Temperature, salinity, and density variability in the central Middle
Atlantic Bight

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[1] Four years of sustained glider observations are used to compute the seasonal cycle of hydrographic fields in the central Middle Atlantic Bight (MAB). Results reveal a large phase lag in near bottom temperatures, with peak values occurring in September at the inner shelf, in October at the mid shelf, and in November at the outer shelf. Unlike the northern MAB, the seasonal cycle explains over 70% of the near-surface salinity variability. At the inner shelf and offshore near the bottom, however, most of the variance is due to pulses in river discharge and to shifts in the position of the shelfbreak front. Cross-shelf density gradients inshore of the 60-m isobath are dominated by salinity during winter and spring, with temperature contributing significantly from August to October. This is because bottom waters near the coast are warm due to the deepening of the thermocline during fall, but offshore waters are still influenced by the cold pool. The vertical stratification seasonal variability is also large. Early in the year, stratification is small and entirely due to salinity. By May, salinity still dominates vertical gradients near the coast, but temperature and salinity contribute equally to the density stratification offshore. During summer, stratification is dominated by temperature. Temperature interannual variability was small during the sampling period, but surface salinity was anomalously low by 1.2 psu in summer 2006. The anomaly was due to larger than average discharge from the Hudson River in early summer during a period of strong upwelling favorable winds.


1. Introduction

[2] The hydrography and stratification in the Middle Atlantic Bight (MAB) are characterized by large seasonal variations [e.g., Bigelow, 1933; Houghton et al., 1982; Beardsley et al., 1985; Linder and Gawarkiewicz, 1998; Lentz et al., 2003; Shearman and Lentz, 2003]. Waters are strongly thermally stratified during summer, with a sharp thermocline at about 20 m depth. A band of cold water usually referred to as the ‘cold pool’ [Houghton et al., 1982] is found below the thermocline at the middle and outer shelf. This vertical structure of the thermal field is eroded during fall primarily due to the increase in the frequency of storms, especially when winds are downwelling-favorable [Lentz et al., 2003]. As a result, shelf waters are vertically homogeneous in temperature during winter, although cross-shelf gradients are still present. As the frequency of storms decreases and the surface heat fluxes increase in spring, thermal stratification rebuilds. Shelf waters are separated from slope waters by a shelfbreak front [e.g., Linder and Gawarkiewicz, 1998].

[3] The salinity field is characterized by low values near the coast increasing offshore toward the shelfbreak front and with depth. In contrast to temperature, the reported seasonal variations in salinity and in cross-shelf salinity gradients are much smaller [Manning, 1991; Linder and Gawarkiewicz, 1998; Lentz et al., 2003; Mountain, 2003; Shearman and Lentz, 2003; Lentz, 2008b]. The notable exception is the inner shelf, where gradients are often enhanced due to increased river discharge during spring [e.g., Ullman and Codiga, 2004; Codiga, 2005; Chant et al., 2008a, 2008b]. The large seasonal variation in stratification is, therefore, primarily associated with the seasonal variation in water temperature.

[4] It is important to note, however, that this description is based mostly on observations from the northern MAB, where the great majority of previous studies in the MAB took place. Lentz [2003] pointed out that of a total of 10,652 acceptable hydrographic profiles extracted from the National Oceanographic Data Center (NODC) archive for the continental shelf between Cape Hatteras and the northeastern tip of Georges Bank, only ~16% were collected to the south of the Hudson Shelf Valley (Figure 1). Since these include also observations from the southern MAB (between
Chesapeake Bay and Cape Hatteras), it is reasonable to assume that less than 10% of the available observations in the NODC archives for the MAB were collected in the central part of the Bight (between Hudson Shelf Valley and Chesapeake Bay). The central MAB is, therefore, a relatively under sampled region, at least concerning the readily available observations from the NODC archives. There are, however, important differences between the central and the northern MAB, including the coastline orientation, which enhances the coastal upwelling response to summertime southerly winds in the central MAB, and the proximity of the Hudson River outflow, which delivers large amounts of freshwater to the central MAB with no equivalent in the northern MAB. Mountain [2003] used bulk shelf water properties to find a significant salinity seasonal cycle in the central MAB, but not in the New England shelf. Observations indicate that interannual variability in salinity on the New England shelf over shadows the seasonal variability. Characterizing the differences in the seasonal evolution of the fields in the central MAB compared to the more studied northern MAB is important, because both temperature and salinity contribute to the structure of the density field which plays a critical role in the dynamics of shelf circulation [e.g., Chapman and Lentz, 1994; Shearman and Lentz, 2003; Lentz, 2008a].

