Estuarine boundary layer mixing processes: insights from dye experiments.

Robert J. Chant\textsuperscript{1*}, Wayne R. Geyer\textsuperscript{2}, Robert Houghton\textsuperscript{3}, Elias Hunter\textsuperscript{1} and James Lerczak\textsuperscript{2}.

September 12\textsuperscript{th}, 2006

Submitted to the \textit{Journal of Physical Oceanography}

*Corresponding author

\textsuperscript{1}IMCS Rutgers University 71 Dudley Road New Brunswick New Jersey

\textsuperscript{2}Woods Hole Oceanographic Institution Woods Hole MA

\textsuperscript{3}Lamont Doherty Earth Observatory, Columbia University New York.
Abstract

A series of dye releases in the Hudson River Estuary elucidated diapycnal mixing rates and temporal variability over tidal and fortnightly timescales. Dye was injected in the bottom boundary layer for each of four releases, during different phases of the tide and of the spring-neap cycle. Diapycnal mixing occurs primarily through entrainment that is driven by shear production in the bottom boundary layer. On flood the dye extended vertically through the bottom mixed layer, and its concentration decreased abruptly near the base of the pycnocline, usually at a height corresponding to a velocity maximum. Boundary layer growth is consistent with a one-dimensional, stress-driven entrainment model (Kato and Phillips, 1969; Trowbridge, 1992). A model was developed for the vertical structure of the vertical eddy viscosity in the flood tide boundary layer that is proportional to \( \frac{u^2}{N^2} \), where \( u^* \) and \( N^\infty \) are the bottom friction velocity and buoyancy frequency above the boundary layer. The model also predicts that the buoyancy flux averaged over the bottom boundary layer is equal to \( .06N^\infty u^2 \), or based on the structure of the boundary layer equal to \( .1N_{bl}^\infty u^2 \), where \( N_{bl} \) is the buoyancy frequency across the flood tide boundary layer. Estimates of shear production and buoyancy flux indicate that the flux Richardson number in the flood-tide boundary layer is 0.1-0.18, and this is consistent with the model indicating that the flux Richardson number is between 0.1-0.14. During ebb, the boundary layer was more stratified, and its vertical extent was not as sharply delineated as in the flood. During neap tide the rate of mixing during ebb was significantly weaker than on flood, due to reduced bottom stress and stabilization by
stratification. As tidal amplitude increased ebb mixing increased and more closely resembled the boundary-layer entrainment process observed during the flood. Tidal straining modestly increased the entrainment rate during the flood, and it restratified the boundary layer and inhibited mixing during the ebb.
1. Introduction

In recent years numerous studies have characterized tidal period variability of vertical mixing in estuarine environments (Lu et al., 2000; Peters and Bokhorst, 2000; Peters, 1997; Rippeth et al., 2002; Simpson et al., 2002; Stacey et al., 1999; Stacey and Ralston, 2005; Trowbridge et al., 1999; Geyer et al., 2000). Simpson et al. (1990) introduced the phenomenon of tidal straining, in which vertically sheared tidal current interacts with the along channel density gradient. Tidal straining was found to augment stratification during the ebb and reduce it during the flood, contributing to the now familiar “tidal asymmetry” in vertical mixing (Jay and Smith, 1990). Turbulence measurements by Peters and Bokhorst, (2000) and Trowbridge et al., (1999) provided direct confirmation of tidal asymmetry in turbulence; they also found significant spring-neap variations in the asymmetry, with greater asymmetry occurring during neap tides. Similar findings were obtained by Geyer et al., (2000), based on an analysis of the estuarine momentum balance.

Although considerable attention has been focused on turbulent stress and dissipation of turbulent kinetic energy in estuaries (e.g., Rippeth et al., 2002; Stacey and Ralston, 2005; Stacey et al., 1999; Peters and Bokhorst, 2000; Trowbridge et al., 1999), the quantification of vertical salt flux (or equivalently, buoyancy flux) has been more limited. Peters and Bokhorst (2001) made estimates of buoyancy flux using the Osborn (1980) relation based on a local turbulent kinetic energy balance. This method requires an
*a priori* estimate of the flux Richardson number, which they derive from laboratory measurements and earlier field studies.

Another approach to obtaining buoyancy flux estimates is to quantify the vertical spreading of a deliberate tracer, such as dye or SF₆. Tracer studies of buoyancy flux have the advantage that they do not require *a priori* assumptions about the turbulence. Ledwell *et al.*, (1993) provided an accurate estimate of the mixing rate in the thermocline of the North Atlantic with SF₆, and similar methods have been used in coastal waters with Rhodamine and Fluorescein dye (Sundermeyer and Ledwell, 2001; Ledwell *et al.*, 2004; Houghton-*et al.*, 2004). In this paper we present the results of a set of dye tracer releases in the Hudson River estuary to quantify the vertical mixing rate in different conditions through the tidal cycle and spring-neap cycle of tidal amplitude. Section 2 describes the dye experiments while section 3 characterizes the tidal period and spring-neap variations in mixing observed with the dye tracer. In section 4.1 we focus on flood-tide mixing and show that boundary layer growth occurs due to entrainment and that this growth rate is consistent with the Trowbridge (1992) model. In section 4.1 we also access the relative roles of entrainment and tidal straining to boundary layer growth. This is followed by analysis that quantifies vertical mixing rates and relates vertical eddy diffusivity to the bottom friction velocity and overlying stratification. In section 4.2 ebb boundary layer mixing is discussed and in section 4.3 we provide estimates of the vertical salt flux and of the flux Richardson number. Conclusions are drawn in section 5.
2. The Dye Experiments

The dye studies were conducted in the Hudson River Estuary in May 2002 in a reach north of the George Washington Bridge (Figure 1) that is relatively uniform in the along channel direction. The channel is laterally asymmetric, with a 15-m deep channel running close to its eastern shore and a shoaling western flank. The mean river discharge (measured at Green Island Dam, near Troy, NY) during May 2002 was 811 m$^3$/s, about 20% above average (Fig. 2). There was a peak flow of around 2,000 m$^3$/s between the two sets of dye releases (shown as horizontal bars on Fig. 2). Tidal range was at its fortnightly minimum (neap) during the first release (Fig. 2), low but increasing during the second release, sharply increasing during the third release, and near its maximum during the fourth release (spring). Stratification was strongest during the first release, and transitioned from highly stratified to nearly mixed during the third release (Fig. 2). The most comprehensive surveys were conducted following the first and third releases, so the main focus of this paper is on them, with some supporting data from the other releases.

