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PROCESSES THAT AFFECT STRATIFICATION IN SHELF WATERS

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Abstract. Processes that affect stratification in the South Atlantic Bight can be divided into those that create it and those that destroy it. Stratification is created by surface heating, freshwater runoff, and rain and is destroyed by evaporation, cooling, and surface and bottom stresses. We show that freshwater runoff is the primary source of buoyancy that creates stratification in the inner and middle shelf near rivers. Heating is equally important over all of the shelf. Up to 20 mW m^{-2} of mixing power may be required to destroy runoff-related stratification. Wind stress can exert a maximum of about 3 mW m^{-2} . Another process that creates and destroys stratification is the advection of buoyancy, a term usually neglected. We show that variations in stratification caused by advective flux of buoyancy often dominate all other processes. The power required to destroy advectively created stratification may reach 3 mW m^{-2} .

Introduction

Stratification of shelf waters reflects the balance between buoyant forces that result from heating and cooling, evaporation and precipitation, and runoff, and mixing forces such as wind stress at the surface and current stress at the bottom. Since most of these forces vary at a variety of frequencies from tidal to seasonal, and even geological for that matter, we can expect variations in stratification to occur at similar frequencies. In this paper we will examine seasonal variations in observed stratification and in the processes that affect stratification. Most of our discussion will be based on data from the southeastern United States continental shelf, the South Atlantic Bight (SAB) (Figure 1). The cycle of stratification in the SAB is dramatic because of the extreme seasonal and latitudinal variations that occur there and thus is a good laboratory for examining stratification processes.

In this paper we will show that while some of the processes that affect stratification are well known, some of the most important terms, such as runoff and evaporation, are difficult to parameterize, and thus a true determination of seasonal variations in buoyancy flux cannot be determined

by summing terms. We will also show that the advective transport of heat into the region may at times exceed other forms of buoyancy flux.

Available oceanographic data from the SAB were used by Atkinson et al. [1983] to create charts showing mean oceanographic conditions at monthly increments. These data were used to calculate the monthly mean bulk stratification for areas of the SAB defined by 1° of latitude and the 0-, 20-, 40- and 60-m isobaths (Figure 2). The mean bulk stratification is the simple difference between near-bottom and near-surface density. During the winter, bulk stratification is low because of high wind stress, low runoff, and negative heat flux. In early spring, bulk stratification increases first in the southern SAB because of decreased winds, increased heating, and runoff. In the summer, bulk stratification increases throughout the SAB partly because of positive heat flux and decreased wind stress but also because of subsurface Gulf Stream intrusions. By October, bulk stratification generally decreases because of increased wind stress; however, an area of increased bulk stratification occurs off southeast Florida because of southward advection of fresher coastal waters.

These mean monthly distributions lead to several conjectures regarding stratification in the SAB:

1. Stratification in the spring and summer increases because of (1) increased insolation, (2) reduced winds, and (3) subsurface intrusions of denser Gulf Stream water.
2. Stratification in the fall and winter generally decreases because of (1) decreased insolation and (2) increased winds.
3. Stratification can increase locally because of the trapping of low-salinity water along the coast.

Stratification is the net result of energy inputs creating buoyancy forces plus mixing forces that destroy stratification. We now discuss the various forms of buoyant energy.

Flux of Buoyant Energy

Buoyancy represents a force that tends to stabilize the water column. Conservation of buoyancy,

Fig. 1. The southeastern United States continental shelf area.