Most of the previous studies on the MAB were based on mooring observations (which are great to address the longer-term statistics but provide limited information on spatial scales), on intense hydrographic surveys (which often resolve the spatial structure but are subject to the question of how representative such short-term observations are [Linder and Gawarkiewicz, 1998]), or on climatologies (which are great to describe mean patterns but are also characterized by some degree of smoothing). Here, we use four years of quasi-synoptic, very high-resolution glider observations to investigate seasonal and interannual variability of temperature, salinity and density in the central MAB. The glider measurements are comprised of a total of 80552 hydrographic profiles, greatly increasing the number of observations in this relatively under sampled region.

2. Field Observations and Processing

Repeated surveys of water over the shelf off New Jersey using autonomous underwater gliders begun in October 2003 as part of the ongoing Rutgers University Glider Endurance Line [Castelao et al., 2008a] (Figure 1). The surveys are composed of cross-shelf sections up to 130 km long, generally extending from the 20 to the 100 m isobath. In some instances, however, the sections were limited to the region close to the coast, or to the region between the 60 and the 100 m isobaths. Since sections do not extend offshore of the 100 m isobath, the shelfbreak front is generally not fully resolved. Hydrographic data are collected

Figure 1. Study area showing glider tracks (black lines). Also shown are location of NOAA NDBC buoys 44025 (circle), 44009 (diamond), NOS station SDHN4 (triangle) and CMAN station at Ambrose Light (square).
using a fleet of Teledyne Webb Research Corporation Slocum Coastal Electric Gliders [Schofield et al., 2007], which are equipped with a Sea-Bird conductivity-temperature-depth (CTD) instrument. Some of the gliders also have optical sensors to measure chlorophyll fluorescence, chroomorphic dissolved organic matter and other biologically relevant properties, although only temperature, salinity and density observations are reported here. We focus on observations from October 2003 to October 2008, but discard observations from 2005. This is because the temporal coverage during that year was very limited. A total of 103 transects are available during the study period. The average number of sections for each month plus or minus one standard error of the mean is 8.58 ± 1.51, with a median of 8. In general, more observations are available during spring and summer than during fall or winter. Note also that because the length of the cross-sections varies, the number of observations used on the computations varies spatially, which can lead to small discontinuities in some of the averaged fields.

The gliders are operational in water depths up to 100 m, cycling from the surface to 3–5 m above the bottom while moving forward at an average speed of 25–30 km per day. Typical along-track resolution is about 300–400 m near the shelf break, improving to about 100–200 m over the shallower regions close to the coast. Each transect takes approximately 4–5 days to be completed. Observations can adequately resolve seasonal variability, but high frequency variations (e.g., tides, internal waves, inertial motions) are not fully resolved. Measured and derived quantities (salinity, temperature, density; computed following Fofonoff and Millard [1983]) from each glider transect were projected into a common line and then averaged vertically to 1 dbar bins and horizontally to 500 m.

The seasonal cycles of temperature and salinity along the transect are estimated by fitting observations to a mean and annual harmonic:

\[ y(t) = \bar{y} + a \sin \omega_1 t + b \cos \omega_1 t, \]

where \( y(t) \) is the variable time series, \( \bar{y} \) is the temporal average of the variable, \( t \) is the time in days, \( \omega_1 = 2\pi/(365.25 \text{ days}) \) is the annual frequency and \( a \) and \( b \) are the coefficients of the harmonic fit. The spatial pattern of the percent contribution of the seasonal cycle and the residual time series to the total variance is computed by assuming that total variance is approximated as the sum of the variance contributed by the seasonal cycle and the residue [Leggoard and Thomas, 2006]. The error due to the assumption is less than 15% for temperature, and less than 8% for salinity. In order to better represent asymmetries in the temperature and salinity seasonal evolution, glider observations were also fitted to a mean and to 3 harmonics (with \( \omega_2 = 2\omega_1 \) and \( \omega_3 = 3\omega_1 \)).

Cross-shelf gradients of temperature, salinity and density were estimated via linear regression for the regions extending (a) from the coast to the 60-m isobath and (b) offshore of the 60-m isobath. The 60-m isobath (∼90 km from the coast) was chosen as the limit between ‘inshore’ and ‘offshore’ because that is the approximate region where the cross-shelf temperature gradient changes sign during the stratified season in the northern MAB [Shearman and Lentz, 2003; Lentz, 2008b].