The dye studies were complemented with moored instruments that included four 1.2 MHz Acoustic Doppler Current Profiles (ADCP) deployed across the channel approximately 3 km north of the George Washington Bridge (Figure 1). The mooring array also included 12 Seabird CT sensors and 2 Paroscientific pressure sensors. The mooring array was deployed on April 22$^{nd}$ and recovered on June 6$^{th}$. Lerczak et al. (in press) provides a detailed description of this mooring array. We refer to the moorings as sites 1 through 4 counting from west to east.
Each of the dye studies commenced with an injection of ~ 40 kg of Fluorescein dye into the bottom mixed layer in the deep channel on the eastern side of the river. The initial dye streak was several hundred meters long in the cross-channel direction, running between the 10 m isobaths. Each release took approximately 15 minutes. Following the release, the dye patch was surveyed by two boats. One boat ran along-channel sections and the second boat ran cross-channel sections. The cross-channel surveys were done in “tow-yo” mode (i.e., continuous vertical profiling while underway at approximately 3 knots) and provided ~50 m spatial resolution of the cross-channel structure of the dye. Typically each patch survey involved 4-6 cross-channel sections. The along-channel surveys were completed with approximately 10-15 profiles and thus the spatial resolution of the along-channel surveys decreased with increasing patch size. Patch surveys took approximately 1 hour to complete. The injection site was selected so that during the first tidal cycle after injection the dye patch would pass through the mooring array while still remaining north of the more complicated channel morphology south of the George Washington Bridge (Chant and Wilson, 1997) at river km 18. Each boat was equipped with a 1200 kHz RDI ADCP and an Ocean Sensor CTD (OS-200) that was mated with a Chelsea Aquatracka III fluorometer for dye tracking. Both the CTD and fluorometer operated at approximately 6 Hz. One day prior to each dye injection we conducted a CTD survey of the estuary to the inland limit of salt.

Over the course of the 4 dye experiments we completed 29 dye-patch surveys. While the mean inventory of these patches was only 25.2 kg, and significantly less than the amount injected, the standard deviation was only 3.9 kg. We suspect the reduced mean inventory may be due to fluorometer calibration. Nevertheless, the patch-to-patch
consistency in dye inventory allows us to evaluate the changing properties of the dye patch.

The variations of current and salinity through the course of the dye study are shown in Fig. 3. Near-bottom currents (1.5 m above bottom) were typically 0.5 m/s during neaps and up to 0.9 m/s during springs. Surface currents showed less spring-neap variation, sometimes reaching 1.5 m/s on strong ebbs during both springs and neaps. As noted above, salinity stratification was strong during the first two releases, and transitioned from strongly stratified to mixed conditions during the third release.

3. Results

Neap Conditions

The May 5th injection occurred during early ebb. The injection was about 2-m above the bottom in the boundary layer, at a target salinity of 17.1 psu. Surface-to-bottom salinity difference exceeded 15 psu at the time of injection. Along-estuary sections through the patch are shown in Figure 4 and vertical profiles in Figure 5. The first panel in each figure corresponds to maximum ebb, then early, mid- and late-flood conditions on the day of the injection. During maximum ebb the bulk of the patch resided near the bottom, although there was some penetration into the lower halocline. The highest concentrations of dye patch remained at the bottom within a few km of injection site (km 25).

During the flood (Figures 4 and 5, panels b-d), the dye patch thickened as the bottom boundary layer grew. At the beginning of the flood, the patch was about 2.5 m
thick, and it increased to 6.5-m by the end of the flood. The upper limit of the dye patch corresponded closely to the location of the velocity maximum, which indicates the upper limit of the influence of bottom stress on momentum.

The average salinity within the dye patch decreased through the flood, indicating the vertical entrainment of lower-salinity water into the bottom boundary layer. The rate of freshening was found to be consistent with one-dimensional vertical mixing, as will be demonstrated in Section 4.1.2. The salinity structure did not show a distinctive change at the top of the boundary layer (defined here by the upper limit of the dye)—there was a continuous increase in salinity gradient, reaching a local maximum near the upper limit of the dye (Fig 5., left panels). Clearly the boundary layer was not a mixed layer, but rather a stratified shear layer with vertically varying gradients.

The gradient Richardson number (Ri) reached values close to 0.25 within the dye patch during the strong ebb as well as during the strong flood, although its average value within the boundary layer was around 0.5 (Fig. 5, first and 3rd panels). During decreasing flow conditions, Ri increased and became variable.

The lateral distribution of dye (Fig. 6) indicates that the dye was confined to the deep, eastern part of the channel during the ebb, but it spread across the channel during the flood. Once the dye spread to occupy the channel, the lateral variations of dye were modest, roughly corresponding to lateral variations in salinity. There was more lateral variability at the landward and seaward limits of the patch, reflecting differential advection.
The May 9th release occurred during flood, when the dye was injected into the bottom boundary layer at a target salinity of 18 psu. Stratification and near-bottom tidal current velocities during this experiment were similar to those during the May 5th release (Figure 3), although there was an increase in surface currents during flood at the diurnal period. The May 9th experiment provided detailed estimates of mixing during the ebb and complements the May 5th experiment that better resolved the flood. Together these two experiments reveal that during neap tide the boundary layer, as defined by the properties of the core of the dye patch, freshens more rapidly during the flood tide than on ebb (Figure 7a). Details of defining the core of the dye patch are given in section 4.1.1.