β , can be expressed in terms of advective and nonadvective components which must balance the divergence in buoyancy flux, N , as follows [Phillips, 1966, p. 21]:

$$\frac{d\beta}{dt} = \frac{\partial\beta}{\partial t} + \vec{v} \cdot \nabla\beta = -\nabla \cdot \vec{N} \quad (1)$$

Local
Advective
Mean
Change
Change
Flux

where $\beta = -g(\nabla\rho/\rho_0)$ is the buoyancy per unit volume, $\vec{N} = \vec{v}'\beta'$ is the three-dimensional buoyancy flux vector, $\vec{v}$ is the three-dimensional velocity vector, and ∇ is the three-dimensional gradient operator. Primed quantities represent fluctuations about the mean. However, in the case of oceanic buoyancy flux the advective term is usually neglected, and (1) reduces to [Phillips, 1966, p. 223]

$$\frac{\partial\beta}{\partial t} = -\frac{\partial N}{\partial z} \quad (2)$$

where $N = \overline{\omega'\beta'}$, the vertical flux of buoyancy. The implication of neglecting the advection of buoyancy will be discussed later. For the purpose of this paper we vertically integrate (2), with the left-hand side of (2) becoming the change in water column buoyancy and the right-hand side becoming the buoyancy flux across the air-sea interface as follows:

$$N(o) = \frac{g}{\rho} \left\{ (Q_s - LE) \frac{\alpha_v}{c_p} - ES + F \right\} \quad (3)$$

where

- $N(o)$ net flux of buoyancy across the air-sea interface;
 g acceleration of gravity;
 ρ density of seawater;
 $Q(s)$ net gain of sensible heat by radiation and conduction;
 L latent heat of evaporation;
 E evaporation rate;
 α_v volumetric coefficient of expansion;
 c_p specific heat at constant pressure;
 S salinity;
 F freshwater flux, $\text{kg m}^{-2}\text{s}^{-2}$.

Equation 3 without F is identical to Phillips' [1966, p. 223] expression.

The $Q(s)$ term represents the rate at which buoyancy is generated by heat additions (insolation, sensible heat, convection, latent heat of evaporation). The second term, LE , is the rate at which buoyancy is lost through the latent heat of evaporation. The third term, ES , is the rate at which mass is left behind by evaporation. The fourth term, F , is the rate at which buoyancy is generated by runoff additions.

We will now quantify the terms in (3) for the SAB. In the discussion of the magnitude of buoyancy flux terms we will also give the power required to dissipate such flux distributed over a 10-m deep water column using the relation

$$\text{Required Power} = \text{Buoyancy Flux} * \text{Density} * \text{Depth}/2 \quad (4)$$

The power is expressed in milliwatts per square meter (mW m^{-2}).

The Heat Flux Term, $Q(s)$

The heat flux $Q(s)$ (Figure 3) was determined by examining the monthly change in the mean monthly temperature in shallow (<20 m) coastal waters for the inner shelf. It was assumed that there were no advective variations in the 2° latitude section off Georgia and South Carolina over which the calculation was made. This is similar to the technique used by Pingree and Griffiths [1977] for the Celtic Sea. Using this technique, the derived quantity actually represents the sum of all heat flux terms, including latent heats. The calculated buoyancy flux varies between -4.22×10^{-8} and $3.28 \times 10^{-8} \text{ m}^2 \text{ s}^{-3}$ (-0.21 to $+0.16 \text{ mW m}^{-2}$). Maximum heat gain occurs from May through June, while maximum heat loss occurs from October through February. Insolation rates as high as $12.66 \times 10^{-8} \text{ m}^2 \text{ s}^{-3}$ ($+0.63 \text{ mW m}^{-2}$) have been estimated for this area from climatological data [Blanton and Atkinson, 1983]. Thus, a considerable amount of heat is probably lost by back radiation and latent heat of evaporation. We assume that these data represent the net nonadvective heat flux to all parts of the shelf.

Fig. 2. Mean monthly bulk stratification [from Atkinson et al., 1983]. Units are σ_t .

Fig. 3. Mean monthly heat flux.

The Evaporation-Precipitation Term, ES

Evaporation in the SAB is approximately 300 cm yr^{-1} [Bunker, 1976], and we must assume that it is aseasonal. Precipitation at coastal stations averages 120 cm yr^{-1} . It is probable that precipitation over shelf waters is less than at land stations, but data are not available to confirm this. We assume that $E - P$ is 180 cm yr^{-1} and aseasonal. This converts to a buoyancy flux of $-6.66 \times 10^{-8} \text{ m}^2 \text{ s}^{-3}$ (-0.3 mW m^{-2}).