Freshwater discharge in the region is dominated by the Hudson River outflow, and was obtained from the U.S. Geological Survey (USGS) Water Data website (available online at http://waterdata.usgs.gov) and subsequently modified to include ungauged portions of the watershed following Chant et al. [2008a]. Wind observations were obtained at the NOAA National Data Buoy Center (NDBC) buoy 44025 located at 40.25°N, 73.17°W (see Figure 1 for location). Neutral wind stress was calculated following Large and Pond [1981] and then low-pass filtered (half-power point of 40 h) to remove short-period fluctuations. Gaps in the NOAA NDBC buoy 44025 record were filled through regression with the nearby NOAA CMAN station at Ambrose Light and NOAA NDBC buoy 44009 located near the Delaware Bay mouth (38.24°N, 74.70°W). Wind observations at buoy 44025 during 2007 presented several gaps, so a substantial fraction of the observations shown here for that year are from buoy 44009 adjusted by linear regression. Using 2007 observations from NOAA NOS station SDHN4 at Sandy Hook, NJ in the regression analyses produces similar results.

3. Temperature, Salinity, and Density Variability

3.1. Thermohaline and Density Fields

Monthly averages of temperature, salinity and density based on the four years of observations are presented in Figures 2–4, respectively. The standard deviations of the 3 variables, grouped by season, are shown on Figure 5. Several of the main characteristics of the seasonal evolution of hydrographic fields in the MAB have already been described in the literature [e.g., Houghton et al., 1982; Beardsley et al., 1985; Mountain, 2003; Lentz, 2008b]. However, the great majority of previous observations in the MAB have been collected in the northern MAB, to the north of the Hudson Shelf Valley [Lentz, 2003]. As will be shown in the next sections, the seasonal evolution of the temperature field in the central MAB is similar to the evolution observed in the northern MAB. The evolution of salinity, on the other hand, is considerably different. We focus here on the differences encountered on the central MAB.

Early in the year, temperatures are low close to the coast, increasing offshore toward the shelfbreak front (Figure 2). Vertical gradients are small, with temperatures increasing slightly with depth. The salinity field is characterized by a similar pattern (Figure 3), with salinities increasing offshore, with low-salinity waters near the coast reaching the bottom, and with small top-to-bottom differences throughout the shelf. Because inshore waters are fresh and cold, and offshore waters are warm and salty, the contributions to the density field tend to cancel each other, and density differences across the shelf are relatively small [e.g., Shearman and Lentz, 2003; Lentz, 2008b] (Figure 4). All these are consistent with previous observations elsewhere on the MAB [Beardsley et al., 1985; Linder and Gawarkiewicz, 1998; Shearman and Lentz, 2003; Linder et al., 2006]. Variations in temperature, salinity and density during that time of the year are mostly depth-independent and small, with no substantial variation across the shelf (Figure 5).

Warming at the surface layer begins somewhat between March and April and continues until August (Figure 2; see also Figure 7) following the seasonal evolu-
tion of surface heating. A shallow thermocline begins to develop sometime around May, reaching peak intensity in August. Below the thermocline, average temperatures in the cold pool [Bigelow, 1933; Houghton et al., 1982] remain approximately constant until late summer (note that the structure of the cold pool is not clearly evident in Figure 2 due to the color scale used). Near the coast, however, the seasonal heating in late summer extends all the way to the bottom. Most of the temperature variability during spring is confined to the top 10 m of the water column and is associated with the seasonal warming (Figure 5). During summer, however, the largest standard deviations of the temperature field occur at the thermocline level. Individual sections reveal that propagating internal waves are a major contributor to the enhanced variability at that level.

**Figure 2.** Monthly averaged temperature (°C) cross-section along endurance line.

**Figure 3.** Monthly averaged salinity cross-section along endurance line.
Figure 4. Monthly averaged density (kg m\(^{-3}\)) cross-section along endurance line.