Neap-to-Spring Transition

The May 22nd injection occurred just prior to a rapid decrease in stratification approaching spring tides. The dye was injected into the bottom boundary layer at a mean salinity of 12.2 psu at slack before ebb. Like the first injection the rate of freshening during the ebb was weaker than during the flood. Furthermore, the vertical distribution of dye showed multiple peaks that indicate advective as well as diffusive processes (Figure 8a). The highest concentrations were not near the bottom, which can only be explained by vertical advection and/or secondary circulation. As the ebb progressed a broad mixing layer developed over the bottom 6 to 8m, evident both in the dye distribution and the salinity profile. This indicates that boundary layer mixing began to dominate over vertical advection in controlling the vertical distribution of the dye.

During flood, the vertical extent of the dye patch was clearly limited by the velocity-maximum (Fig. 8b). The velocity, salinity and dye distributions within the
boundary layer were similar to the neap case, but the boundary layer grew significantly more rapidly. The dye patch continued to freshen during the flood and mixed both horizontally and vertically through isohalines. At the beginning of flood the near bottom salinity at the core of the dye patch was at 11 psu, while by the end of flood the near bottom salinity of in the core had decreased to 9 psu. Again this indicates the influence of entrainment of overlying lower-salinity fluid. The Richardson numbers in the boundary layer were consistently at or below 0.25 during the flood, except near the velocity maximum where they increased to around 1.

A May 24th injection occurred during spring tides. Dye was injected near the bottom at a target salinity of 7 psu and top to bottom salinity stratification was weak and less than 2 psu (Fig. 3). The spring tide currents in the presence of weak stratification rapidly mixed the dye throughout the water column. Consequently, this data was not suitable to characterize boundary layer growth, because the boundary layer occupied the entire water column for most of the experiment.

4. Discussion

4.1 Mixing on Flood

4.1.1 Bottom Boundary Layer Growth

The following analysis quantifies the extent that entrainment, rather than vertical advection, drives the vertical growth rate of the dye patch. This analysis required estimates of the growth and freshening rate of the dye. This was done for both the properties within the core of the dye patch and with patch-mean properties. Results from
these two approaches reached similar results and therefore we only present the analysis on the relatively homogenous core of the dye patch. The core of each patch was defined by selecting only profiles with maximum dye concentrations greater than $e^{-1}$ times the maximum dye concentration in the patch. For each of these profiles the top of the dye patch was defined as the depth where dye concentration was 10% of the maximum dye concentration for each cast. Because of the sharp vertical gradients at the top of the dye patch and the self-similarity of the vertical structure of the dye (see for example figure 12) these estimates of boundary layer height were relatively insensitive to reasonable changes in the 10% criteria.

During flood, there was a period of nearly steady boundary layer growth. The mean boundary layer height of the core of the patch grew at a rate of $1.9 \times 10^{-4}$ m/s during the neap tide experiment (May 5th) and $3.8 \times 10^{-4}$ m/s during the stronger tidal conditions on May 22nd (Fig 9a). Boundary layer growth, however, could be due to horizontal convergence and/or vertical entrainment. In order to determine the relative role of flow convergence to entrainment the volume-integrated salt within the core of the dye patch were evaluated. Mixing causes its volume to increase monotonically with time, and an entrainment velocity can be defined which represents the normal velocity of that volume (which is typically oriented vertically). The conservation of volume of the dye patch can be written as

$$\frac{dV}{dt} = A_p w_e \quad (1)$$

where $V$ is the volume of fluid that contains dye, $A_p$ is the surface area of that concentration “surface”, and $w_e$ is the area-averaged outward velocity of that surface.
relative to the local water, or the “entrainment velocity” of water into the dye patch. The quotes are used because the term “entrainment” is used in many different contexts. What is meant here is that in a reference frame that moves with the diffusing dye on a constant isopleth, there is a net fluid motion toward higher dye concentration. The more conventional use of the term “entrainment” applies to the conditions during the flooding tide, in which there is an abrupt change in turbulent intensity at the top of the boundary layer, corresponding to a sharp drop in dye concentration. If the patch of dye touches the bottom everywhere, then the surface across which the entrainment occurs is only the upper surface, and it is approximately equal to the horizontal projected area of the dye patch $A_h$. These are the conditions, particularly related to the flood tide conditions, that will be considered in this analysis. The average thickness of the dye patch is defined simply by $\bar{h} = V/A_h$, and the time rate of change of the height is

$$\frac{d\bar{h}}{dt} = \frac{1}{A_h} \frac{dV}{dt} - \frac{V}{A_h^2} \frac{dA_h}{dt}$$ (2)

The first term on the right-hand-side is the entrainment term, and the second is the change in patch thickness associated with horizontal convergence or divergence of the patch. Equation 2 can be rewritten as

$$\frac{d\bar{h}}{dt} = w_e + w_a$$ (3)

where again, $w_e$ is the entrainment velocity, and $w_a$ is vertical advection. Note that $w_e$ must be positive, as it is associated with the irreversible spreading of the dye patch, whereas $w_a$ can be of either sign, depending on the geometry of the flow. Only entrainment, $w_e$, will cause the dye patch and boundary layer to freshen. Therefore the
extent to which the boundary layer growth was driven by entrainment, rather than flow convergence, was determined by a comparison between the freshening rate of the core of the dye patch and $\frac{d\bar{h}}{dt}$. The average salinity within the core of the patch can be related back to the entrainment velocity via the equation

$$
\frac{d\bar{s}}{dt} = -\frac{(\bar{s} - s_h)}{h} w_e
$$

(4)

where $\frac{d\bar{s}}{dt}$ is the time-rate of change of salinity following the dye, and $s_h$ is the average salinity at the top of the boundary layer, which is being entrained into the patch. Values for the quantities in Eq. 4 are listed in table 1.

