Long-Term Moored Array Measurements of Currents and Hydrography over Georges Bank: 1994-1999

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Abstract:

In conjunction with the GLOBEC (Global Ocean Ecosystems Dynamics) program, measurements of moored currents, temperature and salinity were made during 1994-1999 at locations in 76 m of water along the Southern Flank of Georges Bank and at the Northeastern Peak. The measurements concentrate on the biologically crucial winter and spring periods, and coverage during the fall is usually poorer.

Current time series were completely dominated by the semidiurnal M$_2$ tidal component, while other tidal species (including the diurnal K$_1$ component) were also important. There was a substantial wind-driven component of the flow, which was linked, especially during the summer, to regional-scale response patterns. The current response at the Northeast Peak was especially strong in the 3-4 day period band, and this response is shown to be related to an amplifying topographic wave propagating eastward along the northern flank. Monthly mean flows on the southern flank are southwestward throughout the year, but strongest in the summertime. The observed tendency for summertime maximum along-bank flow to occur at depth is rationalized in terms of density gradients associated with a near-surface freshwater tongue wrapping around the Bank.

Temperature and salinity time series demonstrate the presence, altogether about 25% of the time, of a number of intruding water masses. These intrusions could last anywhere from a couple days up to about a month. The sources of these intrusions can be broadly classified as the Scotian Shelf (especially during the winter), the Western Gulf of Maine (especially during the summer), and the deeper ocean south of Georges Bank (throughout the year). On longer time scales, the temperature variability is dominated by seasonal temperature changes. During the spring and summer, these changes are balanced by local heating or cooling, but wintertime cooling involves advective lateral transports as well. Salinity variations have weak, if any, seasonal variability, but are dominated by interannual changes that are related to regional- or basin-scale changes.

All considered, Georges Bank temperature and salinity characteristics are found to be highly dependent on the surrounding waters, but many questions remain, especially in terms of whether intrusive events leave a sustained impact on Bank waters.

*Regional Index Terms:* Northeast US shelf; Gulf of Maine; Georges Bank

*Key Words:* Wind-driven Circulation; Buoyancy-driven Circulation; Stratification; Seasonal/interannual Variability
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1. Introduction

The Global Ocean Ecosystem Dynamics (GLOBEC) Georges Bank project focused on understanding the processes that determine the success of cod and haddock populations in this highly productive fisheries region. Both fish have a similar life cycle (Wiebe et al., 2002) in that they spawn over the northeastern portion of the Bank during mid-winter (typically January-February), and then the eggs and developing larvae drift southwestward with the ambient currents along the southern flank of Georges Bank. By late spring, the surviving fish are mobile enough that they no longer drift with the currents. Since GLOBEC is particularly interested in physical-biological questions (involving targeted fish and zooplankton species), such as how a disruption of the along-bank flow might affect year-class success, attention was concentrated on the portion of the year when the fish were planktonic (not motile), hence presumably most susceptible to changes in the physical environment. In a sense, each year class represents only one degree of freedom, so observational replication required a multi (five)-year field program.

The larval drift paradigm (summarized by Lough (1984)) drove the physical thinking in the GLOBEC program. Aspects of the field program included multiple, roughly monthly, hydrographic and shipboard acoustic current surveys (e.g., Flagg and Dunn, 2003), extended drifter studies (e.g., Brink et al., 2003), focused process studies (such as of the onset of springtime stratification, e.g., Lentz et al., 2003a), and a moored array which is the subject of this study. The GLOBEC long-term moored array (Irish et al., 2005) was intended to provide continuous Eulerian time series measurements (during the seasons of biological interest) of currents, stratification and water mass properties. From a broader perspective, the time series were intended to document any anomalies that could correlate with biological shifts.

The Bank’s hydrographic structure provided a critical context for the GLOBEC field study (Flagg, 1987). Like much of the eastern North American continental shelf, there is a clearly defined, year-round shelfbreak front (typically rooted near the 100-120 m isobath), which usually slopes upward offshore. It separates Bank water, which is typically fresher by about 2 and cooler, from saltier slope waters offshore. Over the central Bank, tidal mixing homogenizes the water column. During the winter (when there is generally no surface warming), this vertical homogenization covers the entire Bank inshore of the shelfbreak front. During the summertime, surface warming provides enough stability that complete vertical mixing is only found in water shallower than about 60 m. Under these conditions, a sharply defined tidal mixing front near the 60-m isobath separates the shallower mixed water from the ambient stratified water.

Our present goal is to use the GLOBEC long-term moored array data to characterize physical oceanographic processes over Georges Bank. Tides (e.g., Moody et al., 1984; Werner et al., 2003) receive some attention because of their overwhelming dominance of kinetic energy. We focus mainly on variations on time scales longer than a day, since the advective importance of a given phenomenon increases with its inherent time scale (e.g. Pringle, 2006). For the velocity field, we are particularly concerned with wind-driven motions, off-bank influences, longer-period tidal effects and seasonal mean flow patterns. In terms of the moored hydrographic variability, our primary concern is the occurrence of various types of anomalous water that occur over the Bank (from Warm Core Rings, the Scotian Shelf and the Gulf of Maine), and their
duration. Providing a clear understanding of the current and hydrographic variability will then provide a sound basis for context for the broader, interdisciplinary issues.

2. Array Design and the Moored Data Set

The moored array design was developed to match the overall GLOBEC Georges Bank program objectives. The southern flank (SF) site (Figure 1 and Table 1) had the highest priority and was maintained during all 5 years of the field program because it could monitor the stratification, currents, and water properties in this region where larval cod pass southwestward early in each year. The site is in 76 m of water and on-bank of most excursions of the shelfbreak front and yet offshore of the summertime tidal mixing front. The seafloor on the southern flank between the 50 to 120 m isobaths is removed from any shelf edge canyons and has a relatively smooth sand/gravel surface with some broken shells. No features larger than scallop dredge tracks were observed by side-scan sonar or fathometer near the SF site.

The SF mooring was maintained from fall 1994 through summer 1999 (Irish, 1997 and 2000). Because of the programmatic emphasis on cod and haddock spawning and larval transport, each year’s deployments lasted from late fall until summer (Figure 2, upper panel). Temperature and conductivity (hence, salinity) measurements were made at 5-meter intervals down from the surface to 50-m depth, then about 1 m above the bottom. Bio-optical instrument packages at 10 m and 40 m also measured the light transmission, photosynthetically available radiation (PAR), optical backscatter and chlorophyll-a fluorescence. Meteorological observations (wind speed and direction, air temperature and relative humidity, short and long wave radiation and PAR) were made on the surface buoy at 3-m elevation. Current profiles (spanning roughly 8 m to 65 m depth in 1-m vertical bins) were obtained with a downward-looking 300 kHz Acoustic Doppler Current Profiler (ADCP) mounted just below the surface buoy. Backscattered acoustic amplitude was also measured as an indication (although not quantitative) of the time and depth variations of the sediment and biological scatterers in the water column. Finally, bottom pressure was measured from a separate bottom instrument frame that isolated the sensor from contaminating mooring motion. The SF instrument deployment and data return are summarized in Figure 2, and by Irish et al. (2005).

The Northeast Peak (NEP) mooring was deployed in program years 2, 3 and 5 (Figure 2, lower panel) in the region where cod and haddock spawn. The site (Figure 1) was located on the same isobath as, but upstream of, the SF mooring. It was also intended to be well positioned to monitor any Scotian Shelf Water crossover events (e.g., Smith et al., 2003). The NEP mooring was initially designed to be the same as the SF mooring, but practical constraints, including mooring and sensor damage, reduced the number of NEP sensors and deployments. For example, sensor allocation was reduced during winter well-mixed times to 10-m vertical spacing. An ADCP, deployed as at the SF site, was included on most deployments. Finally, a lightly instrumented mooring (CR) was deployed briefly at the top (crest) of the Bank in 43 m of water, but the resulting data are not used here. The NEP and CR data return is summarized in Figure 2 and by Irish et al. (2005).

All moorings were designed with elastic tethers near the bottom (Irish and Kery, 1996; Irish, 1997) in order to maintain a consistent, yet compliant, mooring tension. An electromechanical
cable supplied power to sensors and transmitted data up from selected sensors to the surface buoy and its data system. All sensors controlled by the buoy (the odd 5-m increments) were sampled at 1-min intervals and were then averaged to one-hour values. Temperature was resolved to better than ±0.001ºC, and equivalent salinity to about ±0.002. The autonomous temperature and conductivity sensors (SBE SeaCats and MicroCats at even 5-m increments) recorded single readings at 2 or 5-min intervals with the same resolution as the buoy temperature and conductivity sensors. The RDI ADCP profiled at about 2-sec intervals, allowing waves to be averaged out, and keeping random measurement error below ±0.5 cm/sec. Vector winds were sampled at 1 Hz, resolved into components relative to the buoy, and averaged. Every minute these averaged components were rotated into magnetic coordinates and accumulated for one-hour averages. The radiation sensors were sampled at 10-sec intervals and averaged to hourly values. The moorings were recovered and thoroughly serviced for about a month each fall, and turned around at sea each spring. Unplanned gaps in the record were caused by data system failures, sensor failures, and moorings breaking loose (mainly due to fishing activity). In several cases, echoes of the ADCP acoustic beams from the bio-optical packages or other instrumentation on the mooring line contaminated velocity measurements at discrete depths. In these cases, the bad data were eliminated and replaced (if needed) by linear interpolation in the vertical.

After each mooring recovery, data were downloaded and normalized using the most recent calibration constants. All temperature and conductivity sensors were calibrated once a year either by the manufacturer or at the Woods Hole Oceanographic Institution (WHOI). The temperature sensors were generally very stable, drifting only a few millidegrees per year. The yearly calibrations showed an average temperature sensor drift of under ±0.002º/yr with a maximum drift of less than ±0.005ºC/yr. The conductivity drift was considerably more. The yearly sensor calibrations showed the electronics drifted an equivalent of ±0.02. Bio-fouling effects on the conductivity sensors were the main source of sensor drift and were reduced by using poison plugs. Many of the salinity records had to be adjusted for drifts of 0.05 over a deployment, with the drifts being toward lower salinity due to bio-fouling. The maximum drift observed was about 0.15 on one sensor. Pre- and post-deployment calibrations did not accurately measure bio-fouling effects. Correction was done iteratively by comparing the vertical salinity structures at moorings to those from nearby CTD profiles, and adjusting the moored salinities to make the resulting mooring density profile consistent.

The raw ADCP data were averaged into hourly or half-hourly temporal bins. Following preliminary data processing, the SF current time series were usually rotated 20° counterclockwise into an along-bank (u), cross-bank (v) coordinate system (the main exception being that the tidal analysis was carried out in un-rotated coordinates). This rotation choice, while made based on topographic information, is also representative of the subinertial major axis direction. At the NEP site, the bottom was flatter, and the choice of coordinate rotation less obvious, so that the time series were left in an east-north orientation. All data were low-pass filtered with a half power point of 1.96 days. For some purposes, we segregate seasons by defining “summer” to be the six months beginning May 17, and “winter” as the rest of the year (consistent with Brink et al., 2003).
The moored array time series were supplemented by several other records. Currents, temperature, conductivity, and surface forcing were measured at the 1995 GLOBEC Stratification Study mooring site ST1, about 30 km southwest along the 76-m isobath (Alessi et al., 2001; Lentz et al, 2003). The ST1 data were used to extend the SF records into 1995. A continuous wind stress record was generated by J. Manning (see Manning and Strout, 2001), based on the nearby NOAA NDBC buoy 44011 (Figure 1), with gaps filled by regression using data from other NDBC buoy locations. In addition, coastal sea level and nearby atmospheric pressure data were obtained from NOAA data archives in order to calculate barometrically adjusted sea level (ASL). A continuous surface heat flux record was generated for the SF site using the fifth-generation Pennsylvania State University–NCAR Mesoscale Meteorological Model (MM5) configured and modified for application to the Gulf of Maine/Georges Bank region (GoM MM5) by Chen et al. (2005) and Hu (2009) at UMass-Dartmouth. The radiation fluxes were computed using satellite data from the International Satellite Cloud Climatology Project while the air-sea momentum and heat flux components were computed using the Fairall et al. (2003) COARE 2.6 bulk flux algorithms with inclusion of Advanced Very High Resolution Radiometer (AVHRR)-derived SST. S. Hu (UMass-Dartmouth) provided Gulf of Maine (GoM) MM5 surface time series at grid points near SF which were interpolated and daily averaged to make the SF record.

The overall GLOBEC moored time series data set is described in greater detail by Irish et al. (2005, 2006).

3. Current Measurements

3.1. Introduction

Several aspects of the current variability are immediately evident from the kinetic energy spectrum of depth-averaged SF velocity (Figure 3). First is the dominance of the semidiurnal tidal currents, primarily the M₂ component at 1.93 cpd. This very strong tidal component is also reflected in its M₄ (3.86 cpd) and M₆ (5.79 cpd) overtones, both of which contribute significant (95% confidence) peaks. Likewise, the diurnal tidal currents are consistently energetic in both seasons. The near-inertial energy (near the Coriolis frequency $f$) is not significantly above the background in either winter or summer. Finally, at frequencies lower than about 0.8 cpd, the winter kinetic energy levels are significantly higher than the summertime values, and have a pronounced, broad bulge centered at about 0.3 cpd. In the following, we treat first the tides, then the “weather band” (roughly 2-10 day periods), and finally monthly and longer time scales.