Figure 5. (a–d) Temperature (°C), (e–h) salinity, and (i–l) density (kg m\(^{-3}\)) standard deviation along endurance line for winter (Figures 5a, 5e, and 5i), spring (Figures 5b, 5f, and 5j), summer (Figures 5c, 5g, and 5k), and fall (Figures 5d, 5h, and 5l).
[14] In contrast to temperature, which presents a seasonal evolution similar to the New England shelf in the northern MAB [e.g., Lentz, 2008b], the evolution of salinity in the central MAB is quite different than that observed in regions to the north. The salinity field during spring and summer is marked by a progressive decrease in the averaged value over the shelf (Figure 3). Shelf-averaged (cross- and depth-averaged) salinities are about 1 psu lower during summer compared to winter. The reduced salinity is at least partially due to increased runoff from the Hudson River, which historically peaks in April. In addition to that, the tendency for the shelfbreak front to migrate offshore during spring/summer [Linder and Gawarkiewicz, 1998] and alongshore advection (equatorward depth-averaged flow peaks during summer [Lentz, 2008b]) presumably contribute to the shelf-wide freshening. There is also a clear transition from a structure where the largest salinity variations are across the shelf to a scenario where a sharp halocline is formed, which is not observed in the northern MAB partially because there is no strong local river runoff there. This shift occurs as upwelling winds begin to blow persistently sometime around May (see Figure 13) transporting low-salinity waters offshore, either from a coastal current by surface Ekman transport [Fong et al., 1997; Hickey et al., 1998; Rennie et al., 1999; Fong and Geyer, 2001; Johnson et al., 2001; Sanders and Garvine, 2001; García Berdeau et al., 2002; Hallock and Marmiono, 2002; Johnson et al., 2003; Lentz, 2004; Choi and Wilkin, 2007; Castelao et al., 2008b] or directly from near the Hudson River mouth by a recently described transport pathway along the southern flank of the Hudson Shelf Valley [Castelao et al., 2008b; Zhang et al., 2009]. The salinity standard deviation during spring is quite large near the coast (Figure 5), a signature of freshwater plumes that reach the region in a buoyancy-driven coastal current. Values are also slightly enhanced in a thin layer near the surface. During summer, however, the pattern closely resembles the temperature standard deviation. The region of largest variability is coincident with the bottom of the surface mixed layer, in part due to internal wave activity along the halocline. Note also that there are intensifications in the salinity standard deviation at the offshore region both at the surface and near the bottom. Those local enhancements are due to surface salty intrusions similar to the one observed at Nantucket Shoals by Gawarkiewicz et al. [1996] and to shifts in the position of the foot of the shelfbreak front, respectively. A detailed description of salty intrusions for one of the years reported here (2006) will be presented elsewhere.

[15] Unlike winter, when the contributions of temperature and salinity to the density field tend to cancel each other, they both contribute to establishing a layer of very low-density waters near the surface during late spring and summer (Figure 4). The average density difference across the pycnocline during July/August is as large as 4 kg m$^{-3}$. The signature of the low-salinity water near the coast during spring is clearly evident in the density standard deviation plot (Figure 5), as is the region of enhanced variability at pycnocline level during summer.

[16] During fall, the frequency of strong wind events in the MAB tend to increase [e.g., Lentz et al., 2003; Gong et al., 2010]. The enhanced mixing leads to a thicker mixed layer and a weaker thermocline in September and October (Figure 2). Maximum temperature variability is still located at the thermocline level, which is deeper than during summer (Figure 5). September also marks the typical end of the upwelling season (see Figure 13). The increase in the frequency of downwelling favorable winds during that time together with the reduced discharge from the Hudson River during summer causes the retraction of the low-salinity waters toward the coast. This establishes a much saltier surface layer compared to summer conditions (Figure 3). The signature of pycnocline and near-surface salty intrusions [e.g., Gawarkiewicz et al., 1996; Lentz, 2003] is clearly seen in the September/November salinity fields. Similarly to summer conditions, the salinity standard deviation is intensified near the surface and near the bottom close to the shelfbreak (Figure 5). The intensifications are associated with salty intrusions and with changes in the position of the foot of the shelfbreak front.

[17] The thermal stratification breaks down in November, when temperatures become nearly constant across the shelf (Figure 2). This leads to a sharp increase in near-bottom temperatures offshore, so that highest temperatures are found during that time (see also Figure 7). Low-salinity waters are once again restricted to very near the coast (Figure 3). During that time, the density field (Figure 4) closely resembles the salinity distribution, since temperature is nearly homogeneous.

3.2. Seasonal Cycles

[18] The seasonal cycles of temperature and salinity were estimated using equation (1), and are shown in Figure 6. The amplitude of the temperature seasonal cycle decreases from 10°C at the surface to about 2°C near the bottom offshore. There is also a tendency for larger amplitudes near the coast compared to offshore, especially near the surface. This is because maximum surface temperatures during summer do not vary much across the shelf but offshore waters are warmer than near-coastal waters during winter due to the influence of the shelfbreak front, which is located closer to the shelf during that time of the year [Linder and Gawarkiewicz, 1998]. The seasonal cycle peaks nearly simultaneously in the surface layer throughout the shelf in August, consistent with the evolution of surface heating. Note, however, that using 3 harmonics in the least squares fit (with $\omega_2 = 2\omega_1$ and $\omega_3 = 3\omega_1$; Figure 7) reveals that the seasonal evolution of temperature at the surface is not symmetric, with the coldest water at the surface not occurring simultaneously across the shelf. The coldest surface waters are found in early March near the coast, in mid-March at the mid-shelf, and in early April offshore.