The Freshwater Term, F

Because of the many small rivers, freshwater flowing into the SAB approximates a line source between Jacksonville, Florida, and Cape Fear, North Carolina. Discharge peaks in spring, and maximum monthly values based on 20-year means range between 1300 and $4000 \text{ m}^3 \text{ s}^{-1}$ [Blanton and Atkinson, 1983] over the 400-km length of shoreline (Figure 4). The effect of coastal runoff on buoyancy flux is difficult to evaluate because unlike heat flux and evaporation, which generally occur equally over a large area, runoff is a coastal source and the manner in which it spreads over the shelf determines the resultant buoyancy flux. Blanton and Atkinson [1983] assume that the rivers flow in along 400 km of coastline and spread over the inner 20 km of the shelf. With this assumption, we calculate a buoyancy flux that ranges between 160 and $400 \times 10^{-8} \text{ m}^2 \text{ s}^{-3}$ (8 to 20 mW m^{-2}). Clearly, the buoyancy flux due to river discharge is highly variable and no doubt is the most important contribution to buoyancy flux for the inner shelf, with diminishing impact offshore depending on the degree of spreading.

Flux of Mixing Energy (Mixing Power)

Several processes apply power at the surface and bottom boundaries which acts, along with a negative buoyancy flux, to weaken, if not destroy, stratification. Wind stress during storms can easily produce vertically mixed water on most areas of shallow continental shelves. Wind stress

also produces subtidal currents which exert stress at the bottom boundary where additional work is performed.

Tidal currents also perform work at the bottom boundary which can vertically mix stratified water. Tidal currents are of relatively constant amplitude and are at a higher frequency than wind-induced currents. Tidal power available for mixing amounts to $0.03\text{--}0.2 \text{ mW m}^{-2}$ [Blanton and Atkinson, 1983].

Wind and Bottom Stress Work

The wind regime in the SAB is quite variable, with southerly winds occurring during the summer months and northerly winds dominating the winter months [Weber and Blanton, 1980]. The effect of wind on stability is twofold: Direct local forcing causes vertical mixing, while wind-induced advection may cause low-salinity coastal water to move into other regions, effectively causing a buoyancy flux in the offshore region or vice versa. Vertical mixing due to local wind forcing is not intense because of the relatively weak mean winds, although occasional storms may cause strong mixing. Wind-induced advection can have a substantial effect on stratification, since significant amounts of fresh water can be advected offshore [Blanton and Atkinson, 1983]. This process is particularly effective in the spring when southerly winds cause the offshore advection of fresh coastal waters.

Work that mixes the water column due to wind can be derived by multiplying wind stress by U_* (friction velocity):

$$\rho_w U_*^2 = \rho_a C_d U_a^2 \quad (5)$$

where ρ_a and ρ_w are the densities for air and water, respectively, U_a is wind speed, U_* is friction velocity, and C_d is a drag coefficient. Since $\rho_a/\rho_w = 10^{-3}$, the friction velocity is

$$U_* = (10^{-3} C_d)^{1/2} U_a \quad (6)$$

Thus the work done by the wind, W_w , is

$$W_w = \rho_w U_*^3 = \rho_w [(\rho_a/\rho_w) C_d]^{3/2} U_a^3 \quad (7)$$

Fig. 4. Mean monthly river flow in units of total cubic kilometers and meters per square meter of shelf area.

TABLE 1a. Typical Values of Buoyancy Flux and Mixing Powers in the SAB

Term	Typical Range	
	Buoyancy Flux, $10^{-8} \text{ m}^2 \text{ s}^{-3}$	Equivalent Mixing Power, mW m^{-2}
1. Heat additions	-4.22 to 3.28	0 to 0.16
2. Evaporation of water	-6.66	none required
3. Runoff of fresh water	160-400	8-20
4. Advective flux	40-400	2-20
5. Wind stress		1.0-3.1
6. Bottom currents		0.5-1.1
7. Tidal current stress		0.03-0.2

Buoyancy flux values have been converted to power units for comparison with the mixing power data by assuming a 10-m deep water column.