3.2. Tidal Pressure and Current Structures

Each SF bottom pressure data segment was analyzed separately for tides using the harmonic method (Dennis and Long, 1971; Irish and Brown, 1986) and the five main constituents were found to be the same for each deployment to within 95% confidence limits. Thus, only the weighted (by record length) average results are presented here (Table 2). Semidiurnal tides dominate, and the results are within error of those from Moody et al.’s (1984) station M4, which was located nearby on the 84-m isobath. The SF N₂ constituent is larger than the S₂ even though, regionally, S₂ tends to be somewhat stronger (Moody et al, 1984). This amplification is
consistent with the Gulf of Maine-Bay of Fundy tidal resonance at a frequency just below the semidiurnal (Garrett, 1972).

Currents from both the SF and NEP sites are analyzed here in all uncontaminated 1-m vertical bins. True eastward and northward components were analyzed using the “T_Tide” software package (Pawlowitz, et al, 2002). The M2 and K1 tidal constituents at 20-m depth show no significant seasonal or long-term differences among deployments, so all velocity results reported here are weighted averages over all data segments (Table 3). The 20-m M2 constituent accounts for 75.0% of the record variance at SF and 88.9% at NEP, while the 20-m K1 constituent accounts for only 1.3% of the record variance at SF and 0.6% at NEP. Overall, the seven principal tidal constituents and their overtones account for 81.9% and 95.6% of the total current variance at SF and NEP, respectively (Table 3), but, by far, the dominant constituents are semidiurnal.

The SF M2 velocity profile in elliptical representation (Figure 4) shows that the major axis velocity is about 1.5 times the minor, and that the velocity vector rotates slightly in the clockwise direction as depth increases. Amplitudes are almost depth-independent in the upper 40 m of the water column but below about 40 m, the velocity decreases more rapidly toward zero, implying a boundary layer of O(30 m) thick (in agreement with the findings of Moody, et al., 1984, and Werner et al., 2003). The Greenwich phase is nearly constant in the upper half of the water column at 5º, and then decreases with depth (by about 10°), so that bottom currents lead shallower currents in time.

The NEP M2 tidal profile (Figure 4), compared to SF, shows about twice the tidal amplitude and stronger depth variations. Velocity is nearly constant in the upper 20 m, and then decreases uniformly down to about 60-m depth. This suggests a much thicker, O(50 m), bottom boundary layer here than at SF, as might be expected with the much stronger tidal currents here. The ellipticity, phase variation, and net tidal veering at this site are comparable to those found at SF, while the overall results are again consistent with those of Moody et al. (1984). The N2 and S2 (the other significant semidiurnal constituents) tides show the same vertical structure (within 95% confidence limits) as the M2 at both the SF and NEP sites, so are not shown. The amplitudes for both are both smaller than the M2 and the Greenwich Phase differs (Table 3), but the inclination is essentially the same – across the bathymetry.

The diurnal K1 (Figure 5) currents are much weaker than the semidiurnal currents at both sites, and are slightly stronger at NEP than at SF. At SF, the amplitude is nearly constant in the upper water column – increasing slightly with depth down to about 25 m, decreasing slowly with depth to about 40 m, and then decreasing more quickly down to 65 m. The Greenwich phase shows a smooth increase with depth of about 20°. Consistent with stronger turbulence effects, the NEP K1 current decreases smoothly over the entire water column, while the Greenwich phase decreases with depth by about 5°. That the tidal phase trends with depth are in the opposite sense with for subinertial tides (e.g., K1) and superinertial tides (e.g., M2) can be qualitatively accounted for using laminar boundary layer theory (e.g., Munk et al., 1970). Overall, our diurnal
tidal parameter estimates agree reasonably with those presented by Moody et al. (1984) who only had access to much shorter in situ records.

The tidal ellipse orientation and tidal current ellipticity $e$ (ratio of minor to major axis amplitude) serve as useful diagnostics of the tidal dynamics. Specifically, using the 20-m ellipse values in Table 3, the SF M$_2$ major axis is aligned towards 31° west of north, about 11° counterclockwise from the cross-isobath direction, with an ellipticity $e = 0.699 \pm 0.019$, very close to the theoretical plane wave value of $\sigma/f = 0.679$ (where $\sigma$ is the tidal frequency and $f$ is the Coriolis parameter). At NEP, the M$_2$ ellipse is also aligned towards 31° west of north, with an ellipticity $e = 0.662 \pm 0.011$, also very close to $\sigma/f = 0.689$. These results support the conceptual model of the M$_2$ tide propagating across the southern flank of the Bank as a locally plane Sverdrup wave into the Gulf of Maine/Bay of Fundy system (Loder, 1980; Brown, 1984). The presence of a propagating, rather than a standing tidal wave over the Bank, is consistent with the existence of substantial tidal dissipation within the Gulf of Maine/Bay of Fundy system (Greenberg, 1979). The other semidiurnal tides have a similar, plane wave, character over the southern flank of the Bank. The diurnal tides are more complex. The K$_1$ ellipse is aligned with the local topography at both SF and NEP, consistent with Daifuku and Beardsley (1983), who showed that the K$_1$ tide can be represented to lowest order as a combination of a barotropic Kelvin and a shelf wave propagating along the continental shelf/slope from Nova Scotia into the Mid-Atlantic Bight.

### 3.3 Overview of Lower Frequency Currents

The low-pass filtered (subtidal) current variability is summarized using the first mode Empirical Orthogonal Function (EOF) at each site, since these tend to represent the variance fairly efficiently. The modes were computed separately for winter and summer by using all available data from each season. On-bank ($v$) and along-bank ($u$) velocity time series (at SF; at NEP, north and east velocities are used) are treated as independent inputs at each depth for this analysis. The resulting modal structures (right panels in Figures 6 and 7) then align roughly along the major axis (e.g., Kundu and Allen, 1976) orientation, so that the modes are a compact way to represent nearly equivalent amplitude (the normalized modal structures are multiplied by the square root of the associated eigenvalue) and direction information. If only data from a single depth were used, this calculation would return the principal axis orientations and standard deviations exactly. The overall subtidal current variability at both locations, as described by the individual principal axis currents at each depth (not shown), is more nearly isotropic than the lowest-mode EOFs suggest, a result equivalent to accounting for higher, more noisy, EOFs. This suggestion that more complex higher modes are more nearly isotropic is perhaps not surprising.

At both SF and NEP, the variability is highly correlated vertically and its amplitude decreases monotonically with depth. Both locations exhibit a counterclockwise veering with increasing depth, as is often found in northern hemisphere wind-driven shelf systems with a surface Ekman layer and compensating flow at depth. Werner (1999) found, that during wintertime at SF, surface mixed layer thickness (which is indicative of boundary layer thickness) varied over about 20-60 m (the entire water column was almost never homogenized), and that during the summer, the mixed layer was rarely deeper than 10 m. The velocity EOF structures do not reflect the surface mixed layer depth scales, but instead the dominant vertical variation is associated with
weakening flow toward the bottom. Winter variability is substantially stronger than summer at both sites, but the difference is more pronounced at SF where the winter amplitude is almost twice the summer’s. In contrast to semidiurnal tidal amplitudes, subtidal current variability at NEP (Figure 7) is substantially weaker than that at SF (Figure 6), especially during the wintertime. The orientation associated with the lowest mode NEP EOF (Figure 7, right panels) is east-west in summer (compared to southward mean flows) and northeast-southwest in winter (roughly the same orientation as the mean flow). In contrast, the primary subtidal velocity fluctuations at SF are aligned year-around with both the mean flow and the local isobaths. We attribute the directional inconsistency at NEP to the relatively flat bottom at that location, which contrasts to the weak, O(0.001), but fairly uniform slope at SF. Integral time scales (the integral of the autocorrelation function from 0 to very large lag), which characterize typical fluctuation time scales, of individual current time series are in the range of 1-4 days, being somewhat shorter during winter than summer. During summer, remarkably, the time series associated with the most energetic SF and NEP EOFS are not correlated between mooring sites. The lack of correlation in this relatively short comparison (84 days of overlap) is associated with a 50-100 day fluctuation at NEP during 1999 that is not present at SF. During winter, however, the lowest mode EOF time series at SF and NEP are well correlated (0.65), with NEP leading in time by 0.25 days.

3.4 Wind Forcing

3.4.1 General

The effects of surface wind stress forcing are summarized here in terms of the most energetic current EOF and the Georges Bank wind stress. By using correlations, we treat the response averaged across all frequency bands. For each case (summer/winter, SF/NEP), we rotate the wind stress vector to find which wind direction and time lag give the maximum correlation with the first current modal time series. In all cases, the maximum response is with winds oriented northeast-southwest, the general large-scale orientation of the coastline and roughly the orientation of the Bank itself (which is oriented about 30° north of east). At NEP, the maximum wind response occurs at zero time lag, with winds 35° north of east in the summer, and 65° in the winter. The NEP wind-EOF correlation is much weaker during summer (0.38) than winter (0.69), but is significant (all correlations cited in this paragraph are significant at 99.9% confidence) in both cases. The strength of the response can be judged by the regression between wind stress (in optimal orientation) and the shallowest current in the EOF, and it is strikingly independent of season, falling in the range of 3-4 (cm/sec)/(dyne/cm²). At SF, the optimal wind stress rotation is 60° north of east during summer and 55° in winter, and with a 0.25-day time lag during both seasons. Again, the summer currents are more weakly correlated with the wind stress (0.59) than during the winter (0.79), and, during both seasons, the SF response is stronger (about 6 (cm/sec)/(dyne/cm²)) than at NEP.

At both locations, wind forcing thus clearly plays a major role in subtidal variability. The present approach probably underestimates the total wind contribution, partly because it considers only local winds, and partly because it focuses on only a single EOF mode rather than the total velocity at each location. At SF, where the local isobath direction is well-defined, wind-driven effects are stronger and more obvious than at NEP, where the “alongshore” direction is much
more poorly defined. Drifter results (Brink et al., 2003) reveal a broad wind response pattern that has the scale of the Gulf of Maine, and is highly coherent with Boston adjusted coastal sea level: roughly speaking, the response appears to be that of an extremely wide, but otherwise uncomplicated continental shelf.

It is useful to put these Georges Bank results into the context of other coastal regions. Shearman and Lentz (2003) made moored current measurements southwest of Nantucket Shoals, about 200 km west of the SF site during fall 1996-spring 1997. Their location is only about 25 km west of where the continental shelf narrows because of the presence of Cape Cod/Nantucket Shoals, so we take this to be a transitional location. During the winter, their alongshore currents were most correlated (typically about 0.8) with wind stress rotated 45° north of east, and the regression with depth-averaged alongshore currents was $O(6 \text{ cm/sec})/(\text{dyne/cm}^2)$, quite comparable to our SF estimate. For comparison with a more “normal” shelf, we note, for example, that Winant et al. (1987) found typical wind-current correlations and regressions over the northern California continental shelf to be of $O(0.7)$ and $O(15 \text{ cm/sec})/(\text{dyne/cm}^2)$ respectively. Their regression is thus much larger than the values at either NEP or SF. Of course, one distinction about the U.S. west coast is that weather systems tend to move northward, in the resonant sense of coastal trapped waves (e.g., Halliwell and Allen, 1984), while in the Mid-Atlantic Bight, a less effective response would occur because weather systems generally move eastward (Ou et al., 1981), in the non-resonant sense (i.e., relative to coastal-trapped wave propagation, which is westward here). Nonetheless, Noble et al. (1983) investigated near-bottom current records from around Georges Bank and through the Mid-Atlantic Bight, and found a tendency for wind-current coupling to be stronger over the bounded New England shelf than over the Bank. Thus, it is likely that the proximity to a coastline, where Ekman transport is blocked, helps to determine the magnitude of wind response.

That the NEP current variability should differ from that at SF is perhaps not surprising, given its location at the curving, flatter, eastern end of the Bank. That summertime variability at NEP and SF is uncorrelated is astonishing given that both modal records are correlated with local wind stress (although with somewhat different orientations). Despite this curiosity, there is clearly pervasive wind-driven variability on the Bank that serves as a background to other types of current variability associated with offshore influences (section 4.3) and other causes.

Boston adjusted sea level (ASL) represents a longer and more continuous time series than those of Bank currents, and it is known to have a clear relation with currents over the Bank, in the sense expected for a coastal-trapped pressure signal extending over Gulf of Maine regional scales (e.g., Brink et al, 2003). In fact, the Boston adjusted sea level is very coherent with SF $u$ during the winter (and significantly, but less impressively so during the summer) over a range of frequencies (Figure 8 was computed using the longest uninterrupted interval where the time series overlap), and offers the considerable advantage of representing a longer and continuous time series. Boston ASL is also very coherent (coherence squared $> 0.8$) with NEP $u$ in the frequency range of 0.2-0.4 cpd during the winter of 1998-1999 (not shown), but the shorter records preclude a calculation as conclusive as that with SF data.