[19] Near the bottom, there is a substantial phase difference in the temperature annual cycle across the shelf (Figure 6). Maximum near-bottom temperatures in shallow regions are found in September, and are associated with the deepening of the thermocline from summer to fall (Figure 2). Offshore of the ~40 m isobath, however, the seasonal warming peaks later in the year, with the latest warming occurring near the 60-m isobath in mid-November. The delay in the seasonal cycle peak offshore occurs presumably because, during summer, warming due to solar radiation is partially offset by cooling due to the southward advection of cold water in the cold pool [Houghton et al.,
leading to only small changes in near-bottom temperatures (Figure 2). Bottom temperatures only increase later in the fall, when the increase in the frequency of strong wind events leads to the breakdown of the seasonal thermocline.

The 1-harmonic fit to the annual cycle explains over 80% of the temperature variance in the surface layer, but less than 40% of the variance below 20 m depth offshore of the 40–50 m isobath. Much of the temperature variability in that

Figure 6. Amplitude and phase of (a–d) temperature and (e–h) salinity seasonal cycle computed using equation (1). Figure 6f shows time of year when freshest water is found over the shelf. Figures 6c, 6d, 6g, and 6h show percentage of total variance explained by the seasonal cycle and by the residue time series. Black circles in Figure 6d indicate location of observations shown in Figure 7.

Figure 7. Least squares fit to temperature glider observations using an equation similar to equation (1), but with three harmonics (with $\omega_3 = 2\omega_1$ and $\omega_5 = 3\omega_1$). Small dots are individual observations. Location of measurements is shown in Figure 6d.
region is related to changes in the position of the foot of the shelfbreak front [e.g., Linder and Gawarkiewicz, 1998].

[20] In contrast to the region north of the Hudson Shelf Valley [e.g., Linder and Gawarkiewicz, 1998; Mountain, 2003; Lentz, 2008b], the central MAB presents a significant salinity seasonal cycle [Mountain, 2003], which is strongly surface-intensified (Figure 6). Amplitudes offshore are as large as 1.4, decreasing with depth and with proximity to the coast. Interestingly, the amplitude of the seasonal cycle is relatively small near the coast. In our observations, surface coastal salinities are low year-round (between 30.5 and 32), with most of the variability occurring on event timescale related to pulses in river discharge and to wind events. Most of the measurements were obtained offshore of the 20 m isobath, however, so it is possible that a larger seasonal freshening takes place inshore of that in a narrow coastal current [Chant et al., 2008a, 2008b]. In these shallow regions, the freshest bottom water is found during fall as mixing increases due to strong wind events. Note, however, that the amplitude of the seasonal cycle there is very small. Offshore of the 30 m isobath, the surface freshening occurs during summer (see also Figure 3), peaking in August. The offshore freshening is due to upwelling favorable winds that transport low-salinity waters offshore [Remie et al., 1999; Johnson et al., 2001; Sanders and Garvine, 2001; Hallock and Marmorino, 2002; Johnson et al., 2003; Lentz, 2004; Choi and Wilkin, 2007; Castelao et al., 2008a, 2008b]. This seasonal freshening is responsible for over 70% of the surface salinity variance. Offshore of the 60 m isobath below 20 m depth, most of the variability (>80%) is due to either changes in the position of the foot of the shelfbreak front [e.g., Linder and Gawarkiewicz, 1998] or to the frequently observed salty intrusions at pycnocline level [e.g., Lentz, 2003].

[21] As mentioned before, we also computed the least squares fit using 3 harmonics in order to better represent asymmetries in the temperature and salinity seasonal evolution (e.g., Figure 7). In that case, the reconstructed time series account for over 90% of the temperature variance in the surface layer, and over 50% below the thermocline offshore of the 40–50 m isobath. For salinity, the variance explained by the reconstructed time series is also larger than before. The coefficients estimated by the least squares fit (equation 1, but with 3 harmonics) for both temperature and salinity are available in the auxiliary material (together with Animation S1, which shows the reconstructed fields) and can be used to reconstruct their seasonal evolution, which might be helpful for modeling initialization and validation efforts in the central MAB.

3.3. Cross-Shelf and Vertical Gradients

[22] Cross-shelf gradients of temperature, salinity and density estimated via linear regression for the regions extending from the coast to the 60 m isobath (‘inshore’) and offshore of the 60 m isobath (‘offshore’) are shown in Figure 8. Note that the color scale for the temperature cross-shelf gradient is inverted, so that red colors in all plots indicate light waters (warm and fresh) toward the coast.