The drag coefficient, C_d , is typically 2×10^{-3} . Mean winds in the SAB would result in $W_w = 1.0 \text{ mW m}^{-2}$, while typical storm winds would yield a higher value, $W_w = 3.1 \text{ mW m}^{-2}$. Similarly, the work done by bottom currents, W_c , is

$$W_c = \rho_w C_d \frac{3}{2} U_c^3 \quad (8)$$

where U_c is the observed near-bottom current. For a typical range of bottom currents the work would be 0.5 to 1.0 mW m^{-2} .

Discussion of Buoyancy Flux and Available Mixing Work

Now that we have discussed each of the normal terms in the buoyancy flux equation and the mixing work terms, it will be useful to compare them.

Table 1 shows that buoyancy flux related to freshwater runoff is an order of magnitude higher than other buoyancy sources. This buoyancy flux is maximum near the coast and decreases offshore and toward the southern and northern portions of the SAB, where runoff decreases. Thus in regions remote from runoff we would predict a net buoyancy flux closer to the sum of the heat flux and ES component. Since, as will be shown, we do observe stratification during the summertime, when buoyancy flux due to runoff is relatively low, we can assume that buoyancy flux due to the heating term $Q(s)$ is greater than the negative buoyancy flux caused by mass left behind during evaporation (ES).

When one compares the available mixing energy to the components of buoyancy flux, it is apparent that most winds would be sufficient to dissipate any stratification resulting from a positive buoyancy flux unless runoff is present or winds are weak. The maximum possible buoyancy flux related to runoff would require more mixing work than is available during most wind events. Thus we would predict that inner shelf waters would remain stratified until strong winds during a large storm event occur or until the advection of surface waters creates a local negative buoyancy flux.

Another source of buoyancy flux is the onshore flow of dense subsurface intrusions [Atkinson, 1977]. These would create local areas of high stratification that also would require more mixing power than is available. The fact that outer shelf waters remain stratified during the summer suggests that advective buoyancy flux overcomes any local mixing forces.

Examples of the Formation and Destruction of Stratification, April 1980

During a large experiment conducted in April 1980, when the shelf was strongly stratified in the vertical due to runoff, a storm occurred that caused stratification to break down over various parts of the shelf in response to variations in wind-induced vertical mixing. Changes in stratification were determined by calculating the potential energy of the water column relative to a

TABLE 1b. Power Required to Vertically Mix Observed Stratified Regimes

Regime	Power, mW m^{-2}
Inner shelf, high runoff (spring 1980)	1.10
Inner shelf, low runoff (autumn)	0.11
Inner shelf, solar heating during May	0.02
Inner and middle shelf in April 1980	0.0-4.00

Fig. 5a. Surface and vertical temperature distributions for April 12-15, 16-19, and 20-23, 1983. Arrows indicate locations of upper and lower sections.

completely mixed state [Simpson et al., 1978]. The potential energy, PE, is calculated by

$$PE = \int_{-h}^0 \beta z \, dz = \frac{1}{\rho} \int_{-h}^0 (\rho - \bar{\rho}) g z \, dz \quad (9)$$

where z is positive upward, h is the water depth, and ρ is density, whose vertical average is $\bar{\rho}$.

We mapped the continental shelf off Georgia and Florida on three occasions during April 12-25, 1980. Winds were northward prior to this period and until a cold front passed on April 14. After-

ward wind stress shifted southward until April 23, when stress again shifted northward. Available evidence [Lee and Brooks, 1979] suggests that the response of shelf currents to wind follows a "frictional equilibrium" model. The sequence of wind stress described above can be summarized as follows:

Before April 10: Winds weak.

April 10-17: Wind stress northward and average currents northward with offshore/onshore flow in the surface/bottom layer.