Boston adjusted sea level is thus treated as a continuous proxy for studying Bank-scale wind-driven current variations. Northeastward wind stress (rotated 40°, the large-scale alongshore
direction) gives maximum (as a function of orientation) correlation with Boston adjusted sea
level. Spectral calculations for wind stress and Boston summer and winter ASL are presented in
Figure 9. The spectra are computed by separately compositing all summertime (May 17-
November 17) and wintertime spectra for 1995-1999. Both wind stress and sea level fluctuations
are significantly (> 95% confidence) more energetic in winter than summer, and all records show
a significant “shoulder” of enhanced spectral energy around 0.1-0.3 cpd. Wintertime wind-sea
level coherence squared is considerably higher at all frequencies than in the summertime.

On a regional scale, wintertime adjusted sea levels from the Gulf of Maine to as far southwest
as Sandy Hook (New Jersey) show a weak northeastward propagation (as does the wind stress).
This behavior is consistent with wintertime variability being dominated by relatively local wind
forcing (e.g., Beardsley et al., 1977). During the summer, when wind-sea level coherence is
lower than during winter, sea level fluctuations show phase propagation toward the southwest at
O(1000 km/d), consistent with theoretical lowest mode long coastal-trapped wave propagation
rate of 820 km/day (using software described by Brink, 2006) for this area. This ability for
information to propagate southwestward is consistent with the possibility that Georges Bank
summertime current variability is more affected by remote, e.g., Scotian shelf, winds (see
Schwing 1992a, 1992b) than is the case during wintertime. Further, the coherence with remote
sea level records shows that variability over the Bank is part of a regional-scale wind response.

3.4.2 “Three-Four Day” Oscillations

Resolving currents as a function of frequency leads to some interesting findings. One striking
feature is a persistent “three-four day” oscillation that is detectable in moored along-isobath
currents. For example, NEP low-pass filtered currents from January-March, 1999 (Figure 10,
lower panels) show a distinct tendency, in both \( u \) and \( v \), for depth-independent oscillations with a
period of about three days. Although these oscillations are particularly obvious in velocity, they
also coincide with wind events and are apparent in Boston adjusted sea level (upper panels). The
three-day oscillations are much less obvious in SF current records (not shown). Spectral analysis
of the kinetic energy shows a distinct peak at around 0.3 cpd at NEP, along with a peak around
0.25 cpd at SF (Figure 11). A shorter 1999 record from north of the NEP site (mooring NFS: see
Smith et al., 2003) also shows a spectral peak centered near 0.3 cpd. At both SF and NEP, the
oscillations come and go with time, and complex demodulation shows that the oscillations at
both locations are much more energetic in winter than in summer. For example, complex
demodulation at NEP shows that an optimal (minimal net phase shift during the calculation)
period is 2.8 days, and that summertime amplitudes are about 1/3 of those observed during
winter. At SF, complex demodulation on currents shows that the oscillation changes frequency
from year to year over a narrower and lower frequency range of about 0.24 to 0.26 cpd.
Altogether, it seems clear that there is a preference, especially near the Northeast Peak, for
current energy to peak at a frequency in the range of 0.24-0.30 cpd.

If these current oscillations represent resonances, there ought to be a systematic phase shift
between forcing and response at the resonant frequency, where the sharpness of the shift depends
upon the strength of the damping. The long, continuous records of wind stress and Boston ASL
allow a detailed examination of this possibility. Wintertime coherence between these records has
very modest peaks at 0.250 and 0.35 cpd, while the summertime coherence peaks at 0.125 and 0.375 cpd (Figure 9). Phase behavior versus frequency is similar in both cases in that the phase does have a substantial change centered on the frequency of peak coherence. Summertime phases shift more rapidly (relative to wintertime phases) near the peak coherences, suggestive of weaker damping during this season. Similar cross-spectra, but computed on shorter segments, show sharper coherence peaks and more dramatic phase shifts at frequencies around 0.25 cpd. The tendency for the frequency of these peaks to move about over a few hundredths of a cpd appears to account for the less dramatic phase shifts in the composite spectral analysis, where frequency-varying signals will be “blurred out”.

A simple model of resonance quality (i.e., fitting the width of the resonant peak, e.g., Ohanian, 1985, page 357) can be used to estimate the frictional decay rate of the oscillations. During individual summers and winters, the phase changes by about 40º near 0.25 cpd, which is consistent with an amplitude damping time for the resonance of about 0.5-1.0 day. This estimate compares favorably to the range of damping times (about 0.5-0.9 days) obtained by Garrett (1972, 1974), based on models and tidal observations for the M2 tidal amplitude. The similarity of these damping rates adds credence to the idea that the three-four day oscillations could be a resonance phenomenon and that it has a decay process, presumably bottom friction, comparable with the M2 tidal case.

3.4.3 A Simple Numerical Experiment

As an attempt to understand what processes might contribute to the frequency selectivity in Georges Bank current response to the wind, we carried out numerical experiments with a linear, regional wind-driven model. Specifically, we used the ROMS model (Shchepetkin and McWilliams, 2005) with real bottom topography, representative summertime density stratification and a curvilinear grid that extends from near Halifax, Nova Scotia to near the mouth of the Chesapeake Bay. Resolution is 8 km or better in the alongshore direction and 7 km or better in the cross-shelf direction. We applied an idealized localized forcing consisting of smoothly varying alongshore winds that oscillate with time and that have peak amplitude on a cross-shelf line that passes offshore through the Gulf of Maine and across Georges Bank. No wind forcing is applied east or west of the Gulf of Maine area, but within that area, the northeastward wind stress is given by

\[ \tau^x = 0.5 A (1 + \cos(\xi)) \sin(\omega t), \]

(1)

where \( \xi \) is a stretched coordinate = ± π at the eastern and western boundaries of the Gulf (along the dashed white lines in Figure 12), \( \omega \) is the forcing frequency and the maximum stress \( A = 1.0 \) dyne/cm². Since we hypothesize a resonant response, the details of the wind forcing ought not to be too important. A full set of model runs (spanning forcing periods from 1.5 to 7 days) were run with both a full baroclinic model and with a barotropic (3 vertical grid points, uniform density and all stresses applied as body forces) version. The use of a barotropic model could be justified by shelf widths in the area typically being wide (> 100 km) relative to regional internal Rossby radii, but the main purpose was to see if any changes in response might occur due to changes in stratification. The following discussion treats the baroclinic model unless otherwise stated. No lateral viscosity is used in these runs, and the linearized bottom stress coefficient is 0.05 cm/sec.
Within the shallow Gulf of Maine, modeled current responses are in phase of the winds to within ± 0.2 day, and there is no sign of coastal trapped wave propagation along the Maine coast. Thus, the response along the northern and eastern Gulf of Maine is in a quasi-steady state for all frequencies considered (0.7 cpd and less). With weaker bottom friction (0.025 cm/sec), there is some sign of alongshore phase propagation along the Maine coast, but winds and currents are still typically in phase with the winds to within ± 0.3 days. In either case, the sea level over the entire western Gulf of Maine rises and falls in near synchrony with the wind stress. Away from the strongest forcing, the sea level variations partly propagate away from the Gulf through the Great South Channel and then along the continental shelf towards the west. The modeled Boston-Sandy Hook phase relation is consistent with westward free wave propagation at about 500-900 km/day, roughly consistent with observations (above). The Gulf of Maine sea level fluctuations also include an eastward propagating disturbance trapped at the abrupt depth change at the northern flank of Georges Bank.

The model shows a broad peak in NEP current response (Figure 13), centered at a forcing frequency of 0.4 cpd (0.32 cpd in the unstratified model, which may be more representative of wintertime conditions). With weaker friction, similar curves are obtained, but the peaks are somewhat higher and sharper. The modeled current amplitude response at this frequency has three distinct spatial maxima: near the Maine coast, over Nantucket Shoals and at the eastern end of Georges Bank (Figure 12). The time evolution of model fields shows that the Bank-edge response strengthens as it propagates eastward at a rate of about 60-80 km/day. The disturbance accumulates at the Northeast Peak and does not appear to propagate away. Indeed, model kinetic energy dissipation (due to bottom friction) has a strong maximum at this location. The propagation rate can be compared with a theoretical topographically trapped wave phase speed of 76 km/day (48 km/day in the barotropic case), computed, with open boundaries, using the coastal-trapped wave software described by Brink (2006). The 2.5-day peak is associated with a positive interference between the Bank-trapped wave as it reaches NEP and the regional-scale (non-propagating) response. At SF, the model has a peak response at about a 9-day period, although (unlike the observed energies: Figure 11 or Figures 6 and 7) the model SF response is weaker than the NEP wind response (Figure 13) for frequencies higher than 0.16 cpd. Thus, the numerical model seems to account for oscillations near the NEP mooring, but it fails to account for a response enhancement along the South Flank.

Why is the wave energy enhanced as it propagates toward the eastern end of the Bank? Calculations with a simple barotropic (straight-isobath) wave model show that both along-isobath propagation speed and maximum allowable frequency decrease as the net depth difference (Bank top to offshore basin) decrease. This is relevant to the model, since the top of the Bank becomes deeper toward the east, and the offshore depth becomes shallower approaching the Northeast Channel. If the wave’s propagation speed decreases eastward (the simple barotropic wave models suggest by a factor of 2-3), its amplitude must increase in order to conserve energy flux, and this could account for the modeled amplification. Similarly, if a wave (while conserving frequency) reaches a location with zero group velocity (a maximum frequency), then energy will accumulate at that location since it can not propagate away. A zero group velocity explanation would seem to account for there being a well-defined frequency of peak response, but idealized wave calculations that include density stratification do not appear to
have a subinertial frequency maximum (see Chapman, 1983 for a discussion of this in a coastal context). In the absence of stratification, either explanation (and they are not entirely distinct) could account for the modeled energy enhancement at the eastern tip of Georges Bank. We do not attempt to apply the simple, straight-isobath theory more quantitatively in this case because it can not match the greater complexity of the actual Bank’s topography.

One unsatisfying aspect of the model result is that it overpredicts NEP wind response relative to SF at periods longer than a few days. Numerical experiments with spatially uniform (larger-scale) winds do not differ qualitatively from those with the idealized wind (1), in that the SF response is always weaker and that the same response peaks are found. This tendency for stronger currents appears in both the stratified and unstratified models. When the model does not include the effects of density stratification, the main difference we would expect (e.g., Huthnance, 1978) and find is that the natural wave frequencies, hence phase speeds, would decrease. This, in turn explains why the barotropic model, even if generally correct, would underestimate response-peak frequencies, at least for the summertime.

Thus, we tentatively conclude that the observed NEP 3.2-day oscillation can be identified with the model’s 2.5 day (3.1 day for unstratified conditions) peak, and that this peak is, in turn associated with the along-isobath amplification of a slowing topographically trapped wave. The fact that the model’s prediction for peak response frequency depends on stratification may help to explain the observed tendency for the peak frequency to vary from time to time.

### 3.5 Monthly and longer time scales

Lower frequency current variations are explored by considering monthly mean SF currents, for those 22 months with sufficient data to create useful means. The 13- and 43 m currents at SF (Figure 14) can be readily compared with the Site A results from Butman and Beardsley (1987: BB87 hereafter) and GLOBEC drifter results (Brink et al., 2003). The present results agree with these earlier studies in that the strongest southwestward (along-bank) flows are found in the summertime, especially at 13 m, and the annual mean flows are comparable, around –10 cm/sec. The GLOBEC results also agree with BB87 in that 1) the monthly mean flows at 13 m are generally stronger than those at mid-depth and 2) standard deviations (computed monthly relative to the monthly along-bank means) peak in the wintertime, when wind variability is also strongest. All of our measures of cross-bank flow (13 m, 43 m, depth-averaged) have maximum off-bank flow during the wintertime, and show a tendency for on-bank flow during the summer (most obvious in 1995). This is exactly the opposite of the weak annual pattern measured by BB87, but we note that cross-shelf flows (especially means) are notoriously sensitive to small changes in the angular definition of the alongshore direction. Although one might expect that standard deviations of cross-bank flow, like those of along-bank flow, would peak during the wintertime, our observations do not support such a conclusion.

The aggregate summer and winter mean flows (Figures 6 and 7, left panels: averaging all summer ADCP records and all winter ADCP data from GLOBEC) observed at NEP and SF are also generally consistent with previous results (BB87; Brink et al., 2003) in that they are clearly part of a clockwise circulation around the Bank. At all depths, the mean flow is southwestward (along isobaths) at SF (Figure 6) and southward at NEP (Figure 7), although substantially weaker.
at NEP than at SF. There are no flow reversals with depth, although at both sites, the velocity veers with depth (either clockwise or counterclockwise), by perhaps 20º or less. Summer mean currents (upper left panels of Figures 6 and 7) at both locations are both stronger and more strongly sheared than winter means (lower left panels). The tendency for stronger near-surface mean flows during the summer has been noted before (e.g., Brink et al., 2003), and appears to be due to stronger stratification hence lateral density gradients during the summer (Chen and Beardsley, 1995).