[23] Consistent with Figure 2, the cross-shelf temperature gradient inshore of the 60 m isobath is positive (temperature increases offshore) and roughly depth-independent from December to March. During April and November, the cross-shelf gradient is nearly zero throughout the water column. For the remaining months (May to October), there is substantial vertical variability in the gradient magnitudes. For the most part, gradients are negative. This is especially true at depth, where gradients in excess of ~0.1°C km⁻¹ are found during September and October. This is because waters near the coast are already warmer due to the deepening of the thermocline during fall, but waters below the thermocline near the 60 m isobath are still dominated by the presence of the cold pool (see Figure 2). Only in November, when thermal stratification is destroyed, does the cross-shelf temperature gradient at depth is reduced. Near the surface, however, the situation is very different. Cross-shelf temperature gradients are much smaller, even becoming positive during July and August. It is clear from Figure 2 that isotherms are tilted upward near the coast, which is consistent with the wind-forcing that is typically upwelling favorable during that time of the year (Figure 13) [Glenn et al., 2004].

[24] In contrast to regions north of the Hudson Shelf Valley [e.g., Lentz, 2008b], the central MAB presents strong seasonal variation in the cross-shelf salinity gradients inshore of the 60 m isobath. Gradients are large from fall to spring when low-salinity waters are restricted to close to the coast, decreasing substantially during summer. This is because winds are predominantly upwelling favorable during that time (Figure 13), and low-salinity waters are transported offshore occupying the surface layer throughout the shelf. As near-coastal waters are transported offshore, the freshwater plumes can detach from the coast [e.g., Lentz, 2004], and isolated lenses of low-salinity waters are sometimes found offshore off New Jersey [Castelao et al., 2008b]. Waters near the coast are then saltier due to the decrease in the Hudson River discharge and to upwelling events [Glenn et al., 2004], reverting the sign of the cross-shelf salinity gradient near the surface during July (Figure 8).

[25] The contribution of the temperature (T) and salinity (S) gradients to the cross-shelf density (ρ) gradients was estimated using a linear relation [Shearman and Lentz, 2003],

$$\nabla \rho = \alpha \nabla T + \beta \nabla S \tag{2}$$

where $\nabla$ is the cross-shelf gradient operator, and $\alpha$ and $\beta$ are estimated by linear regression. Their contribution to the density gradients tend to cancel out during winter, but to add to each other during summer/fall. So even though salinity gradients during winter are as large as during fall, density gradients reach a clear maximum in September/October (similar to Shearman and Lentz’s [2003] observations in the northern MAB), with temperature gradients contributing more near the bottom, and salinity gradients more near the surface. A second intensification is also found in April near the surface, associated with the influence of river plumes (see also Figures 4 and 5). In July, both temperature and salinity gradients contribute to the negative cross-shelf density gradients observed near the surface.

[26] The seasonal variability in the cross-shelf density gradients is reflected in the geostrophic flow at the shelf via
the thermal wind relation. Several previous studies support a thermal wind balance in the MAB, even over the inner shelf [e.g., Lentz et al., 1999; Garvine, 2004; Codiga, 2005]. The geopotential anomaly at the surface was computed assuming a level of no motion along the bottom (as in the work by Linder and Gawarkiewicz [1998]), and is shown in Figure 9. The cross-shelf average of the geopotential anomaly was removed for each month to facilitate comparisons. The geopotential anomaly generally decreases offshore indicating southward flow during most of the year. Using several mooring data records at least 200 days long, Lentz [2008b] concluded that maximum equatorward flow due to cross-shelf buoyancy gradients onshore of the 60 m isobath for sites west of 72°W tends to occur in the spring. In the glider observations, the largest differences in the geopotential anomaly across the shelf are found in April and October, when density gradients are largest (Figure 8). During July, the geostrophic flow is to the north inshore of about 50 km from the coast, consistent with isopycnals tilting upward toward the shore (Figure 4) resulting from prevailing upwelling favorable wind-forcing at that time (see Figure 13).

[27] Similarly to the region close to the coast, both temperature and salinity increase offshore during winter offshore of the 60-m isobath. The temperature and salinity gradients are mostly depth-independent (Figure 8), and magnitudes are larger than in the region close to the coast due to the proximity of the shelfbreak front. Their contributions to the density gradients, however, partially cancel each other, and as a result cross-shelf density gradients are not much stronger than in the region close to the coast. During summer and early fall offshore of the 60-m isobath, temperature and salinity gradients continue to oppose each other, since shelf temperatures are lowest at the center of the cold pool, increasing offshore. This is in contrast to the region near the coast, where temperature and salinity gradients added together to generate strong cross-shelf density gradients [Shearman and Lentz, 2003; Lentz, 2008b]. Note, however, that while cross-shelf salinity gradients offshore of the 60-m isobath are large throughout the water column peaking at the surface, cross-shelf temperature gradients are only large below the thermocline. As a result of that, cross-shelf density gradients are relatively large near the surface, decreasing below 20 m depth. It is important to point out that sections don’t extend offshore enough to fully capture the shelfbreak front. Also, observations below 60 m offshore of the 60-m isobath were discarded from the gradient computations because of the small amount of data available to the regression analysis due to the sloping bottom.
Therefore, gradient magnitudes reported here are likely not representative of the shelfbreak front.