The freshening rate of the core of the dye on May 5th required an entrainment velocity of $2.3 \times 10^{-4}$ m/s, consistent with the observed growth rate of $1.9 \times 10^{-4}$ m/s. Similarly the requisite entrainment rate on May 22nd, based on the freshening rate that day, is $4.1 \times 10^{-4}$ m/s is roughly consistent with the observed growth rate on flood of $3.8 \times 10^{-4}$ m/s. Consequently we conclude that the growth of the bottom boundary layer is almost entirely due to entrainment and has little contribution from a vertically advective process.

Estimates of the patch-mean thickness growth rate and freshening rate were also made. While both growth and freshening rates are about 20-30% slower than those obtained by the analysis of the patch’s core, the results are similar in that they also indicate that boundary layer growth is primarily driven by entrainment. The moderate differences between the properties of the core and the patch mean properties may be due
to patch-scale variability in mixing and redistribution by three dimensional circulation.

However, these processes appear to be secondary to the flood tide boundary layer
dynamics. We acknowledge that this analysis is 2-dimensional and neglects lateral
variability in channel depth. The impact of lateral variations in depth on boundary layer
processes are discussed after section 4.1.2.

4.1.2 Comparison with Trowbridge (1992) model

Next we compare these growth rates with a boundary layer model developed by
Trowbridge, (1992) that is an extension of earlier models developed by Kantha et al.,
(1977) and Price et al., (1986). The model assumes that the Richardson number in the
boundary layer is held at the critical value of 0.25 by a balance between shear forcing
associated with a surficial stress and buoyancy supplied due to entrainment of pycnocline
fluid into the boundary layer. The observations reported here are approximately
consistent with the assumption that Ri=0.25 in the boundary layer. The growth of the
boundary layer is written as:

$$\frac{\partial \bar{h}}{\partial t} = \frac{Ri_c^{1/2} \gamma}{N_\infty} \frac{u^*}{h} \frac{u^*}{h}$$  \hspace{1cm} (5)

where \(Ri_c = 0.25\) is the critical Richardson number, \(N_\infty\), is the buoyancy frequency of
the pycnocline, \(\bar{h}\) is boundary layer thickness in the thalweg, \(u^*\) is the friction velocity
and \(\gamma = 1.22\) is a constant of integration from Trowbridge, (1992). We estimate bottom
friction using the quadratic drag law \(\tau_b = C_d \rho \nu |\nu|\). Based on direct estimates of stress
and near bottom velocities Trowbridge et al., (1999) found \(C_d = 0.0017\) in the Hudson
applied for currents at \(d=3.5\)m above the bed, which corresponds to a \(C_d = 0.0021\) for our
near bottom velocity 1.35 meters above the bed. Time series of friction velocity for sites 2, 3 and 4 following the first and third dye injections are shown in figure 10. Site 4 is deeper than site 3, which results in stronger baroclinic forcing at site 4, and thus relatively weaker stress during the ebb. However during the flood bottom stresses are approximately equal. For the purpose of determining the intensity of mixing during the flood we used the stress estimate from mooring 4. Note that at site 4 the stress was much weaker during the ebb than the flood during neap tides, but it became more symmetrical during spring tide conditions. This variation in bottom stress is an important contributor to tidal asymmetry, as will be discussed below. We note that the biggest uncertainty in this comparison our estimates of $u^*$ which we feel are likely to be accurate to within 20 percent.

Other variables needed for comparison are the thickness of the boundary layer and the stratification at the top of the boundary layer. The stratification at the top of the layer is taken to be the mean stratification at the top of the core of the dye patch and the thickness of the boundary layer is taken as the mean elevation of the top of the core at mid flood, when the estimate of $\frac{\partial h}{\partial t}$ is well defined.

In general the model agrees well with the observed growth of the bottom boundary layer (table 1). During neap tides the model predicts a growth rate of $1.7 \times 10^{-4}$ m/s and the observed growth rate is $1.9 \times 10^{-4}$ m/s while during the May 22nd experiment both the observed growth rate and the modeled growth rate is $3.8 \times 10^{-4}$ m/s. The agreement between the model results and observations provides support for the application of the Trowbridge, (1992) entrainment model to these results and further emphasizes the dominance of 1-D processes in driving bottom boundary layer evolution.
Moreover, these results imply that vertical mixing is primarily accomplished though a
entrainment processes.

While the above comparison suggests that the Trowbridge model works locally, in reality boundary layer growth will be modified by the lateral variation in bathymetry. As the boundary layer grows vertically, the bottom layer also grows laterally due to the sloping channel. This lateral divergence tends to reduce growth rate because a portion of the increased volume associated with entrainment spreads laterally onto the flank. However, according to eq. 5, the boundary layer growth rate increases with decreasing boundary layer thickness on the flank and thus tends to compensate for the former effect. The fact that the boundary layer in the thalweg appears to grow as fast as 5 predicts indicates that the mean entrainment over the boundary layer exceeds that predicted by 5, and this is consistent with more rapid entrainment on the flank where \( \bar{h} \) decreases.