Fig. 5b. Surface and vertical salinity distributions for April 12-15, 16-19, and 20-23, 1983. Arrows indicate location of upper and lower sections.

April 17-23: Wind stress strong and southward accompanied by southward currents except for on-shore/offshore flow in surface/bottom boundary layer.

April 23-27: Wind stress and currents again northward as in the period April 10-15.

Thus the wind regime and the hydrographic response can be divided into three episodes. Episode 1 was composed of alongshelf wind stress northward from April 12 to 17; episode 2 was composed of alongshelf stress southward from April 17

to 22; and episode 3 represents a return to northward flow. We have a complete map of the hydrographic structure of the inner and middle shelf representative of each episode (Figure 5).

At the beginning of episode 1 the water was stratified at all stations. After April 17 the inner shelf was well mixed while the middle shelf continued to lose PE. Thus northward wind stress was quite effective in mixing middle shelf waters. During episode 2 the inner shelf remained well mixed while PE on the middle shelf continued

Fig. 5c. Surface and vertical density distributions for April 12-15, 16-19, and 20-23, 1983. Arrows indicate location of upper and lower sections.

to decrease at the same rate. By episode 3 the entire inner and middle shelf was nearly totally mixed.

Using equation (9), we calculated PE for the three surveys and then calculated the difference in PE at each station in each successive survey (Figure 6). Potential energy during the first mapping varied from 200 J m^{-2} in inshore waters to 3000 J m^{-2} over the middle shelf. During the second mapping, PE over the inner shelf was near zero, and middle shelf PE had decreased by 500 to 1500 J m^{-2} . By the third mapping, PE over the

middle shelf decreased another $500\text{--}1500 \text{ J m}^{-2}$. Parts of the inner and middle shelf were already vertically mixed and thus could lose no more energy.

Both middle and inner shelf decreased in PE by about 1100 J m^{-2} over the 5-day period of episode 1. This represents a power of $\delta\text{PE}/\delta t = 2.25 \text{ mW m}^{-2}$. During episode 2 the middle shelf lost about another 1000 J m^{-2} . The inner shelf was already well mixed, although in some places PE increased during episode 2.

Boundary work from wind and bottom stress must

Fig. 6. Horizontal distribution of potential energy PE and change in potential energy, ΔPE .

be approximately 2-4 mW m^{-2} , based on changes suggested in Figure 6, to vertically mix the stratified waters observed on the middle and inner shelf. We used wind speed data in equation (7) and current speed data in equation (8) to estimate boundary work for April 1980.

Wind data from the Savannah Navigational Light Tower (SNLT) were used to estimate the average work done by wind on the inner shelf. Data from a National Oceanic and Atmospheric Administration (NOAA) buoy on the middle shelf was used to estimate wind work there. Currents were also measured 8 m off bottom at SNLT. There was a current meter located 3 m off bottom in 28 m of water off Savannah. The drag coefficients used for the estimates of work were as follows:

$$\begin{aligned} C_d &= 2 \times 10^{-2} \text{ (wind drag)} \\ C_d &= 5 \times 10^{-3} \text{ (current drag at 8 m)} \\ C_d &= 10 \times 10^{-3} \text{ (current drag at 3 m)} \end{aligned}$$

Drag coefficients are often observed to be higher for observations located closer to bottom [Winant and Beardsley, 1979]. Using the above values for C_d , we were able to estimate the total work done by wind stress on the inner shelf and the total work done by bottom currents for the first two episodes (Table 2).