Vertical gradients of temperature, salinity and density in the top 30 m of the water column were also estimated for each month (Figure 10). In regions shallower than 30 m, top-to-bottom gradients were computed instead. The contribution of the temperature and salinity gradients to the density vertical stratification is estimated using equation (2), except that the cross-shelf gradient operator is replaced by a vertical derivative (Figure 11). All variables are characterized by strong seasonal variability in the gradient magnitude. Early in the year, thermal stratification is virtually nil, and density stratification is entirely determined by the salinity field (March, Figure 11). From April on, temperature and salinity stratification increases rapidly (Figure 10). The salinity stratification is strongly enhanced near the coast in May and June due to the influence of the Hudson River buoyancy-driven coastal current, which is surface-advected during that time of the year [Yankovsky and Chapman, 1997; Chant et al., 2008a, 2008b]. In May, salinity gradients are still the largest contributor to density vertical gradients inshore of about 30 km from the coast. Offshore of that, however, the temperature vertical gradients contribution becomes just as important (Figure 11). During the peak of the summer, salinity stratification is roughly uniform in the 80 km closest to the coast, but increases offshore due to the presence of the shelfbreak front below the surface layer (Figure 10; see also Figure 3). Temperature stratification in the top 30 m reaches its seasonal peak in July/August at about 90 km from the coast due to the influence of the cold pool below the thermocline. Temperature stratification is the major contributor to the density stratification throughout the shelf during that time (July, Figure 11). After reaching peak intensity in July/August, the magnitude of the salinity stratification decreases rapidly as the low-salinity surface water is mixed with ambient waters, advected along the coast, or transported back onshore by downwelling favorable winds. Near-surface salinity vertical gradients during fall are often very small on the outer shelf (see also Figure 3), a result of salty intrusions at the surface [e.g., Gawarkiewicz et al., 1996]. By November, thermal stratification is already destroyed (Figure 10), and density stratification is once again entirely due to salinity gradients.

3.4. Interannual Variability

The four-year-long time series of quasi-synoptic high-resolution observations along a single cross-section provide a unique opportunity to look at differences in hydrography between the years. Empirical orthogonal decompositions (EOFs) of the temperature and salinity fields were performed after the respective temporal means at each bin were removed from the corresponding time series. This allows for the dominant modes of variability of the system to be extracted, and interannual variability can often be easily identified [e.g., Strub and James, 2002]. Results from EOF decompositions using observations from 2004 only have already been reported by Castelao et al. [2008a]. Therefore, the focus here is on deviations from the patterns reported previously.
Temperature EOFs computed using observations from 2004, 2006–2008 are similar to the EOFs computed using observations from 2004 only [see Castelao et al., 2008a, Figure 3], and thus are not shown here. The similarities occur both for the spatial EOFs and for the amplitude time series, indicating a large degree of repeatability on the thermal structure and little interannual variability. The reader is referred to Castelao et al. [2008a] for a detailed description of the two dominant modes.

The first EOF mode of the salinity field is related to the seasonal widening of the region influenced by low-salinity waters during summer (similar to the seasonal cycle, Figure 6), and is again similar to results obtained using data from 2004 only [see Castelao et al., 2008a, Figure 4] (not shown here). The second mode obtained using the full data set, however, is notably different than when observations from 2004 only are used. The second EOF mode, which explains about 15% of the total variance, is characterized by small values everywhere, except near the surface offshore of ~80 km from the coast, where values are large and negative (Figure 12). There is also a slight positive increase in the values offshore near the bottom. The amplitude time series (broken down by year to facilitate comparisons) show that, except for a few isolated observations, values are small in 2004, 2007 and 2008. In 2006, however, values are large and positive in July and August, and small for the remainder of the year. Therefore, surface salinities offshore of 80 km from the coast were approximately 1.2 fresher in summer 2006 than in other summers, while bottom waters offshore were ~0.5 saltier.

Castelao et al. [2008b] used glider observations, HF radar surface current measurements, and the analytical model developed by Lentz [2004] to investigate the details of the salinity distribution in the surface layer off New Jersey during 2006. Several lenses of low-salinity water were found offshore of ~80 km from the coast. Observations indicate that the freshwater was transported offshore by an offshore-directed jet situated roughly along the southern flank of the Hudson Shelf Valley [Castelao et al., 2008b; Zhang et al., 2009], and that the jet transport was correlated with upwelling favorable winds. The second EOF mode of the salinity field (Figure 12) indicates that the occurrence of the lenses in 2006 with very low-salinity waters was an anomalous event, which was not repeated in other years.