The annual cycle of monthly mean currents computed from the ADCP data shows more detail (Figure 15a). The 95% confidence intervals on the mean are typically about 6 cm/sec in winter and 2 cm/sec in summer. Interestingly, the summertime along-bank flow is strongest toward the southwest (negative) at 20-35 m depth, rather than near the surface. Previous observations of the along-bank current on the south flank of Georges Bank did not have the vertical resolution to detect this subsurface maximum (Butman and Beardsley, 1987). During June, for example, the vertical shear over the upper 15 m is quite large, roughly $\partial v/\partial z = (2 \text{ cm/sec})/10 \text{ m}$. This is equivalent to a mean on-bank $\sigma_t$ increase of 0.2 over 10 km if the shear is in thermal wind balance. In contrast, the shear between 40 -70 m is consistent with an across-bank increase (decrease) in $\sigma_t$ in the deep (shallow) layers. We might expect such a subsurface along-bank current maximum near a summertime tidal mixing front, as is found farther on-bank of SF.

Although it is useful to compare local mean density gradients to the observed shears, existing direct hydrographic measurements do not yield much evidence for the expected cross-bank density gradients, presumably because of coarse spatial resolution. We thus take an indirect approach to estimating hydrographic gradients, inspired by Werner et al (2003), who plot the $T$, $S$ and $\sigma_t$ variations in time over a tidal cycle near the SF location and show that the changes are consistent with tidal advection and the known across-bank hydrographic gradients. We thus consider the tidal-band advection equation for temperature (for example)

$$\frac{\partial T'}{\partial t} + u'\frac{\partial T'}{\partial x} + v'\frac{\partial T'}{\partial y} + w'\frac{\partial T'}{\partial z} = Q' + M', \quad (2a)$$

where $(x, y, z)$ and the along-bank, cross-isobath and vertical directions respectively. Velocity components $(u, v, w)$ are in the $(x, y, z)$ directions, respectively, $T$ is temperature, $Q$ is the net source (heating), $M$ represents vertical mixing, a prime represents high-pass filtered (half power point of 1.96 days) data and $\{ \}$ indicates a mean over a one-month period. All terms are taken to be functions of the vertical coordinate. We assume that the temperature gradients themselves do not vary systematically with the tides, so that terms like $\{u'\partial T'/\partial x\}$ do not enter. The high-pass filtered data are all dominated by the $M_2$ tidal variations, so that $Q$ (mainly diurnal in this band) is incommensurate with the tidal period, so this term is neglected. The vertical velocity is estimated as

$$w' = (-z/h)\, v_B'\, \partial h/\partial y, \quad (2b)$$

where $h$ is the water depth and $v_B'$ is the on-bank current observed closest to the bottom. Since we know $T'$, $u'$, $v'$ and $\{\partial T/\partial z\}$ from mooring observations, the remaining unknowns are just the two horizontal gradient components that are representative of scales of about a tidal excursion, or 10-15 km at SF. Similar equations apply to salinity and density. It is then straightforward to
solve (2a) for $\partial T/\partial x$ and $\partial T/\partial y$, using a least squares fit, at all depths where moored hydrographic data exist. Separate fits are done using all of the January data, all of the February, etc.

Once the monthly cross-shelf density gradients are calculated, it is straightforward to integrate vertically (over the 5-6 depths where there are typically gradient estimates) and obtain the thermal wind along-isobath current relative to the bottom (Figure 15b). The thermal wind results are generally in good agreement with the directly measured monthly means, in that current magnitudes agree well in the summer, the general patterns of low shear in late winter and strong shear in later summer match, and both show a subsurface southwestward flow extremum in early summer (June and July). The main disagreement lies in the amplitude of the maximum mean flow in winter, when thermal wind leads to a substantial underestimate. This discrepancy is evidently due to relatively weak horizontal density gradients in winter and to a wintertime free surface tilt that is thus not reflected in the density structure. Stated another way, the bottom is evidently a much better reference level in summer than winter. We thus conclude that the early-summer subsurface current extremum is consistent with geostrophy.

Further, (2a) leads to salinity and temperature gradient estimates that are consistent with salinity increasing off-bank at depth and decreasing off-bank in the upper 10-15 m. Also, temperature decreases off-bank at depth and increases off-bank in the top 10-15 m at SF. We conclude that warm, fresh near-surface water exists south of the mooring during late spring and summer (see section 4.2.3 below) and that this water is transported southwestward along-bank by the mean flow.

Most existing hydrographic climatologies (e.g., National Oceanographic Data Center: NODC; GLOBEC, Marine Resources Monitoring, Assessment and Prediction: MARMAP; and National Marine Fisheries Service: NMFS) do not show a density structure consistent with that found using (2). The one exception is the Bedford Institute of Oceanography climatology (http://www.mar.dfonpo.gc.ca/science/ocean/scotia/ssmap.html) which shows a weak minimum in surface salinity during summertime over the south flank. The near absence of a surface salinity minimum in historical observations might be due to the relatively wide spacing of hydrographic stations on traditional surveys. Another corroboration is a high-resolution “snapshot” SeaSoar survey, conducted near the SF site during June, 1999 by J. Barth on R/V Oceanus Cruise 343 (http://globec.whoi.edu/jg/serv/globec/gb/process/1999/seasoar_ctd.brev1?cruiseid%20eq%20OC_343), which shows the anticipated near-surface salinity minimum and temperature maximum, but somewhat farther on-bank than (2) suggests. Averaging Barth’s eight local synoptic cross-bank sections yields a mean density field that is qualitatively consistent with the density gradients derived from (2) (although, again, with a spatial offset).

3.6 Mean Flow Generation

Butman et al. (1983) pointed out that fortnightly and monthly modulations in along-bank currents can be used to test tidal rectification theories such as that of Loder (1980). They attempted to test this at site A (near our SF site), but were unable to confirm the theory, partly because the bottom is so gently sloping that only a weak rectified flow is expected, and partly because their current record was too short to obtain statistically significant amplitudes for both
fortnightly and monthly tides. We attempt to improve on their analysis by using 917 days of SF ADCP data to carry out a least-squares fit at fortnightly (354 hour) and monthly (661 hours) periods, for both depth-averaged and individual-depth SF and NEP low-pass filtered current records. The amplitudes of these low-frequency fluctuations would then be indicative of the strength of the tidally rectified mean flow. Our results are similarly inconclusive to those of Butman et al. (1983): in no case was a low-frequency tidal flow modulation found that was significantly non-zero at 95% confidence, and only a few records were found where a coefficient was non-zero at 67% confidence. We repeated the analysis using the 2142-day merged record at 45 m from the SF, ST1 and Butman (USGS) with only a modest improvement: the regression coefficient relating the alongshore flow to the monthly cosine component is \(-0.61 \pm 0.32\) cm/sec at 67% confidence, or nonzero at almost 95% confidence. We can confidently conclude that, at SF (NEP), these low-frequency modulations are less than about 0.9 (1.3) cm/sec in amplitude, and that the tidally driven mean flow at SF is thus about 1.2 cm/sec and no more than about 2 cm/sec. These numbers are not inconsistent with the Butman et al. (1983) theory, which predicts that the modulations ought to be less than 0.5 cm/sec and that the mean flow associated with tidal rectification is about 1 cm/sec here. Again, these extremely weak rectification effects are expected over such a gentle bottom slope.

The long current record at SF also provides a test of the “Neptune” concept: that current variability above a sloping, but bumpy, bottom can be rectified into a mean along-isobath flow (e.g., Haidvogel and Brink, 1986; Holloway, et al., 1989; Holloway, 2002). Viewing the theory in its simplest form, enhanced current variability at a given location ought to be correlated with strengthened along-isobath (southwestward, \(u < 0\), in this case) flow. The existing literature does not provide grounds to estimate how strong the mean flow or modulation would be. Since wind variability is greatest in the wintertime, one might expect strongest monthly mean flows during the winter: this is not obviously the case (see section 3.5). Instead, the monthly mean along-isobath flow is significantly (95%) correlated with monthly mean along-bank wind stress (0.48), and the monthly standard deviations relative to monthly means are also well correlated with wind stress standard deviation (0.60). Although there is a negative correlation between standard deviation of cross-bank flow and mean along-bank flow (-0.32), it is not significant, and it can be accounted for by the fact that monthly means and standard deviations of along-bank wind stress are well correlated (-0.52). Thus, although there is the type of correlation between variability and mean alongshore flow one might expect from Neptune, the correlation between monthly mean winds and wind variability provides a suitable explanation for the Neptune-like correlation. We conclude that the SF current data lend no clear support to topographic rectification theories. There is also nothing here that invalidates the theory either, since the bottom slope at this mooring location is not great, and the wind-driven effects may simply be masking a weaker topographic rectification.

4. Hydrographic Time Series

4.1 Introduction

Long moored time series of temperature and salinity were obtained at SF and NEP with the primary goal of identifying and characterizing multi-day to seasonal to interannual changes in
water mass types that could potentially be associated with important biological changes, such as importation of nutrients (e.g., Churchill et al., 2003) or plankton. For background, it is important to review the types of water masses (Figure 16) that are likely to be observed at these sites, and what their properties might be (http://www.mar.dfo-mpo.gc.ca/science/ocean/scotia/ssmap.html provides an excellent climatology). The freshest, coldest water at any time of the year is associated with “cross-over” events from the Scotian Shelf: temperatures lower (<5 ºC in winter and < 8 ºC in summer) than those found on Georges Bank and salinities of 32 or less (Bisagni et al., 1996). Another fresh end member found on the Bank during the summertime is that associated with the western Gulf of Maine (to be identified with the “Maine Surface Water” of Hopkins and Garfield, 1979): these waters have been warmed at the surface, but are still quite fresh due to their Scotian Shelf origin and to the further addition of Gulf of Maine river runoff. We take the summertime core waters to have salinities around 32 or less (Flagg, 1987, defines Maine Surface Water salinities as 31.6-33) and, at the surface, to be typically warmer than those on Georges Bank (e.g., 17 ºC versus 12 ºC in the summer: see Figure 11). Unstratified central Georges Bank waters usually have salinities in the range of 32 to 33.2, and temperatures, while seasonally variable, tend to be lower than those in surrounding waters (Flagg, 1987). Shelf-slope frontal water and slope water, found offshore of the shelfbreak front, are relatively saltier (33.6 to 35.0) (e.g., Lentz et al., 2003). The warmest (up to almost 25 ºC) water found in the area is associated with Warm Core Rings, whose properties strongly resemble those of Gulf Stream waters, having salinities above 35.5 (e.g., Flagg, 1987; Brink and Lee, 2006). For many (but not all) purposes, salinity by itself is a good discriminator of water type, and fresher water tends to be lighter than saltier water throughout the year. It is thus relatively easy to detect, and often to identify, intrusions of different water masses at a particular site.

4.2 Events

4.2.1 Introduction

Three primary types of events are discussed below. These are associated with Scotian Shelf crossovers, western Gulf of Maine waters, and salinity increases due to offshore influences. The passage of Hurricane Edward in September 1996 is a special case and is already treated by Williams et al. (2001). The offshore influences can take many forms, including onshore motion of the shelfbreak front, slope water intrusions and Warm Core Ring intrusions. The frequency of these events is summarized in Table 4 for both of the mooring locations, but this table should not be taken to provide more than order of magnitude information, since some subjectivity was involved in defining events, and the data coverage is not uniform, either vertically or in time. Further uncertainties are also due to the possibility that a single mooring can miss at least some finite-sized features as they pass nearby. Ashjian et al. (2001) provide interesting biological context for several of the classes of features that we discuss here.

In the following discussion, the reader should bear in mind that there are several biases in the data sets (Figure 2). For example, the fall period is less well-sampled. Likewise, at SF, salinity information was only obtained below 50 m about 70% of the time that salinity data were available shallower in the water column. Finally, the summary information is based on a qualitative inspection of the observed time series, so that weak events (e.g. with salinity changes less than 0.1) are not accounted for.
4.2.2 Scotian Shelf Water Crossovers

Scotian Shelf water crossover events are characterized by relatively fresh and cool upper ocean water passing across the Northeast Channel and onto Georges Bank (e.g. Bisagni et al., 1996; Smith et al., 2003). Study of satellite images or hydrographic climatologies (http://www.mar.dfo-mpo.gc.ca/science/ocean/scotia/ssmap.html) show that Scotian Shelf water is almost always cooler than Georges Bank water (Figure 17), so that we use surface temperature to discriminate between Scotian Shelf and the warmer western Gulf of Maine surface waters.

With one possible exception, all of the crossover events we detected occur between December and April, when background stratification on the Bank is weak or nonexistent. (The one ambiguous event, in July 1999, is more likely a Gulf of Maine intrusion, and is described separately below). Likewise, Smith et al. (2001) suggest, based on current observations, that Scotian Shelf crossovers ought to be most prevalent during the winter. A typical event at NEP is illustrated in Figure 18, and shows the fresh anomaly most pronounced in the upper 30 m of the water column, where it temporarily introduces stratification. The crossover events do not appear to be systematically related to anomalies in currents, winds or sea level. This lack of correlation with velocity could be consistent with dynamically passive (density compensated) patches being passively advected by the ambient flow, but the crossover waters are generally associated with a distinct decrease in density (hence, increase in stratification during an otherwise homogenized time of year). A passive advection explanation thus seems unlikely, and it is more likely that the lack of correlation is due to the complex spatial structure of either the density or velocity field. Of the 11 crossover events observed at the two moorings, 9 are confined to the upper 30 m or less of the water column, while two (February and March 1996) occur throughout the water column as might be expected with deep wintertime mixing. The observed crossover events have maximum salinity and temperature decreases of 0.2-1.0 and 0.3-1.5 ºC, respectively. There was no obvious tendency for the water mass anomalies to be stronger at NEP than at SF, although the events occurred more frequently at NEP than at SF (Table 4). Smith et al. (2003), based on their study of 1999 cross-over events, conclude that their most likely cause is mesoscale activity that sweeps fresher water parcels south or southwestward across the Northeast Channel, and this finding is consistent with our longer but sparser data set. Many aspects of the cross-over events are still not clearly understood, such as the apparent tendency for these intrusions only to be seen in the mooring data during winter.