Total boundary work (Table 2) varied between 1.4 and 4.2 mW m^{-2} . The higher value occurred because of the high winds and bottom currents observed during the first episode. In all cases, wind stress work was higher than bottom stress work by a factor of 2 to 3. We were unable to balance the potential energy changes observed in the hydrography: less PE was lost than energy put in. By combining all hydrographic stations at depths less than 25 m in the inner shelf regime, we observed PE changes during episode 1 that were within 40% of the boundary work expended. PE changes in the middle shelf regime (25-40 m) were

TABLE 2. Summary of Potential Energy Changes Compared With Work Done by Wind and Bottom Currents on the Inner and Middle Shelf for the Two Episodes

	Inner Shelf	Middle Shelf
<u>Episode 1</u>		
Energy lost	1100	1100
Wind stress work	1340(3.10)	690(1.60)
Bottom stress work	480(1.11)	260(0.60)
Total boundary work	1820(4.21)	950(2.20)
<u>Episode 2</u>		
Energy lost	already mixed or increased	1000
Wind stress work	520(1.20)	410(0.95)
Bottom stress work	260(0.60)	200(0.46)
Total boundary work	780(1.80)	610(1.41)

All energy values are in joules per square meter. Numbers in parentheses represent power expenditures in $10^{-3} \text{ mW m}^{-2}$.

Fig. 7. Typical summertime distribution of temperature, salinity, and density.

Fig. 8. Typical temperature, salinity, and density profiles.

within 20% of the available boundary work expended. These discrepancies are partially due to uncertainties in drag coefficients and the parameterizations leading to equations (7) and (8). We have neglected the advection of buoyancy between the middle and inner shelf within each episode, which would require additional boundary work to vertically mix the additional input of PE.

As an additional test on the balance we calculate the drag coefficient by assuming that the observed wind stress exactly caused the observed loss in PE. Thus, from equation (7),

$$\rho_w [(\rho_a / \rho_w) C_d]^{3/2} U_a^3 = \frac{\delta PE}{\delta t} \quad (10)$$

The average wind speed off Savannah during episode 1 was $U_a = 9.65 \text{ m s}^{-1}$. With $\rho_a / \rho_w = 10^{-3}$ we calculate a drag coefficient, C_d , of 2.0×10^{-3} . This value, perhaps fortuitously, is identical to the "traditional" C_d of 2×10^{-3} used in many studies [Blanton, 1981].

This example demonstrates that winds, even relatively strong winds, were not sufficient to totally mix the middle shelf waters because of the excess buoyancy caused by low-salinity surface waters. Only when the surface waters were advected away did total mixing occur.

The observations we have just discussed demonstrate that stratification can be created and destroyed in a relatively short period of time depending on the balance of advective buoyancy flux and wind and bottom stress. The discussion of the observations also introduced the magnitude of the mixing processes that destroy stratification which is developed during seasonal heating and the critical nature of the balance between advective buoyancy flux and wind-related mixing.

Example of Advective Buoyancy Flux

In this paper we have so far neglected advective buoyancy flux because it is usually difficult to quantify. In this section we will use data from the summer of 1981 to show the importance of advective buoyancy flux due to intrusions.

During the summer in the SAB the total heat flux can reach 0.6 mW m^{-2} and runoff is low; thus the total nonadvective buoyancy flux is in the $0.4\text{--}0.8 \text{ mW m}^{-2}$ range. Note that we are using the required mixing power for a 10-m deep water column, so this flux range would be less by a factor of 4 at the 40-m isobath. Mixing power during the summer is low, with up to 0.3 mW m^{-2} from tides and less than 1 mW m^{-2} from winds. However, even though the mixing power is low, it is probably higher than the nonadvective buoyancy flux. Thus we would expect to see the SAB shelf in the summer to be mixed; however, this is not the case. Evidently, because of Gulf Stream intrusions, the shelf receives a high enough advective buoyancy flux, in terms of the subsurface onshore flow of high-density water, to offset mixing forces. A typical cross-shelf section (Figure 7) shows a strong two-layered system with a very strong pycnocline. The pycnocline extends onshore but eventually breaks down, and very nearshore waters are typically mixed. The horizontal maps of surface and bottom temperatures show the extensive area impacted by the intrusions. In the absence of intrusions the bottom temperatures would be nearly the same as surface temperatures.