**4.1.3 The influence of tidal straining on flood**

Tidal straining will also contribute to boundary layer growth. To assess the relative importance of tidal straining to entrainment in driving boundary layer growth we considered an idealized boundary layer of thickness \( \delta \) and a constant vertical shear with currents zero at the bottom and increasing to \( \Delta u \) at \( \delta \) (Figure 11). The boundary layer extends across a mixed bottom layer of thickness \( \delta_M \) and penetrates into a linearly stratified layer of thickness \( d\delta \) above (Figure 11). In this conceptual model the boundary layer advects upstream at \( \Delta u/2 \) while the stratified top of the boundary layer advances at \( \Delta u \). Salinity at the top of the boundary layer is \( S_M - \frac{\partial S}{\partial z} d\delta \), where \( S_M \) is the salinity in \( \delta_M \) (or \( S_2 \) in the case of figure 11). At the top of the boundary layer tidal straining
drives $\frac{\partial s}{\partial t}$ at a rate $\frac{\Delta u}{2} \frac{\partial s}{\partial x}$ faster than the mean value of $\frac{\partial s}{\partial t}$ in the boundary layer.

Therefore the salinity at the top of the boundary will decrease to $S_{M}$ in time

$$\frac{dt}{dt} = 2 \frac{\partial s}{\partial z} d \delta \left( \frac{\Delta u}{\partial s} \right)^{-1}$$

causing the mixed region to grow by $d\delta_{x}$. Therefore the rate that the boundary layer grows due to tidal straining is:

$$\frac{d\delta_{x}}{dt} = \frac{\Delta u}{2} \frac{\partial s}{\partial z}$$

Combining equations 5 and 6 we can write an expression for boundary layer growth that includes both entrainment and tidal straining by noting that the Richardson number in the boundary layer is about $\frac{1}{4}$ and thus $\Delta u = N_{BL} \delta / \sqrt{Ri}$.

$$\frac{\partial h}{\partial t} = \left( w_{i} + \frac{\partial s}{\partial t} \right) = w_{i} \left( 1 + \frac{\varepsilon^{2} N_{BL}}{2 \gamma N_{w} Ri} \right)$$

where $N_{BL}$ is the buoyancy frequency in the “mixed” portion of the boundary layer,

$$Ri_{i} = \frac{g \frac{\partial \rho}{\partial x} h^{2}}{\rho \frac{\partial x}{u_{i}^{2}}}$$

is the horizontal Richardson number (Stacey and Ralston, 2005; Monismith et al. 2002) $h$ is the total water depth and $\varepsilon$ is the fraction of the water column occupied by the bottom boundary layer. The second term on the right hand side of 7 is the relative contribution of tidal straining to boundary layer growth during flood. (Later
in the paper, based on the use of a structure function in the boundary layer, we find that \( N_{\infty} = \sqrt{3} N_{\text{BL}}, \) which for \( \varepsilon = 0.5 \) indicates that this term is approximately equal to 0.2 \( \text{Ri}_x \). On the May 5\textsuperscript{th} experiment, the tidal straining term was approximately 0.3 suggesting that tidal straining had a modest contribution to boundary layer growth, while during the May 22\textsuperscript{nd} flood it was less than 0.1 and tidal straining was of minimal importance.

In general, however, the close agreement between the entrainment model and the observations suggests that during flood tidal straining plays a secondary role in bottom boundary layer growth particularly during the May 22\textsuperscript{nd} experiment.

### 4.1.4 Quantifying the vertical structure of mixing rate during flood

Analysis presented in the preceding sections suggest that the bottom boundary layer growth rate and freshening rate are to first order a one-dimensional process. While we recognize that tidal straining, lateral circulation and other multi-dimensional processes in general influence the boundary layer dynamics, based on the results above we neglect them in the following analysis.

The analysis involves the use of a structure function to model the vertical structure of the dye. The use of a structure function was necessary to avoid problems associated with noise which was amplified by taking vertical gradients. By taking advantage of the analytic properties of the structure function we can estimate the mixing rate and vertical eddy viscosity. The time-dependent, vertical distribution of dye is approximated by the similarity form
\[ c(z,t) = c_o(t)f \left( \frac{z}{\overline{\delta}(t)} \right) \]  

where \( c_o \) is the time-varying concentration at the base of the profile, \( f \) is a shape function (to be determined from the observations), and \( \overline{\delta} \) is the time-varying scale height of the profile and proportional to the boundary layer thickness. In our Lagrangian reference frame (following the core of the dye patch) a conservation equation is written

\[
\frac{dc}{dt} = -\frac{\partial}{\partial z} \left( \overline{c'w} \right) \tag{9}
\]

where the primes indicate turbulent quantities and \( c \) and \( w \) are dye concentration and vertical velocity. Integrating vertically over the boundary layer height and substituting in (8) we obtained

\[
\overline{c'w} = -\delta \frac{dc_o}{dt} \int_0^\zeta f(\zeta')d\zeta' + c_o \frac{d\delta}{dt} \int_0^\zeta \zeta' df \tag{10}
\]

where \( \zeta = z/\delta \) and \( \zeta' \) is a variable of integration. Integrating the second term on the right in parts we obtain

\[
\overline{c'w} = -\frac{d}{dt} (c_o\delta) \int_0^\zeta f(\zeta')d\zeta' + c_o \frac{d\delta}{dt} \zeta f(\zeta) \tag{11}
\]

In the one-dimensional model first term on the right is a conservation statement of dye and is identically zero and therefore the vertical flux of dye is simply related to the rate of growth of the layer, and the shape function The form of the eddy diffusivity
\( K_z = -e^z w' / (\dot{c} c / \partial z) \) can be easily derived from this simplification of (11)

\[
K_z = -\frac{\partial}{\partial t} \frac{d\partial}{\partial \zeta} \zeta \frac{f(\zeta)}{f'(\zeta)}
\] (12)

where \( f' \) is the vertical derivative of the shape function. Finally, substituting 5 into 12 we arrive an expression that relates the vertical eddy viscosity in the boundary layer to the friction velocity and overlying stratification

\[
K_z = -\frac{\text{Ri}^{\frac{1}{2}} u_z^2}{N_x} \frac{f(\zeta)}{f'(\zeta)} = -\chi \frac{u_z^2}{N_x} G(\zeta)
\] (13)

where \( \chi \) is a constant and equal 0.61 and \( G(\zeta) = \zeta \frac{f(\zeta)}{f'(\zeta)} \).