4.2.3 Western Gulf of Maine Water

We find unambiguous western Gulf of Maine water to be fresher and warmer than ambient Bank water. These intruding waters originally come from the Scotian shelf, but circulate counter-clockwise around the Gulf of Maine, where they are freshened to some degree by New England river outflows, and warmed during the summertime. Very often, in the summer, the warmest waters in the region are found offshore of Boston and extending eastward toward and around the Bank (e.g. Figure 17 or http://www.mar.dfo-mpo.gc.ca/science/ocean/scotia/ssmap.html). These waters are typically quite fresh as well. It is possible that, during the winter, the Gulf provides some of the fresh, cool intrusions that we have associated with the Scotian Shelf, but there is good precedent (e.g. Bisagni et al., 1996) to believe that the cooler intrusions generally reach the...
Bank from the east. One ambiguous case occurred in July, 1999, when fresh, warm water appeared in the upper 15 m at NEP shortly after the onset of density stratification. It is conceivable that this is Scotian Shelf water (initially cooler) that has warmed before reaching the NEP site. We nonetheless provisionally attribute this intrusion to a Gulf of Maine source, since that is the simpler explanation for warm water.

A representative example of a western Gulf intrusion at NEP is shown in Figure 19. All of these events observed at the moorings are confined to the upper 15 m at NEP and the upper 35 m at SF. These intrusions occur about as frequently at NEP as at SF (Table 4). Associated anomalies can range up to +5 °C in temperature and down to −0.8 in salinity so that these events invariably tend to stabilize the upper portion of the water column. There is again no systematic coincidence of these fresh events with winds or currents. Evidence for the path of the intrusions can be found in a warm tendril in climatological sea surface temperature that reaches around the northern and then eastern end of the Bank (Figure 17) and tapers to a lateral width of O(10km). Individual April-October satellite snapshots of sea surface temperature often show this narrow tendril extending around the Bank past the SF site. In some early spring cases, the intrusions provide a temporary springtime stratification that vanishes as the patch is advected away.

The most dramatic Gulf of Maine intrusion event (see Figure 24, middle panel, below) was observed for 41 days at SF during July-August 1998, accounting for the freshest water observed at SF, and following record-breaking stream flow volumes in Maine during June 1998. Butman et al. (2004) document this same event by relating Merrimack River outflow and upper ocean salinity records observed near Boston. A similar flood event in April-May, 1996 is also reflected in the Butman et al. (2004) records and, about a month later, in upper-ocean SF salinity (Figure 24). Thus, extreme flooding events in northern New England can increase the freshwater transport in the Maine Coastal Current, and eventually give rise to unusually low salinities over the Bank. The 1- to 1.5-month lag time between peak stream flow and freshening at SF is consistent with advection at a reasonable speed of 10-20 cm/sec. This type of event, while dramatic, is evidently also rather rare, and so its cumulative importance to the Bank salinity balance is unclear.

The concentration of observed SF Gulf of Maine intrusions in the warm months makes sense because a) the clockwise circulation around the Bank is distinctly stronger during the stratified season (e.g. Limeburner and Beardsley, 1996; Brink et al., 2003), and b) these events are more readily identified during the summertime (vs. the winter, when surface temperatures do not vary so much in space), when western Gulf of Maine water is unambiguously warmer than Bank surface waters.

### 4.2.4 Offshore Influences

This category includes all events involving warmer, saltier water, which comes presumably from offshore. There are several categories of saltier water (e.g. Churchill, et al., 2003), including shelfbreak frontal water, slope waters, and very salty Warm Core Ring waters: all of these evidently are observed at different times. Potential ambiguity is also introduced because of the presence of salty water (> 34) at depths greater than around 100 m in the Gulf of Maine, so it is possible that deep salty anomalies could originate from upwelled Gulf of Maine waters,
especially at NEP, which is relatively close to the Gulf source. We do not believe any of our salty events originate from the Gulf, since these salty Gulf waters are associated with very low temperatures (e.g., 7-8 °C), and all of our salty events observed at NEP are associated with either slight temperature change or with warming. Further, salty events occur much more frequently at SF, which is relatively close to offshore saltier waters, than at NEP, which is closer to the Gulf of Maine (Table 4).

The most dramatic salty intrusions are those associated with Warm Core Ring invasions (e.g. Figure 20), where temperatures can increase temporarily by up to 13 °C, and salinities by up to 3.8. This example (July, 1997) was intensively mapped out by Brink and Lee (2009), who showed that water having almost undiluted Gulf Stream properties is found as far on-bank as about the 70-m isobath (Figure 21). In some places, the intrusion water is found at all depths, but often the salty water is concentrated in the upper half of the shelf water column. At the SF site, the intrusion is apparent for about 10 days at all depths and is followed by several 2-5 day pulses in only the upper 30 m over the next 30 days. At the end of this pulsing period, there is no obvious sign of salty water having been left behind on the Bank. The pulsing time scales are not surprising, because of the 20-40 km scale intrusion and the ambient O(10 cm/sec) currents. Another, slightly less impressive (temperature and salinity increases of 9 °C and 3), 11-day Warm Core Ring intrusion occurs at SF during the fall of 1995, but the data coverage at that time is poorer. There are no data available at NEP during either of the Warm Core Ring intrusions measured at SF, although the 1997 event does not appear to reach anywhere near the NEP site (Brink and Lee, 2009). Finally, during this event, there is also some evidence in the satellite imagery (figure 21) that water is being drawn off the Bank east of the onshore intrusion, but this expulsion is entirely undetected by the moored data.

Bottom-trapped salinity increases, which extend upward no more than 30 m, are found three times at SF (e.g., Figure 22) and twice at NEP. Twice, these are associated with temperature increases (2-3 °C) and twice with decreases (1-3 °C), while salinity increases by 0.2-1.0 in each case. The events each last from 1-5 days at NEP and 13-21 days at SF. A credible explanation of these changes would be that the foot (at least) of the shelfbreak front moves into shallower water. Representative cross-bank sections (Flagg, 1987) show that temperatures should normally be expected to increase (going from the Bank’s subsurface cold band offshore into slope waters) as salinity increases (frontal signature moves to shallower water), but there are cases with cooler water at depth offshore of the frontal gradients so that cooling could potentially coincide with increasing salinity when the front moves on-bank. It is tempting to compare these deeper, salty intrusions to those repeatedly observed in the autumn by Mountain et al. (1989) in the Great South Channel, but our features do not appear at any preferred time of year.

Finally, there are five events where salinity and temperature both increase by 0.4-2.0 and 0.3-6 °C, either at all observed depths (four cases, e.g. Figure 23), or, in one case, where the perturbation is only observed in the upper 40 m of the water column. These events occur throughout the year and are tentatively associated with onshore intrusions of frontal or slope water, or perhaps of diluted or less dramatic Warm Core Rings. In each case, when the intrusion withdraws, no obvious residual of the slope water is left behind on the Bank.
One of the most dramatic forms of onshore intrusions are the “SMax” intrusions (e.g., Lentz, 2003) that are 20-50 m thick salty tongues extending into shallower water along the seasonal pycnocline. These features are strikingly absent in our measurements. One reason may be their geographical scarcity here: Lentz (2003) shows that they are to be found in less than 5% of hydrographic sections in this area, and that the frequency of occurrence might be expected to drop by roughly another factor of two between the 100 m isobath and the SF mooring’s 76-m isobath. Further, the moorings have their temperature and conductivity sampling concentrated in the upper 50 m, so we would not anticipate that the whole feature would often be resolved. Even with these qualifications, we find the absence of even a single clear SMax intrusion event to be striking.

4.2.5 Conclusions on Events

A wide range of temperature/salinity events occurs at the two mooring sites. Given the proximity to the source waters, it is probably not surprising that Warm Core Ring events seem to be more prevalent at SF than NEP, and that the Scotian Shelf and western Gulf water appear more at the NEP site. One of the most striking aspects of intrusion events is that the hydrographic variability shows so little relation to wind stress or (except in the case of a Warm Core Ring intrusion) observed currents. Perhaps the most curious aspect of the events is that the various intrusions come and go without leaving behind any obvious water mass changes. This does not mean that the intruded water simply sloshes back off the Bank. Rather, the water parcels are continually moving (mainly along isobaths), so that a patch of a few tens of km’s in size will rapidly pass by a particular fixed site. Thus, moorings are not well adapted to sample the evolution and mixing of a patch. Further, once a single patch does mix in with its surroundings, its individual contribution to the whole Bank’s freshwater or saltwater inventory would be so small as to be difficult to measure. It would be instructive to estimate the net potential contribution of the observed intrusions to Bank-wide property balances, but this is indeed difficult to do in the absence of more specific information on patch size. What we can say with confidence is that, at our two sites, which we take to be somewhat representative, “foreign” water masses can be observed about 20-25% of the time.

4.3 Seasonal and Longer Time Scales

4.3.1 Introduction

We investigate the seasonal and interannual variability in temperature, salinity and density ($\sigma_t$) on the southern flank of the Bank using the five-year SF hydrographic time series. While these time series contain some gaps, sufficient data are available to determine mean seasonal cycles and identify longer time scale variations that dominate the GLOBEC field period.

The daily-mean SF temperature, salinity, and density ($\sigma_t$) time series at all measurement depths are shown in Fig. 24 for 1994-1999. There are clear seasonal (annual) cycles in both temperature and density but not salinity at SF. The surface layer temperature experiences the largest seasonal change, with warming from the late winter minimum starting around mid-March and reaching a maximum around late August. This is followed by wind-driven mixing and surface cooling in late fall and winter that destroys the seasonal thermocline by late October with
continued water column cooling until the start of the next warming cycle in March. While local
density depends significantly on both temperature and salinity throughout the year, the lack of a
substantial seasonal cycle in salinity causes the density cycle to track the temperature cycle
closely. Thus, the surface layer becomes less dense starting around late February and reaches its
lowest values around late August, before becoming denser in late fall until maximum values are
reached around late February. As would be expected, the depth-averaged temperature, salinity,
and density time series at SF (Fig. 25) exhibit distinct seasonal cycles in water-column
temperature and density, with minimum and maximum temperatures (and maximum and
minimum densities) occurring in March and late August, respectively.

Lack of a distinct seasonal cycle in salinity at SF is perhaps surprising, since the top of the
Bank exhibits a clear cycle with maximum (minimum) salinity of 33.0 (32.4) during late March
(late September) based on historical hydrographic data (Flagg, 1987; Lentz et al., 2003a;
Mountain, personal correspondence). However, neither Lentz et al. (2003a) nor Mountain found
a salinity seasonal cycle over the southern flank in the 60- to 100-m depth range, consistent with
our results. The mean salinity (in the top 30-m) at SF is 32.5 ± 0.5, similar to Mountain’s mean
value of 32.8 ± 0.2 at two stations on the 68-m isobath near SF.

The SF time series also exhibit significant interannual variability, specifically in the form of
1) notable differences in the temperature and density annual cycle amplitudes, and 2) occurrence
and strength of the intrusive events described above (section 4.2). The salinity variations are
dominated by event and lower frequency variability. In addition to the high-salinity intrusion
events in 1995 and 1997, water-column salinity exhibits a series of freshening steps, in 1995 and
1996, then a small increase in 1997, followed by a large decrease in 1998 to its lowest value of
31.8 before increasing back to 32.6 in August, 1999. This overall drop in salinity of about 1
between early 1995 and mid-1998 (before a partial recovery by mid-1999 of about 0.8) reflects a
large change in the upstream source water salinity during this period (Smith et al., 2001;
Mountain, 2004), which in combination with the regional surface heat flux causes a large (order
0.5 kg/m³) low frequency (subannual) change in density at SF. The resulting SF density seasonal
cycle has more year-to-year variation than temperature, which is set by the more repeatable
regional surface heat flux.

To further illustrate the dominance of the seasonal cycles in temperature and density but
lower frequency variability in salinity, we compute periodograms for the 10-, 20-, 30-, and 40 m
records (which have good seasonal coverage during the 4.8-yr GLOBEC field program) (Figure
26). These are computed by least-squares fitting with a fundamental frequency $f_i = 0.2$ cycle per
year (cpy) and a total of 10 frequencies. Both temperature and density periodograms exhibit
pronounced peaks at 1 cpy with a second but much weaker peak at 2 cpy, and a general decrease
in amplitude in increasing depth. The salinity periodogram is more “red”, with more variance at
lower frequencies (especially below 1 cpy) and no clear peaks at 1 and 2 cpy. The resulting
density periodogram reflects salinity’s increased low-frequency variability, particularly in
relation to the annual peak at 1 cpy.

