing in the waters of the Gulf of Maine, with reference to the total productivity in the area. In: *Oceanography from Space* (J.F.R. Gower, Ed.), Plenum, New York, pp 303-312.

Figure Captions

Fig. 1 Area map of the Gulf of Maine - Georges Bank - Bay of Fundy region, showing names of major topographic features and several ship tracks. The filled circle shows the location of a current meter mooring at Lindenkohl Sill. Weather data were obtained from a station located on Manana Island, near the central Maine coast. Depth contours are given in meters.

Fig. 2 Cotidal chart for the M2 tide, showing sea level amplitude (cm) and Greenwich phase lag (degrees) constructed from coastal and offshore pressure measurements. Redrawn from Brown (1984).

Fig. 3 Bathymetry (m) and a 60 x 40 element rectangular grid used for a numerical model of the tides in the Gulf of Maine region. The horizontal resolution of the grid is about 12 km.

Fig. 4 Sea level height contours (m) computed on the numerical grid of Fig. 3 for the 3.67 hours prior to high water at the head of the Bay of Fundy. The M2 tide enters the gulf around the western end of Nova Scotia, then spreads into the Bay of Fundy along the Scotian side. After reflecting from the head of the bay, the wave spreads westward and southward into the rest of the gulf. The near-resonance is indicated by the large amplitude at the head of the bay. The open-ocean tidal forcing amplitude was 0.5 m. Times shown in hours before high water at the head of the Bay of Fundy.

Fig. 5. Depth weigthed M-2 tidal ellipses for current meter data from the Gulf of Maine region (Cook, 1990). The rotation sense of the ellipses is defined by the arrowheads. The tidal ellipses of stations P1, P2, 83A and 84A have been offset from their actual locations, indicated by the dots. The scale for the western Gulf stations CP, MO, CL, 82A, 83B, 84A and 83A is 10 cm/s for the length of the scale bar shown in the figure; for the other stations the length of the scale bar corresponds to 40 cm/s.

Fig. 6. Predicted frontal positions for tidal and summertime wind mixing (___), using tidal dissipation rates calculated from Greenberg's (1983) model. The positions of the log (h/Dt) = 1.9 contour (...) and the 50 m isobath (---) are also shown. Figure from Loder and Greenberg, 1986

Fig. 7 Bigelow's (1927) classical circulation schematic for the Gulf of Maine region in summer months, based on multiple experiments with surface drift bottles, hydrography, and plankton distributions.

Fig. 8. Mean Eulerian current measurements from Georges Bank area (Butman et al., 1982). The boldface number at origin of vector keys measurements to data in table shown by Butman et al. (1982). Number in parentheses following identifier indicates water depth at that station. The number at the tip of the vector indicates the depth of measurement in m. The length of the vector is proportional to the mean current speed. See Butman et al. (1982) for additional details.

Fig. 9. For June 2-15, plan view map of surface temperature (a) and salinity (b) for the Gulf of Maine area. The contour interval is 0.5 °C for temperature < 9 °C and 1 °C otherwise; the salinity contour interval is 0.2 ppt for salinities >31 ppt and 1 ppt otherwise. The heavy arrows show velocity vectors (scale indicated) from 10-week mean currents measured at 25 m depth at moorings "A" and "B" (Figure from Brooks, 1985).

Fig. 10. As in Fig. 9, except showing the depth of the 34 ppt surface (left) and the sea surface dynamic height relative to 100 decibars. Hydrographic data from the ship tracks identified are shown next.

Fig. 11. Temperature and salinity along the "C" hydrographic transect shown in Fig. 10. The temperature minimum of Maine Intermediate Water extends across most of the section.

Fig. 12 (Upper panels): Temperature and Salinity along the "D" hydrographic transect shown in Fig. 10. This section crosses the Northeast Channel near its sill. View looking out of the gulf. (Lower panel): A schematic diagram of the vernal circulation in the upper and lower layers of the Gulf of Maine, inferred from June hydorgraphic surveys (Brooks, 1985). The hollow arrows show general patterns of water movement in the top 75 m during the time of strongest interior circulation. The solid arrows show the inward and westward spreading paths of deeper Atlantic slope water.

Fig. 13. For June, 1986 depth of the 34 ppt surface (m) on the right of the heavy dashed line, and minimum temperature (deg. C) of MIW on the left. The arrowheads show geostrophically inferred surface flow. A current meter mooring was located at Lindenkohl Sill near station C-7 (circled).

Fig. 14. Low pass filtered currents from the Lindenkohl Sill mooring, and wind stress computed from data obtained at Manana Island (Fig. 1). North is "up" in the figure.

Fig. 15. Regressions on data for river runoff (upper curve), Scotian Shelf water inflow (central curve) and inward transport below 75 m in the Northeast Channel (lower curve); see text for details and data references.

Fig. 16. Integrated dynamic height difference (dyn-cm) relative to the surface for pairs of stations in the gulf from June 1986. Station names: NEC (Northeast Channel), GB(Georges Basin), JB(Jordan Basin), WB(Wilkinson Basin). SILL refers to a station at the NEC sill.