The function \( f = \tanh(\alpha(1 - \zeta)) \) was found to be a good representation of the vertical structure of the dye (Fig. 12) with \( \alpha = 3 \). The value of \( \alpha^{-1} \) represents the fraction of the boundary layer that has strong vertical gradients. Consequently, the vertical structure of the eddy viscosity in the boundary layer is:

\[
K_z = \chi \frac{u_z^2}{N_x} \frac{\tanh(\alpha(1 - \zeta))}{\alpha \sech^2(\alpha(1 - \zeta))}
\] (14)

The vertical structure of the vertical flux of dye and the eddy diffusivity are also plotted. The greatest vertical flux occurs in the upper part of the boundary layer, but the flux abruptly goes to zero at the top of the layer. The eddy diffusivity is maximal close to the bottom, where the stratification is weakest. The diffusivity decreases markedly through the layer and goes to zero at the top of the layer.
Using estimated boundary-layer growth rates during the 1\textsuperscript{st} and 3\textsuperscript{rd} dye releases, the dimensional values of diffusivity were calculated and presented in the bottom of table 1. Two values are reported: the maximum value, which occurs near the bottom of the boundary layer, and the value at the level of maximum dye flux (“upper”), which provides a characteristic value for the region of significant buoyancy flux. These values are 3.7 cm\(^2\)/s and 13.2 cm\(^2\)/s for the 1\textsuperscript{st} and 3\textsuperscript{rd} releases, respectively. The maximum values are approximately 5 times higher, but they have little consequence for mixing due to the weak stratification near the bottom.

We found that the eddy diffusivity at the top of the boundary layer was relatively insensitive to the choice of the structure function. However the distribution of $K_v$ in the lower water column was very sensitive, as it depends inversely on the gradient of structure function as it approaches the boundary. A sensitivity analysis indicated that $K_v$ is accurately determined by this method (to within $\pm 20\%$) in the upper 2/3\textsuperscript{rd}s of the boundary layer, but it is essentially indeterminate by this method in the bottom third. While the choice of the structure function impacts estimates of the maximum eddy diffusivity it has little impact on estimates of vertical fluxes.

Finally, the above analysis can be used to characterize the tidal monthly variations in the vertical buoyancy flux during the flood tide. Buoyancy flux at the top of the boundary layer is equal to $K_v N^2_{\infty}$ which after substituting equation 14 is found to be proportional to $u^2_{*} N_{\infty}$. With data from mooring site 4 we estimated the quantity $u^2_{*} N_{\infty}$ as a proxy for vertical salt flux. Over the spring/neap cycle friction velocity is maximum during spring tide and minimum during neap tide (Fig. 13). In contrast, stratification is
maximum during neap tides and minimum during spring tide. Consequently, the vertical salt flux (or buoyancy flux) reaches a maximum during the transition from neap to spring on days 130 and day 142. Following these two peaks in vertical salt flux there is a rapid decline in stratification, and despite the increase in friction velocity (and eddy viscosity) salt flux during the flood drops because of the weakened stratification. Enhanced vertical salt flux following the second spring tide is due to persistent stratification associated with the large discharge event one week earlier (Fig. 2).

4.2 Mechanisms influencing ebb mixing

The mixing of dye during the ebb was not the simple boundary layer entrainment process that was observed during the flood. Localized convergence of near-bottom flow caused upward advection of the dye (e.g., fig. 4, 1st panel), and produced local maxima in dye that was separated from the boundary layer (e.g., fig. 8a). As the dye patch became separated from the bottom, mixing would not necessarily result in a reduction of the salinity of the patch, but would spread the dye in salinity space. The combination of boundary layer entrainment and mixing of elevated portions of the patch caused both freshening of the patch and spreading in salinity space. During the neap tide experiments, the mixing was weak, and upward advection often displaced maximum dye concentrations from the bottom boundary layer. This was most evident during the ebb as seen in Fig. 8a. During the transitional experiment advective processes also displaced the peak dye concentrations off the bottom. However, as the ebb velocity intensified, a bottom boundary layer developed, within which Richardson numbers were generally below ¼ (Figure 8a).
The contrast in mixing dynamics between the flood and the ebb appears to be due to two main factors: 1) the bottom stress was weaker during the ebb, causing relatively less intense boundary-layer mixing; and 2) the influence of tidal straining competed with mixing during the ebb, reducing its effectiveness in producing a bottom mixed layer. Using the scaling of Stacey and Ralston (2005) for the influence of tidal straining on boundary layer growth, the height of the boundary layer should be limited by the relation

\[ \delta \leq \sqrt{R_{fc}} \frac{u_*}{N_x} \]  

(15)

where \( R_{fc} \approx 0.15 \) is the critical flux Richardson number and \( N_x = \left( \frac{g \partial \rho}{\rho \partial x} \right)^{0.5} \) is a measure of the horizontal density gradient, with a value of approximately 0.002 s\(^{-1}\) during this time in the Hudson. During the early and late ebb, when \( u_* < 1 \text{ cm/s} \), tidal straining limits the vertical extent of mixing to less than 2 m, and the overlying water column should increase in stratification due to the baroclinically induced shear. These observations suggest that the boundary layer was not as strongly limited by straining as the Stacey and Ralston relation would indicate; during maximum ebb (\( u_* = 1.5 \text{ cm/s} \)), the observed boundary layer thickness reached 6-8 m (Fig. 8a), compared to a predicted height of 3-m. Nevertheless, the influence of tidal straining was evident in the tendency of the water column to restratify during the early and late ebb.