4.3.2 Seasonal Cycle
Seasonal cycles for temperature and density are presented based on Fourier analysis using a fundamental frequency $f_1 = 1$ cpy and two harmonics (2 and 3 cpy). This approach focuses on the average annual cycle with the interannual and event timescale variability reflected in the uncertainty in the Fourier coefficients. A structure function approach (e.g., Kosro, 1987) was used to estimate the independence time scales ($T_I$) for the residual time series at the 10-, 20-, 30-, and 40 m levels. Gaps in the time series were filled using interpolation of data measured immediately above and below the gap. Deviations were computed as the difference between the observed values and the Fourier series fit (using $f_i = 0.2$ cpy and ten frequencies to remove the low frequency variability) at each time. The temperature, salinity, and density structure functions all showed reduced variance at greater depth, with a rough plateau for lags between about 20 and 45 days. Using the mean of this plateau to scale the structure function, we estimate the record-average $T_I$'s to be about 6-11 days for temperature, 8-12 days for salinity, and 15-20 days for density, with smaller values at depth, consistent with the variance decrease with depth. Uncertainties in the fits are then computed using these $T_I$'s, the standard deviations between the time series and fitted annual cycles, and record lengths. The resulting 95% confidence limits are then approximately ±0.3 °C and ±0.06 kg/m³ for temperature and density, respectively.

The resulting seasonal cycles in temperature and density are shown in Fig. 27 and 28. The water column cools with very little vertical stratification during winter and reaches a minimum temperature of about 4.5 °C around the third week in March. The water column then starts to warm, with more rapid warming in the upper 30 m becoming obvious by May. Temperatures at all levels continue to increase, with the surface layers (1-5 m) reaching a maximum of about 17.5 °C in late August before decreasing while the deeper layers reach their lower maximum values later in the season, e.g., the lower half of the water column (≥ 40 m) reaches its maxima in early October, about five weeks after the surface layer maximum (Table 5). By late October, temperatures in the entire water column are falling, with very little vertical structure observed after November. The processes that contribute to this “de-stratification”, vertical mixing driven by winds, surface cooling, and advection are discussed in section 4.4.2. Vertical mixing by the tidal currents at SF is confined to the bottom boundary layer that reaches a maximum scale height of about 25 m during spring tide in the nearly unstratified conditions in winter (Werner et al, 2003).

The density seasonal cycle is similar in structure and timing, especially during the May-November period of first formation and then destruction of strong temperature stratification (Table 5). The primary difference occurs during late winter, when water column density reaches a maximum of about 25.9 in late February, roughly three weeks before the minimum water column temperature occurs. This late winter difference appears real, but the summer density inversion between 5-10 m and the winter inversion between 50-72 m are roughly within the density uncertainty and likely not significant. While the cause of this late winter phase difference is unclear, there are two potential explanations. First, the large interannual changes in density (caused by the large low-frequency variation in salinity) could cause more uncertainty in the Fourier analysis, especially in the February-March period of maximum density. Second, there could be a weak seasonal cycle in salinity, with the water column freshening during February-March enough to allow the upper layers to start to become less dense before the temperature starts to increase. A change in salinity of about 0.1 would be sufficient to delay the time of maximum density to late March. Such a small change could easily escape detection in the
historical hydrographic data and in the SF moored data in view of the large low-frequency salinity variability.

4.4 Heat Budget

4.4.1 Introduction

The 1995 GLOBEC Stratification Study was designed to investigate the late-winter-summer evolution of stratification on the southern flank and identify the physical causes. The ST1 mooring, located 23 km southwest of the SF site along the same isobath, was instrumented to measure water-column currents, temperature and salinity and the surface momentum, heat, and water fluxes. More lightly instrumented moorings were deployed at ST2 (located 12 km on-bank of ST1) and on the crest (CR) and Northeast Peak (NEP). Lentz et al (2003a) used the ST1, ST2 and SF measurements to describe in detail the temperature and salt balances at ST1 for the period February-August 1995.

Georges Bank and the Gulf of Maine experience a large seasonal variation in surface heat flux, which has long been thought to drive the large seasonal cycle in temperature (e.g. Bigelow 1927). To test this idea, Lentz et al (2003a) examined the depth-averaged heat balance to determine the relative contributions of surface heating/cooling and horizontal advection to changes in heat content. The heat balance, expressed as temperature, is

\[ \langle T \rangle + \langle uT \rangle + \langle vT \rangle + \langle wT \rangle = \frac{Q}{\rho_o c_p h}, \]

where \( T \) is temperature, \( u, v, \) and \( w \) are the along-bank, on-bank, and vertical velocity components, \( \langle \rangle \) indicates an average from surface to bottom, \( Q \) the surface heat flux, \( \rho_o a \) reference density \((1025 \text{ kg/m}^3)\), \( c_p \) the heat capacity of seawater \((3999 \text{ J kg}^{-1}{}^\circ C^{-1})\), and \( h \) the water depth \((76 \text{ m})\). PAR profile measurements made over the bank during May-June 1995 indicate that less than 0.1% of the surface shortwave radiation penetrates beyond 40 m, thus eliminating direct bottom warming in (3) (Chen et al, 2005). Following Lentz et al (2003a) to focus on the longer time scales, the temperature balance (3) is integrated forward in time to get

\[ \langle T \rangle = \langle T_i \rangle + Q_{\text{cum}} - \int_0^t \left( \langle uT \rangle + \langle vT \rangle + \langle wT \rangle \right) dt', \]

where \( T_i \) is the initial temperature,

\[ Q_{\text{cum}} = \int_0^t \frac{Q}{\rho_o c_p h} dt', \]

is the cumulative surface heat (expressed in temperature) flux, and the last terms in (4) represent the cumulative horizontal and vertical heat fluxes, of which the horizontal components could be estimated using data from the along-bank (ST1-SF) and on-bank (ST1-ST2) moorings.

Lentz et al (2003a) found “close” agreement between \( \langle T \rangle \) and \( Q_{\text{cum}} + \langle T_i \rangle \), the one-dimensional (1-D) heat balance, with “most” of the depth-averaged seasonal temperature variation at ST1 between February-August 1995 caused by surface heating/cooling. Horizontal advection did cause shorter time scale (days to weeks) variations, notably a wind-driven cooling event in February and intrusive events of Shelf/Slope Frontal Water (SSFW) in May and August (similar in character to those described in section 4.2). They also concluded that the strong tidal
currents did not contribute significantly to the horizontal advective temperature flux, so we have used daily mean \( T \) and \( Q \) data in our analysis.

### 4.4.2 SF Analysis

Here we extend this analysis using the longer SF measurements to see if the 1-D heat balance also applies at lowest order during the fall-winter period of the year and to other years. Without nearby mooring data, we can not directly estimate advective fluxes so our analysis is limited to comparisons of \( \langle T \rangle \) and \( Q_{\text{cum}} \) and consider any significant differences to be due to advection.

Meteorological measurements were made at SF during the 1996-1999 period, however, large gaps in some measurements (e.g., incident long-wave radiation) occurred and the resulting surface wind stress and heat flux time series (Fig. 29) have large breaks in coverage. Despite these breaks, the SF net surface heat flux time series \( Q \) exhibits a clear seasonal variation, with shortwave warming dominant in spring-summer and latent and sensible cooling important in fall-winter due in part to increased winds (especially during “Nor-easters” and other lows that carry cold, dry continental air offshore) (Beardsley et al, 2003). Due to the gaps in \( Q \), we use the continuous time series \( Q_{\text{MM5}} \) at SF for the heat budget analysis. These two time series compare reasonably well, with \( \bar{Q} - Q_{\text{MM5}} = -21 \pm 66 \text{ W/m}^2 \) (correlation = 0.92 over the 1160 days of comparison). The \( Q_{\text{MM5}} \) annual cycle (Fig. 30) is relatively symmetric about an annual mean flux of 53 W/m², with a mean heating rate of about 190 W/m² during April through August, a mean cooling rate of about -95 W/m² during November through February, and with weak mean fluxes in the March and September transition months. Due to the increase in storms in fall through winter, the variability in daily and monthly flux is greater in winter than summer.

The mean water column warming/cooling rate is defined here as \( \bar{Q}_{\text{wc}} = k\Delta \langle T \rangle / \Delta t \) and the mean surface heat flux as \( \bar{Q}_{\text{sf}} = k\Delta Q_{\text{cum}} \) where \( \Delta \) is the change over a given time interval \( \Delta t \) and \( k = \rho c_p h \). The mean advective heat flux is \( \bar{Q}_{\text{adv}} = \bar{Q}_{\text{wc}} - \bar{Q}_{\text{sf}} \). The experimental uncertainties in the computed fluxes vary with both season and length of \( \Delta t \). The uncertainty in \( \langle T \rangle \) is due to individual sensor error (± 0.05 °C; Lentz et al, 2003a), the number of sensors used in the vertical average (6 to 13), and the sampling error associated with the difference between the true temperature profile and the interpolated profile. Assuming the sensor errors are independent, the combined sensor error in \( \langle T \rangle \) is about ±0.02 °C while the sampling error can vary from roughly ± 0.01 °C in winter (when the water column has little vertical thermal stratification) to ± 0.1 °C in summer (when thermal stratification is largest). The resulting uncertainty in \( \bar{Q}_{\text{wc}} \) is about ± 5 W/m² in winter increasing to about ± 20 W/m² in summer for \( \Delta t = 30 \text{ days} \). The experimental uncertainty in \( \bar{Q}_{\text{sf}} \) also varies with \( \Delta t \). The maximum uncertainty in hourly \( Q \) at ST1 due to sensor and other errors is roughly ± 35 W/m² (Beardsley et al., 2003). Assuming similar uncertainty in the SF hourly surface heat flux and independent hourly errors, the uncertainty in daily mean \( Q \) is about ± 7 W/m², with uncertainty in \( \bar{Q}_{\text{sf}} \) decreasing further with increasing \( \Delta t \). Since it seems unlikely that errors in hourly \( Q \) estimates are completely independent, the actual uncertainty is probably larger. For lack of
better knowledge, we will consider the uncertainty in $\overline{Q}_{sf}$ to be roughly ± 10-15 W/m² independent of analysis interval, and round off all estimates to the nearest 5 W/m².

The SF $\langle T \rangle$ time series has been combined into ten “legs” for analysis (Table 6). Figure 31 compares these legs with $Q_{\text{cum}}^{\text{MM5}}$ (which has been detrended to remove long-term biases and shifted vertically by 7.5 °C to facilitate visual comparison). The tendency for $\langle T \rangle$ to follow $Q_{\text{cum}}^{\text{MM5}}$ during the late spring/summer months is clear in all five years (legs 2, 4, 7, 9, 10), with the primary heat balance being 1-D. Shorter advective warming and cooling events occurred during the late spring/summer months in some years (especially notable in 1995 and 1997) (see section 4.2). During the fall/winter months, $\langle T \rangle$ tends to drop from its seasonal high more rapidly than $Q_{\text{cum}}^{\text{MM5}}$, indicating that both surface and advective cooling contribute to water column cooling. Fall/winter storms with high winds can drive both strong surface cooling and strong currents (Brink et al. 2003). Strong advective cooling can also occur on short time scales when wind forcing is relatively weak. Figure 32 shows three examples of very strong advective cooling events lasting 7, 1, and 3 days. On October 10-17 1998, the water column temperature dropped rapidly from 13.1 to 10.4 °C, resulting in $\overline{Q}_{wc} = -1340$ W/m², $\overline{Q}_{sf} = -55$ W/m², and $\overline{Q}_{\text{adv}} = -1285$ W/m², with almost all the cooling due to advection. The largest advective cooling event occurred November 26-27 1998, with $\overline{Q}_{wc} = -2615$ W/m², $\overline{Q}_{sf} = -25$ W/m², and $\overline{Q}_{\text{adv}} = -2590$ W/m². A weaker but classic winter intense cooling event occurred during January 29-February 2 1999 when a strong southward wind pulse of about 0.4 N/m², the water column remained vertically well-mixed and became colder, more saline and denser, with $\overline{Q}_{wc} = -815$ W/m², $\overline{Q}_{sf} = -345$ W/m², and $\overline{Q}_{\text{adv}} = -470$ W/m².

These SF heat budget comparisons (Table 6, Figs. 31-32) suggest the following conclusions. First, during the period from about mid-April through September when the seasonal thermal stratification is established, water column warming is due primarily to surface heating on time scales of monthly and longer, i.e., consistent with the primarily one-dimensional heat budget paradigm found at ST1 by Lentz et al. (2003a). On shorter time scales, deviations from this balance are caused by either on-bank intrusions of warmer, more saline Shelf/Slope Frontal Water (SSFW) or cooler, fresher water from presumably along-isobath and/or on-bank. The SSFW intrusions, while causing large positive fluctuations in water column mean temperature, occur intermittently (absent in some years) and do not appear to have a significant lasting effect on the local heat budget, which is consistent with the shorter studies of Churchill et al. (2003) and Brink and Lee (2009). Second, from about October to early April when the water column tends to become well mixed vertically and is cooled to its seasonal minimum, water column cooling is due to both surface cooling and episodic advection, with the latter clearly dominant on time scales from a few days to several weeks. Third, the seasonal decrease in vertical stratification starts in September while there is still net water column warming and surface heating, indicating that advection plays a significant role in de-stratification. This process continues well into October until the onset of strong storms begins to contribute significant surface cooling and deep mixing which, combined with advection, leads to persistent vertical well-mixed conditions at SF. (Note that this de-stratification process differs from the more
classical one-dimensional process observed on the New England mid-shelf by Lentz et al., 2003b).