Since the transport of freshwater offshore seems to be related to upwelling winds [Castelao et al., 2008b], it is instructive to compare the cumulative (time-integrated) wind stress [Pierce et al., 2006; Barth et al., 2007] between the different years, and relate that to the Hudson River discharge (Figure 13). There are large differences between the years in both time series. Most notably, the Hudson River discharge in 2006 was spread out throughout the first half of the year, with the largest discharge pulse occurring in late June/early July. River discharge at around 1000 m³ s⁻¹ continued until late July. This is in contrast to the other years, when most of the river discharge was concentrated
earlier in the year (March and April in 2007 and 2008; peak in late May in 2004), after which discharges were close to zero for at least a couple of months. In those years, therefore, peaks in river discharge were followed by a few weeks of fluctuating winds (downwelling or weakly upwelling favorable) before the ‘upwelling season’ begun. In 2006, on the other hand, the peak in river discharge in early June, the large peak in late June/early July and the subsequent pulses in July and early August occurred during a time when winds were already strongly upwelling favorable. Since the integrated Hudson River discharge from March to mid-July was nearly identical in 2006, 2007 and 2008 at \( \sim 12 \text{ km}^3 \), this suggest that the relatively large difference in the freshwater content in the offshore region (Figure 12) was due to the timing of the pulses in river discharge relative to the winds. When large discharge pulses coincide with strong and persistent upwelling winds as in 2006, offshore transport of freshwater is enhanced, both by surface Ekman transport and via the transport pathway along the southern flank of the Hudson Shelf Valley [Castelao et al., 2008b; Zhang et al., 2009], helping explain the anomalously freshwater found offshore. The reason for the slight increase in bottom salinities offshore during summer 2006 (Figure 12) is unclear.

4. Summary and Conclusions

[34] The seasonal and interannual variations in shelf hydrography in the central Middle Atlantic Bight (MAB) are described based on four years of high-resolution, quasi-synoptic glider observations. Observations were collected off New Jersey in a region that has been relatively under
sampled compared to the northern MAB. The evolution of the thermal field is consistent with observations from the New England shelf [e.g., Beardsley et al., 1985; Lentz, 2008b], with the surface layer warming up from March/April to August, and with the thermal stratification breaking down during fall. In contrast to the northern MAB, there is also substantial seasonal variation in salinity (and in the cross-shelf salinity gradients) with the amplitude of the fit to the first annual harmonic as high as 1.4 in the surface layer offshore. The seasonal variability in the surface layer explains over 70% of the total salinity variance in the central MAB, and is mostly related to upwelling winds transporting low-salinity waters offshore during spring/summer. On shorter time scales, most of the salinity variability during the stratified season is related to internal waves propagating along the halocline, to surface and pycnocline salty intrusions, and to shifts in the position of the foot of the shelf-break front. Cross-shelf temperature, salinity and density gradients are depth-independent during winter, but vary considerably in the vertical during spring/summer. Density gradients are positive most of the time, consistent with the mean southward flow characteristic of the MAB. During summer, however, isotherms are tilted upward toward the coast in response to predominantly upwelling favorable winds. Upwelling winds also bring saltier water at depth close to the coast to replace the fresher water that is transported offshore. The temperature and salinity fields, therefore, combine to create negative cross-shelf density gradients in the upper water column and a northward geostrophic flow near the coast during summer. All variables are characterized by a strong seasonal cycle in their vertical gradients. While early in the year density stratification is entirely due to salinity, temperature is the major contribution to density vertical gradients during summer, when stronger stratification in the top layer is found offshore due to the influence of the cold pool and of the shelfbreak front. The transition between salinity-dominated to temperature-dominated stratification occurs during the spring, when salinity contributes more near the coast, while temperature is more important offshore.

[35] During the four years observed, temperature interannual variability was small. Surface salinity, on the other hand, was anomalously low during summer 2006 by ~1.2 psu, especially offshore of the 60 m isobath. The anomaly was due to pulses in discharge from the Hudson River at a time when winds were already predominantly upwelling favorable, which favors offshore transport of low-salinity waters. This was in contrast to the more common situation when the peak discharge occurs earlier in the year when winds fluctuate often between downwelling and weakly upwelling favorable.

[36] The repeated high-resolution glider observations allowed for a robust estimation of the seasonal cycle of the hydrographic fields. We believe this can be quite useful for modeling initialization and validation efforts, since reconstructed time series using three harmonics (available in the auxiliary material) explain a large fraction of the temperature and salinity total variance in the central MAB.

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