The occurrence of elevated maxima in dye can be explained in part by tidal straining. As the water column restratifies, patches of dye can become isolated from the bottom boundary layer. However this does not explain elevated maxima that occur soon after the injection, prior to significant vertical mixing. It appears that horizontal convergence processes are more likely to occur during the ebb than the flood, leading to
vertical advective motions that carry the dye into the interior. Classical estuarine theory indicates that there should be a net vertical velocity in a partially mixed estuary (Chatwin, 1976); these observations suggest that the vertical velocity is modulated within the tidal cycle, occurring preferentially during the ebb. The dynamical origins of this process are beyond the scope of this paper.

4.3 Vertical Buoyancy Flux and the Flux Richardson Number

In this final section we present estimates of the vertical turbulent buoyancy fluxes ($B_f$), shear production ($P$) and their ratio ($P/B_f$) the flux Richardson number ($R_f$). Shear production estimates were made with the moored ACDP data, while buoyancy flux estimates are based on the analysis presented in section 4.1.4. However, since this analysis is only relevant during flood we limit our estimates of the vertical salt flux (and buoyancy flux) and the flux Richardson number to the flood. The vertical structure of salt in the bottom boundary layer during flood is similar to that of the dye and thus we assumed a salt profile in the boundary layer as $s(z) = s_h + (s_0 - s_h) \tanh(\alpha(1-\zeta))$, where $s_h$ is the salinity at the top of the boundary layer and $s_0$ is the bottom salinity. Using equation 4 to relate $\overline{dh/dt}$ to $\overline{ds/dt}$ and equation 11 to relate $\overline{dh/dt}$ to $\overline{|w's|^2}$, where the brackets represent the mean over the boundary layer, we obtain

$$\overline{|w's|^2} = \overline{h} \overline{\frac{ds}{dt}} \frac{\alpha \tanh(\alpha(1-\zeta))}{\tanh(\alpha(1-\zeta))} \approx 0.4 \overline{h} \overline{\frac{ds}{dt}}.$$ (16)
With a linearized equation of state $\rho = (1000 + \beta S)$, we estimate the depth averaged buoyancy flux over the boundary layer as

$$|B_f| = 0.4g \frac{\beta \bar{S}}{\bar{\rho} dt} \bar{h}$$

(17)

where $\beta = 0.77 \text{ kg/(m}^3\text{psu})$. With the dye data in figures 7 and 9 we estimated the term $\bar{h} \frac{d \bar{S}}{dt}$ between each successive surveys of the patch. Here, $\bar{h}$ is the mean thickness of the core of the dye patch across two successive surveys and $d \bar{S}/dt$ is estimated based on the change in the salinity of the core of the patch between surveys. With these estimates we used 16 to determine the mean buoyancy flux over the boundary layer.

Mean boundary layer shear production is estimated by assuming that stress decreases linearly from $\rho u_*^2$ at the bottom to zero at the top of the boundary layer, and that vertical shear is the maximum velocity of the mid-depth velocity jet $u_\infty$ divided by the boundary layer depth defined by the height of the jet, i.e.

$$|P| = 0.5u_*^2 \frac{u_\infty}{\bar{h}}$$

(18)

While these assumptions differ from the dynamics of a constant stress layer, observations presented here and in (Geyer et al., 2000) suggest that both the stress and velocity profile
are approximately linear in the boundary layer. Bottom friction velocity is estimated as it was done in section 4.1.2.

Results from this analysis are presented in figure 14. In figures 14a and 14b the buoyancy flux estimates are multiplied by 5 so that for a flux Richardson number of 0.2 these buoyancy flux estimates are on the same scale as the shear production. On May 5th shear production during peak flood was $2 \times 10^{-5} \text{ m}^2/\text{s}^3$ and approximately five times larger than the buoyancy flux. Data from May 22nd shows a similar relationship with peak shear production around $5 \times 10^{-5} \text{ m}^2/\text{s}^3$ and buoyancy flux at $10^{-5} \text{ m}^2/\text{s}^3$. A scatter plot between all estimates of shear production and buoyancy flux (Figure 14c) has an $r^2=0.64$ and regression of 0.14. The regression represents an estimate of the flux Richardson number. Given the standard error of the fit and 90% confidence limits our analysis places the flux Richardson number between 0.1 – 0.18. Since the maximum value that the flux Richardson number can take in stratified turbulence is 0.2 (Tennekes and Lumley, 1982) these estimates of the flux Richardson number suggests that shear production in the boundary layer is efficient in driving a vertical buoyancy flux.

Using our expression for the vertical eddy diffusivity (14) and the structure function for the density profile in the boundary layer, $\rho = \rho_h + \Delta \rho \tanh(\alpha(1-\zeta))$, the model predicts a buoyancy flux ($B_f$)

$$B_f(\zeta) = \frac{g}{\rho} k \frac{\partial \rho}{\partial z} = \frac{N_w^2}{N_z} u_z^2 \zeta \tanh(\alpha(1-\zeta))$$  (19)
where \( N_{\text{BL}} = \frac{g \Delta \rho}{\rho h} \) and \( \Delta \rho \) are the buoyancy frequency and density difference across the boundary layer respectively. Note that \( N_\infty \) occurs at the top of the boundary layer where the density gradient is maximum, and with the use of the structure function we find that \( N_\infty^2 = \alpha N_{\text{BL}}^2 \). Thus averaging 19 over the boundary layer we find:

\[
|B_f| = 0.063 N_\infty u^2 = 0.11 N_{\text{BL}} u^2
\]

(20)