These results suggest the following seasonal paradigm: the heat budget at SF is primarily one-dimensional on monthly and longer time scales during the warming season (ignoring the SSFW intrusions) but three-dimensional during the cooling season. Since the along-bank flow at SF is always directed towards the southwest on monthly time scales and does not exhibit large seasonal change, the effective upstream horizontal temperature gradient must be small so that \( \overline{Q}_{sf} \ll \overline{Q}_{adv} \) during the warming season but larger during the cooling season. A rough estimate of the winter \( \overline{Q}_{adv} \) can be made using a mean along-bank current (towards the southwest) of 5 cm/s and the along-bank gradient in monthly-mean SST* over a 50-km along-isobath transect northeast of SF for November through February when SST is a close proxy for depth-averaged temperature. The mean gradient averaged over the four winters (1995-96 through 1998-99) is 
\[ -0.006 \pm 0.004 \, ^\circ C/km. \]
The resulting advective cooling flux is -90 W/m², large enough to be significant on monthly time scales.

There are three potential sources of colder water that can contribute to the advective flux at SF during the cooling season. During this period, water on the shallow crest of the Bank is generally cooler than on the flanks, so off-bank transport of that water past SF can cause cooling (Lentz et al., 2003a). Water from the northern flank that continues around the northeast peak to the southern flank to SF could be cooled during the transit, presumably by surface cooling and lateral mixing with crest and outer flank water. Early in the cooling season, water from the northern flank could include Maine Intermediate Water, which is cooler and more saline at depth and known to exit the western Gulf of Maine along the northern edge of the Bank (Hopkins and Garfield, 1979). Later in the cooling season, the wind- and convection-driven surface mixed layer in the adjacent Gulf can reach depths up to 150 m (Brown and Beardsley, 1978; Mupparapu and Brown, 2002), creating water that is both cold and more saline and that can be carried onto the Bank upstream of the southern flank. The third potential source is Scotian Shelf Water (SSW), which is colder and fresher than Georges Bank water during winter and can occasionally cross the Northeast Channel to the eastern perimeter of Georges Bank (Bisagni et al., 1996). As described by Smith et al. (2003), these SSW “cross-over” events are generally episodic (on time scales of days to a few weeks), surface intensified in the roughly upper 35 m with very cold initial surface temperatures (below -1.0 °C) so that they can be tracked by AVHRR, and exhibit large interannual variation in occurrence. Brunner and Bisagni (2008) used satellite-derived surface temperature data from 1985-2000 to show that SSW cross-overs tend to depart from the southern-most tip of Browns Bank and arrive on the outer flank of Georges Bank east of about 66.3°W, where they are presumably absorbed into the Bank and Bank-edge flow and lose their extremely cold distinction before they reach SF. Thus, while they can not be tracked very far southwest using surface temperature, Lentz et al (2003a) used water column temperature and salinity data to track cross-over SSW from the Bank’s northeast flank, along the southern flank, and to SF and ST1 during March - early May 1995, where it contributed to advective cooling. It is likely that all three sources play a role in cooling the southern flank (and by extension water entering the Mid-Atlantic Bight), however the relative role of each source is unclear. The large interannual variability in SSW cross-over occurrences suggests that this is not the dominant source of cooling (Brunner and Bisagni, 2008).
Clearly more work is needed to investigate the heat budget on the southern flank of Georges Bank and place this information within the context of the broader Gulf of Maine system. Both the kinematics and dynamics of the flows that control the seasonal temperature variability on seasonal and longer time scales are quite unknown, yet this information has implications beyond simply physical oceanography. For example, if SSW is an important source of colder water on the southern flank of Georges Bank during winter, then it may also be an important source of new nutrients and biota during this period.

* The 1995-1999 AVHRR monthly mean SST composites were kindly provided by A. Thomas and R. Weatherbee of the Satellite Oceanography Data Lab, University of Maine.

5. Conclusions and Discussion

Currents over the Bank are, of course, dominated by tidal variability, especially at the semidiurnal $M_2$ frequency, but also at its overtones and at the diurnal $K_1$ frequency. At somewhat longer periods, wind driving plays a critical role which is most obvious during the wintertime when wind-current correlations are high. The ocean’s response is particularly pronounced near a 3-4 day period, and this enhancement seems to be associated with a wind-driven topographic wave that propagates and amplifies eastward along the northern side of the Bank. During the summer, wind-current correlations are considerably lower and there is evidence, based on sea level information, that equatorward coastal-trapped wave propagation plays a role in spreading information alongshore. On these time scales, the fluctuating flow is nearly unidirectional, with a top-to-bottom veering of typically about 20°. At the longer time scales resolved by monthly current averages, wind forcing is still evident, as is a summertime increase in vertical shear. This increase in summertime shear is consistent with model results (Chen and Beardsley, 1995; Chen et al., 1995) that show that the thermal wind associated with the tidal mixing front accelerates surface flow. This is also consistent with near-surface drifter data (Brink et al., 2003) that show upper ocean flow accelerating during the summertime. Interestingly, although subinertial frequency current variations at both of our primary mooring locations (SF and NEP) are coherent with the wind, they are uncorrelated with each other, during the summer of their relatively short common period, evidently because the NEP observations have more energy at lower frequencies. Finally, the overall record mean velocities are consistent at all depths with a clockwise gyre around the Bank.

Wind driving, either local or remote, is thus important at subinertial frequencies over the Bank. This forcing, though, is less efficient (measured as current response per unit wind stress) than in other coastal regimes having a nearby coastal wall. This locally weaker response need not mean that the overall wind response is weaker than these other areas, but could simply mean that the wind-driven energy is distributed over a wider area, i.e., the Gulf of Maine-Georges Bank complex, which is about 300 km wide.

Pringle (2006) points out forcefully that advective processes are dominated by the lowest frequencies in the currents. In this sense, monthly and longer time scales are likely to be most
important in governing changes in water properties and biological materials over the Bank. It seems intuitive that longer period motions could have a substantial component at larger spatial scales, and could entail changes in the alongshore advection just beyond the shelfbreak, as shown by Loder et al. (2001). With this reasoning in mind, it appears likely that year-to-year changes over the Bank, such as in salinity (Fig. 24), are a response to changes at far larger, basin scales.

The five-year time series of temperature and salinity (e.g. Fig. 24), at first glance, show a dominance by the longest time scales interrupted by very energetic warm core ring events. The temperature changes are completely dominated by the annual cycle, where the spring-summer warming trend is accounted for remarkably well by a one-dimensional, local heating calculation. This does not mean that the water was stagnant or that heat was not transported during this time, but it does mean that the water being warmed stayed within a very homogeneous environment, presumably that between the tidal mixing and shelfbreak fronts and surrounding Georges Bank, during the warming season. Salinity, on the other hand, shows no obvious annual cycle on the southern flank while the crest of the Bank exhibits a clear but small cycle of 32.4 in late March to 33.0 in late September (Mountain, personal correspondence). The explanation for this is that while the Gulf of Maine rivers contribute ~ 19% of the freshwater in the Gulf, only a small fraction of that water (<1.5% by volume) is found on the upper 30 m on the Bank crest, and instead, the primary source of freshwater for the Gulf and that found on the Bank is that which flows southwest along the Scotian Shelf and into the northern Gulf of Maine around Cape Sable (Houghton and Fairbanks, 2003; Loder et al, 2001). The two primary sources for the western Scotian Shelf freshwater are the St. Lawrence estuary and Labrador Shelf water, which drove the 1995-1999 interannual variability observed in salinity on the southern flank of the Bank (Khatiwala et al, 1999; Houghton, personal correspondence)

At time scales shorter than the seasonal, Georges Bank waters are affected by a range of event types. These include crossovers from the Scotian Shelf (predominantly in winter), warmer, fresh intrusions from the coastal Gulf of Maine (predominantly in summer), smaller-scale slope water salty intrusions and Warm Core Ring invasions. These are all unambiguous signs that waters from away from the Bank reach its confines, but the implications of these events are less clear. Moorings are not well suited for tracking the life of a water patch, especially one with the small spatial scales, O(10 km), that seem to characterize these intrusions. Because of this difficulty in tracking, it is hard to say how important a role they play in the Bank’s budgets of salinity, nutrients or plankton. Assumptions have to be made, such as that of Brink and Lee (2009), who assumed that all of a Warm Core Ring intrusion’s water remained on the Bank, and even so, they found that it would not make a major perturbation to the Bank’s temperature or salinity budgets. Further, since water masses have different “tags” in each of these properties, it is quite conceivable that a given event type could play a major role in, say, the nutrient budget, but no significant role in the salinity budget. A much more Lagrangian and station-oriented approach, such as that applied by Churchill et al. (2003), must be employed.

Two particular aspects of our findings have broader relevance for the interdisciplinary GLOBEC program. First, Georges Bank is far from being a closed system. The around-Bank circulation (as seen especially in drifter data: Brink et al., 2003) draws water from the western Gulf of Maine, passes it around the Bank and then forwards it into the Mid-Atlantic Bight. Further, intrusions from various directions (section 4.2) are present for about 25-30% of the time at our moorings.
These intrusions are expected to have substantial impact on biological activity, for example, by introducing springtime static stability (e.g., Ji et al., 2006) and so prompting phytoplankton blooms. Some intrusions, such as those associated with Warm Core Rings, have even more dramatic impacts (at least temporarily) on stability and water masses (e.g., Ryan et al., 2001), and so it is hard to imagine that they do not play a substantial biological role, but at this time, that role is fairly conjectural. Second, our results show substantial year-to-year changes in stratification (Fig. 24) and monthly-averaged along-bank flow (Fig. 14). Southwestward flow along the southern flank is always present, so that the circulation so central to the Lough (1984) larval fish paradigm is always present, but the strength of that flow in the early months of the year varies by at least a factor of two from year to year. Lewis et al. (2001) point out that it is these flow time scales that are important for advecting plankton patches, and so one should expect substantial interannual variations in the location of, for example, spring bloom phytoplankton patches. A patch’s location could very well have substantial implications for zooplankton populations and, in turn, higher trophic levels. Building a case for the biological impact of specific physical events takes a dedicated interdisciplinary effort, but, to date, these efforts have generally taken the form of numerical modeling studies (e.g., Ji et al., 2006; Hu, 2009) which have the advantages of providing complete flow fields and allowing experimental manipulation.

In the end, we have shown that Georges Bank is a dynamic, leaky environment. Currents are clearly affected by tides and winds, while hydrographic properties are strongly affected by surface heating and by Scotian Shelf, Gulf of Maine and offshore conditions. Intrusion events occur frequently, but their broader impact, either physically or biologically, has yet to be sharply defined.

**Acknowledgements**

This work took place as part of the GLOBEC Northwest Atlantic/Georges Bank field project, and was sponsored through NSF Biological Oceanography grants OCE-80644500 and OCE-80644501. S. Lentz, J. Bisagni, D. Mountain, and C. Flagg provided helpful comments and insights to this work. This is U.S. GLOBEC contribution number xxxx (number given after manuscript is accepted).
References


Tables

Table 1
Mooring location information.

<table>
<thead>
<tr>
<th>Name</th>
<th>Latitude (N)</th>
<th>Longitude (W)</th>
<th>Nom. Depth (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Crest (CR)</td>
<td>41° 24.5´</td>
<td>67° 32.5´</td>
<td>43</td>
</tr>
<tr>
<td>Northeast Peak (NEP)</td>
<td>41° 44.0´</td>
<td>66° 32.0´</td>
<td>73</td>
</tr>
<tr>
<td>South Flank (SF)</td>
<td>40° 58.0´</td>
<td>67° 19.0´</td>
<td>76</td>
</tr>
<tr>
<td>NDBC Buoy 44016</td>
<td>41° 06.6´</td>
<td>66° 34.8´</td>
<td>88</td>
</tr>
</tbody>
</table>

Table 2
Tidal constants for SF bottom pressure. Following Moody et al. (1984), pressure is converted to cm using 1 dbar = 99.5 cm.

<table>
<thead>
<tr>
<th>Constituent</th>
<th>Period (hrs)</th>
<th>Amplitude (cm)</th>
<th>Greenwich Phase (°)</th>
</tr>
</thead>
<tbody>
<tr>
<td>O₁</td>
<td>25.81934</td>
<td>6.05±0.52</td>
<td>179.5±2.2</td>
</tr>
<tr>
<td>K₁</td>
<td>23.93447</td>
<td>7.06±0.52</td>
<td>173.3±2.0</td>
</tr>
<tr>
<td>N₂</td>
<td>12.65835</td>
<td>9.46±0.27</td>
<td>342.3±1.4</td>
</tr>
<tr>
<td>M₂</td>
<td>12.42060</td>
<td>38.64±0.27</td>
<td>6.1±0.4</td>
</tr>
<tr>
<td>S₂</td>
<td>12.00000</td>
<td>8.25±0.52</td>
<td>29.0±1.7</td>
</tr>
<tr>
<td>M₄</td>
<td>6.21030</td>
<td>1.1±0.11</td>
<td>199.6±4.8</td>
</tr>
<tr>
<td>M₆</td>
<td>4.14020</td>
<td>1.0±0.10</td>
<td>354.2±8.0</td>
</tr>
</tbody>
</table>
Table 3

Major Tidal Constituents for Currents at 20 m depth. Negative minor axis amplitude denotes clockwise rotation of the velocity vector. Greenwich Phase is the lag of the observed maximum velocity behind the passage of the sun or the moon over the Greenwich Meridian. Inclination is the angle measured clockwise from North. All errors bounds represent 95% confidence limits.