Vertical gradients (Figure 8) are extremely high. We have found that cold, dense Gulf Stream water invades the shelf waters periodically, producing a varying input of buoyancy. As an example of these variations we show the time variation in temperature, salinity, density (Figure 9), and PE (Figure 10) at three locations on the shelf at the 20-, 32-, and 40-m isobaths at 30°N . The surface currents at the 45-m isobath are also shown (Figure 11). The locations were highly stratified mainly because of temperature. During the study we identified three cooling events at the 45-m station but only one at the 32-m isobath, indicating the varying distance the subsurface intrusions move onshore. Examination of the current meter record suggests that the events at the 45-m isobath are related to Gulf Stream variations. At the 32-m isobath, only the event of July 25–30 was observed, reflecting the effect of the large eddy event that occurred then. The change in PE at the three stations shows that the mean at the 45-m isobath was higher than the mean at 32 or 20 m, reflecting the proximity of the Gulf Stream. Adjusting for depth, the 45-m PE is still higher than the 32- or 20-m station.

Potential energy varies by about 800, 1000, and 2000 J m^{-2} over a 10-day period at the 20-, 32-, and 45-m isobaths, respectively. The variation at the 45-m isobath between July 20 and 27 was related to the passage of an eddy (see Figure 11) that caused eddy-induced upwelling at the shelf break, weakening the pycnocline and reducing the potential energy. Simultaneously, cold water was forced further onshore (see Figure 9), increasing PE at the 32-m isobath. At the 20-m isobath the onshore buoyancy flux was not observed; either it did not extend that far, or mixing removed it. PE variations at the shelf break (45 m) are more

Fig. 9. Time variations in temperature, salinity, and density at inner, middle, and outer shelf locations off St. Augustine during the summer of 1981.

Fig. 10. Time variations in the stratification parameter (potential energy) at the three locations shown in Figure 9.

dependent on direct contact with the Gulf Stream, while further onshore the variations are of a more indirect nature. After the July 20-27 event, PE again increased at the 45-m isobath because of the influx of warmer surface Gulf Stream waters. At the 35-m isobath, PE remained high until wind events in mid-August caused mixing, thereby reducing PE.

During the July 20-27 intrusion event, PE increased at the 35-m isobath by about 1000 J m^{-2} , which would require about 1.5 mW m^{-2} of mixing power to remove. This is more than could be supplied by tidal or wind-generated currents, and winds would have to be higher than normal to dissipate this buoyancy. Thus, as expected, the buoyancy advected in on July 20-27 remained.

The buoyancy increases at the outer shelf are no doubt mainly removed by advective processes rather than mixing forces, since over 2 mW m^{-2} of mixing work would be required, and such energy is not available during the summer.

Conclusions

As we have shown in the previous discussions, stratification in the SAB varies depending mainly on the balance between buoyancy provided by runoff, heating, and positive advective buoyancy flux, and mixing resulting from winds and negative advective buoyancy flux. Runoff is most important in the inner and middle shelf, while negative and positive advective buoyancy flux is more important over the middle and outer shelf. Winds are important sources of mixing energy only during storm events. Mean winds are too weak to affect stratification.

In a seasonal sense, runoff is more important during the spring, when it reaches a maximum. Advective buoyancy flux is more important when shelf waters are of a differing density than near-bottom Gulf Stream waters at the shelf break. This occurs mainly during the summer, when shelf waters are heated and subsurface intrusions are observed. Latitudinally, the effects of runoff are most pronounced adjacent to the rivers with less effect to the north and south.

Stratification variations due to advective buoyancy flux probably occur more frequently toward the south, where the density contrast is greater [Atkinson, 1977; Atkinson et al., 1983],

Fig. 11. Currents and temperature at the 45-m isobath. Forty-hour low-pass-filtered data from four depths are shown. Data were provided by T. Lee, University of Miami.

but can extend across the shelf at probably any location given favorable conditions. As our example has demonstrated, the stratification variations due to advective buoyancy flux can be equal to or greater than all other forms of buoyancy or mixing.

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