<table>
<thead>
<tr>
<th>Southern Flank</th>
<th>Major (cm/sec)</th>
<th>Minor (cm/sec)</th>
<th>Greenwich Phase (°)</th>
<th>Inclination (°)</th>
<th>% Total Record Variance</th>
</tr>
</thead>
<tbody>
<tr>
<td>O₁</td>
<td>2.00±0.58</td>
<td>-1.46±0.60</td>
<td>103.0±17</td>
<td>57.1±18</td>
<td>0.18</td>
</tr>
<tr>
<td>K₁</td>
<td>5.24±0.58</td>
<td>-4.11±0.58</td>
<td>137.1±8.6</td>
<td>62.9±8.8</td>
<td>1.31</td>
</tr>
<tr>
<td>N₂</td>
<td>8.96±0.48</td>
<td>-6.52±0.46</td>
<td>329.2±10</td>
<td>325.8±11</td>
<td>3.63</td>
</tr>
<tr>
<td>M₂</td>
<td>41.29±0.48</td>
<td>-28.86±0.46</td>
<td>5.2±1.6</td>
<td>328.9±1.6</td>
<td>75.01</td>
</tr>
<tr>
<td>S₂</td>
<td>6.09±0.46</td>
<td>-4.17±0.41</td>
<td>40.8±14</td>
<td>332.3±15</td>
<td>1.61</td>
</tr>
<tr>
<td>M₄</td>
<td>0.72±0.18</td>
<td>-0.31±0.19</td>
<td>151.8±28</td>
<td>317.8±27</td>
<td>0.02</td>
</tr>
<tr>
<td>M₆</td>
<td>0.52±0.10</td>
<td>-0.12±0.10</td>
<td>276.0±18</td>
<td>328.2±15</td>
<td>0.01</td>
</tr>
</tbody>
</table>

Total Record Variance (TRV) 1690 (cm/s)²

These 7 tidal constituents contribute to the TRV 81.9% var

All tidal constituents contribute to the TRV 89.1% var

<table>
<thead>
<tr>
<th>Northeast Peak</th>
<th>Major (cm/sec)</th>
<th>Minor (cm/sec)</th>
<th>Greenwich Phase (°)</th>
<th>Inclination (°)</th>
<th>% Total Record Variance</th>
</tr>
</thead>
<tbody>
<tr>
<td>O₁</td>
<td>1.88±0.32</td>
<td>0.43±0.32</td>
<td>45.2±22</td>
<td>29.0±22</td>
<td>0.04</td>
</tr>
<tr>
<td>K₁</td>
<td>6.62±0.29</td>
<td>-3.73±0.32</td>
<td>80.7±7.2</td>
<td>21.0±6.8</td>
<td>0.60</td>
</tr>
<tr>
<td>N₂</td>
<td>16.06±0.48</td>
<td>10.82±0.44</td>
<td>343.2±8.0</td>
<td>333.4±8.0</td>
<td>3.92</td>
</tr>
<tr>
<td>M₂</td>
<td>76.96±0.52</td>
<td>-50.98±0.48</td>
<td>14.2±1.6</td>
<td>329.0±1.6</td>
<td>88.93</td>
</tr>
<tr>
<td>S₂</td>
<td>11.72±0.52</td>
<td>-7.47±0.44</td>
<td>44.67±10</td>
<td>331.1±8.8</td>
<td>2.02</td>
</tr>
<tr>
<td>M₄</td>
<td>1.62±0.28</td>
<td>-0.56±0.28</td>
<td>146.2±23</td>
<td>348.9±22</td>
<td>0.03</td>
</tr>
<tr>
<td>M₆</td>
<td>0.97±0.08</td>
<td>-0.52±0.08</td>
<td>73.0±13</td>
<td>25.7±13</td>
<td>0.01</td>
</tr>
</tbody>
</table>

Total Record Variance (both components) 4790 (cm/s)²

These 7 tidal constituents contribute to the TRV 95.6% var

All tidal constituents contribute to the TRV 97.1% var
Table 4
Summary of intrusion events.

<table>
<thead>
<tr>
<th>Site</th>
<th>Water Type</th>
<th>Number of Events</th>
<th>Total Days</th>
<th>Days of Observations</th>
<th>% Occurrence</th>
</tr>
</thead>
<tbody>
<tr>
<td><em>NE Peak</em></td>
<td>Scotian Shelf</td>
<td>6-7</td>
<td>29-72</td>
<td>527</td>
<td>6-14</td>
</tr>
<tr>
<td></td>
<td>Western Gulf</td>
<td>3-4</td>
<td>35-78</td>
<td>470</td>
<td>7-17</td>
</tr>
<tr>
<td></td>
<td>Slope Water</td>
<td>4</td>
<td>13</td>
<td>535</td>
<td>2</td>
</tr>
<tr>
<td><em>South Flank</em></td>
<td>Scotian Shelf</td>
<td>5</td>
<td>39</td>
<td>1351</td>
<td>3</td>
</tr>
<tr>
<td></td>
<td>Western Gulf</td>
<td>6</td>
<td>125</td>
<td>1351</td>
<td>9</td>
</tr>
<tr>
<td></td>
<td>Slope Water &amp; WCRs</td>
<td>8</td>
<td>135</td>
<td>1376</td>
<td>10</td>
</tr>
<tr>
<td></td>
<td>Slope Water (deep)</td>
<td>3</td>
<td>78</td>
<td>943</td>
<td>5</td>
</tr>
</tbody>
</table>

Table 5
A. Temperature maximum and time of maximum taken from the SF seasonal cycle shown in Figure 27. STD is the standard deviation between the observed temperature and Fourier fit.

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>Tmax (°C)</th>
<th>Time</th>
<th>STD (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>5</td>
<td>17.20</td>
<td>8/23</td>
<td>1.46</td>
</tr>
<tr>
<td>10</td>
<td>16.63</td>
<td>8/28</td>
<td>1.61</td>
</tr>
<tr>
<td>20</td>
<td>14.87</td>
<td>9/12</td>
<td>1.62</td>
</tr>
<tr>
<td>30</td>
<td>13.88</td>
<td>9/25</td>
<td>1.40</td>
</tr>
<tr>
<td>40</td>
<td>13.14</td>
<td>10/3</td>
<td>1.12</td>
</tr>
<tr>
<td>50</td>
<td>12.82</td>
<td>10/3</td>
<td>1.13</td>
</tr>
<tr>
<td>72</td>
<td>11.33</td>
<td>10/3</td>
<td>0.99</td>
</tr>
</tbody>
</table>

B. Density minimum and time of minimum taken from the SF seasonal cycle (Figure 27).

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>σt min</th>
<th>Time</th>
<th>STD</th>
</tr>
</thead>
<tbody>
<tr>
<td>5</td>
<td>23.81</td>
<td>8/23</td>
<td>0.24</td>
</tr>
<tr>
<td>10</td>
<td>23.72</td>
<td>8/28</td>
<td>0.29</td>
</tr>
<tr>
<td>20</td>
<td>24.20</td>
<td>9/12</td>
<td>0.22</td>
</tr>
<tr>
<td>30</td>
<td>24.48</td>
<td>9/25</td>
<td>0.20</td>
</tr>
<tr>
<td>40</td>
<td>24.66</td>
<td>10/3</td>
<td>0.20</td>
</tr>
<tr>
<td>50</td>
<td>24.79</td>
<td>10/3</td>
<td>0.22</td>
</tr>
</tbody>
</table>
Table 6
Results of SF heat budget analysis. Start/end times, record length, mean water column, surface and advective heat fluxes, and comments are presented for each leg. The longer legs are subdivided into shorter segments to provide more detail into seasonal variations.

<table>
<thead>
<tr>
<th>Leg #</th>
<th>Start/Stop</th>
<th>Length</th>
<th>$Q_{wc}$</th>
<th>$Q_{sf}$</th>
<th>$Q_{adv}$</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 (1994-1995)</td>
<td>10/28-03/03</td>
<td>126</td>
<td>-275</td>
<td>-110</td>
<td>-165</td>
<td>Strong advective and surface cooling; record mean</td>
</tr>
<tr>
<td>2 (1995)</td>
<td>04/27-08/03</td>
<td>98</td>
<td>175</td>
<td>220</td>
<td>-45</td>
<td>~ 1-D</td>
</tr>
<tr>
<td></td>
<td>08/03-10/02</td>
<td>60</td>
<td>390</td>
<td>70</td>
<td>320</td>
<td>SSFW intrusion warming, with Tmax &gt; 20 °C and Smax&gt;35</td>
</tr>
<tr>
<td></td>
<td>04/27-10/02</td>
<td>158</td>
<td>255</td>
<td>165</td>
<td>90</td>
<td>Record Mean</td>
</tr>
<tr>
<td>3 (1995-1996)</td>
<td>11/01-01/12</td>
<td>72</td>
<td>-355</td>
<td>-170</td>
<td>-185</td>
<td>Strong surface and advective cooling</td>
</tr>
<tr>
<td></td>
<td>01/12-03/05</td>
<td>53</td>
<td>-160</td>
<td>-60</td>
<td>-120</td>
<td>Strong advective cooling with some surface cooling</td>
</tr>
<tr>
<td></td>
<td>11/01-03/05</td>
<td>125</td>
<td>-275</td>
<td>-125</td>
<td>-150</td>
<td>Record mean</td>
</tr>
<tr>
<td>4 (1996)</td>
<td>04/07-06/11</td>
<td>65</td>
<td>170</td>
<td>200</td>
<td>-30</td>
<td>~ 1-D</td>
</tr>
<tr>
<td></td>
<td>06/11-09/06</td>
<td>87</td>
<td>110</td>
<td>210</td>
<td>-100</td>
<td>Strong surface warming with some advective cooling</td>
</tr>
<tr>
<td></td>
<td>04/07-09/06</td>
<td>152</td>
<td>135</td>
<td>205</td>
<td>-75</td>
<td>Record mean</td>
</tr>
<tr>
<td>5 (1996)</td>
<td>10/27-12/18</td>
<td>52</td>
<td>-225</td>
<td>-60</td>
<td>-165</td>
<td>Strong advective cooling with some surface cooling</td>
</tr>
<tr>
<td>6 (1997)</td>
<td>02/15-03/23</td>
<td>39</td>
<td>-55</td>
<td>0</td>
<td>-55</td>
<td>Advective cooling</td>
</tr>
<tr>
<td></td>
<td>03/23-05/07</td>
<td>39</td>
<td>90</td>
<td>135</td>
<td>-45</td>
<td>Strong surface warming with some advective cooling</td>
</tr>
<tr>
<td></td>
<td>02/15-05/07</td>
<td>81</td>
<td>20</td>
<td>70</td>
<td>-50</td>
<td>Record mean</td>
</tr>
<tr>
<td>7 (1997-1998)</td>
<td>06/22-10/20</td>
<td>120</td>
<td>135</td>
<td>135</td>
<td>0</td>
<td>~ 1-D (with major SSFW intrusion 7/2-12, with Tmax ~ 24 °C and Smax ~35.8; see Brink and Lee, 2009)</td>
</tr>
<tr>
<td></td>
<td>10/20-02/18</td>
<td>121</td>
<td>-265</td>
<td>-85</td>
<td>-175</td>
<td>Strong advective cooling with some surface cooling.</td>
</tr>
<tr>
<td></td>
<td>02/18-04/17</td>
<td>58</td>
<td>20</td>
<td>90</td>
<td>-70</td>
<td>Surface warming ~ advective cooling</td>
</tr>
<tr>
<td></td>
<td>06/22-04/17</td>
<td>299</td>
<td>-50</td>
<td>35</td>
<td>-85</td>
<td>Record mean</td>
</tr>
<tr>
<td>8 (1998)</td>
<td>05/07-06/10</td>
<td>34</td>
<td>170</td>
<td>230</td>
<td>-60</td>
<td>Strong surface warming with some advective cooling</td>
</tr>
<tr>
<td>9 (1998-1999)</td>
<td>07/03-10/10</td>
<td>99</td>
<td>150</td>
<td>145</td>
<td>5</td>
<td>~ 1-D</td>
</tr>
<tr>
<td></td>
<td>10/10-02/18</td>
<td>131</td>
<td>-210</td>
<td>-95</td>
<td>-115</td>
<td>Strong advective and surface cooling</td>
</tr>
<tr>
<td></td>
<td>02/18-04/01</td>
<td>42</td>
<td>-10</td>
<td>-50</td>
<td>40</td>
<td>Surface cooling ~ advective warming</td>
</tr>
<tr>
<td></td>
<td>07/03-04/01</td>
<td>272</td>
<td>-50</td>
<td>0</td>
<td>-50</td>
<td>Record mean</td>
</tr>
<tr>
<td>10 (1999)</td>
<td>05/14-08/16</td>
<td>94</td>
<td>230</td>
<td>210</td>
<td>20</td>
<td>~ 1-D; record mean</td>
</tr>
</tbody>
</table>