Using the values in table 1 for \( N_\infty \) and \( u_\infty \), 19 predicts buoyancy fluxes of \( 1.3 \times 10^{-6} \text{ m}^2/\text{s}^3 \) and \( 2.0 \times 10^{-6} \text{ m}^2/\text{s}^3 \) for May 5th and May 22nd respectively. Using the maximum values for \( u_\infty \) on those days based on figure 10 (2 cm/s for May 5th and 3 cm/s for May 6th) yields buoyancy fluxes of \( 2.8 \times 10^{-6} \text{ m}^2/\text{s}^3 \) and \( 4.3 \times 10^{-6} \text{ m}^2/\text{s}^3 \) for those two days. These values fall within the range of measured values as shown in figure 14c. Finally, equating 20 with 18 times \( R_i \), both with represent expressions for the boundary layer mean buoyancy, we find

\[
.22 \frac{N_{\text{BL}} h}{u_\infty} = .22 \sqrt{Ri_{\text{BL}}} = R_i
\]

(20)

Richardson number estimates made in the boundary layer on May 5th and May 22nd are approximately 0.5 and 0.25 respectively (figures 5 and 8) and correspond to a flux Richardson number, according to 20, of 0.16 and 0.11. Both of the estimates, made with equation 20, fall within the error bars of the flux Richardson number estimate that we made with the experimental data (figure 14c) and suggests that the coefficients that we derived for the boundary layer model are essentially correct.
5. Conclusions

A series of dye studies conducted in the Hudson River Estuary during May 2002 elucidated bottom boundary layer mixing processes, their rates and temporal variability over tidal and fortnightly timescales. Results emphasize that diapycnal mixing occurs through entrainment generated by bottom boundary layer turbulence. During neap tides there is a strong tidal asymmetry in salt flux, with stronger fluxes during the flood. Approaching the spring/neap transition this asymmetry is reduced. During flood the bottom boundary layer is clearly defined by the dye patch which mixes up to the top of the boundary layer and by the depth of a sub-surface velocity maximum. The growth of the boundary layer during flood is primarily due to entrainment with minimal contribution to flow convergence or tidal straining. Observed boundary layer growth rates during flood are consistent with a 1-D entrainment model (Trowbridge, 1992). We presented a 1-D model that provides an expression for vertical structure of eddy viscosity in the boundary layer which is proportional to \( \frac{u^2}{N_\infty} \) and that the mean buoyancy flux over the boundary layer is equal to \( 0.11N_{bm}u^2 \). The 1-D model was used to estimate vertical turbulent salt flux over the neap/spring cycle and suggests that maximum vertical salt flux occurs as tidal range is increasing mid-way between neap and spring tide conditions. Estimates of the flux Richardson number in the boundary layer made both with the model and the observations agree and lie between 0.1-0.18. Finally, tidal straining limits boundary layer height during ebb as suggested by Stacey and Ralston (2005).
Acknowledgments: The work was supported by the National Science Foundation grant OCE00-95972. (WRG, JL), OCE00-99310 (RWH) and OCE00-95913 (RJC, EH). We thank two anonymous reviewers for their highly constructive comments that helped to significantly improve earlier versions of this paper. We thank John Lipscomb and the Hudson River Riverkeeper for their tireless and enthusiastic efforts in the field program and John Trowbridge for providing critical insight in the boundary layer analysis. The paper is contribution number XX, XX and XX from Rutgers, Woods Hole and Lamont-Doherty Earth Observatory.
References


### Table 1

#### Parameters used to estimate boundary layer growth rate

<table>
<thead>
<tr>
<th>Parameter</th>
<th>May 5&lt;sup&gt;th&lt;/sup&gt;</th>
<th>May 22&lt;sup&gt;nd&lt;/sup&gt;</th>
</tr>
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<tbody>
<tr>
<td>( \frac{\partial h}{\partial t} ) (observed)</td>
<td>0.19 mm/s</td>
<td>0.38 mm/s</td>
</tr>
<tr>
<td>( \bar{s} )</td>
<td>2.1 ( 10^{-4} )</td>
<td>2.1 ( 10^{-4} ) psu/s</td>
</tr>
<tr>
<td>( \Delta s = \bar{s} - s_h )</td>
<td>4.5 psu</td>
<td>4 psu</td>
</tr>
<tr>
<td>( \bar{h} )</td>
<td>5.0 m</td>
<td>8.0 m</td>
</tr>
<tr>
<td>( N_\infty )</td>
<td>0.12 s&lt;sup&gt;-1&lt;/sup&gt;</td>
<td>0.08 s&lt;sup&gt;-1&lt;/sup&gt;</td>
</tr>
<tr>
<td>( w_e ) (equation 4)</td>
<td>0.19 mm/s</td>
<td>0.41 mm/s</td>
</tr>
<tr>
<td>( w_e ) (equation 5)</td>
<td>0.17 mm/s</td>
<td>0.38 mm/s</td>
</tr>
<tr>
<td>( u_{flood} )</td>
<td>1.3 cm/s</td>
<td>2.0 cm/s</td>
</tr>
<tr>
<td>( u_{ebb} )</td>
<td>0.7 cm/s</td>
<td>1.5 cm/s</td>
</tr>
<tr>
<td>( K_z ) (upper)</td>
<td>3.7 cm&lt;sup&gt;2&lt;/sup&gt;/s</td>
<td>13.2 cm&lt;sup&gt;2&lt;/sup&gt;/s</td>
</tr>
<tr>
<td>( K_z ) (max)</td>
<td>18 cm&lt;sup&gt;2&lt;/sup&gt;/s</td>
<td>62 cm&lt;sup&gt;2&lt;/sup&gt;/s</td>
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**Figure 1** Study site. In small scale map the black dots show mooring sites. Grey dots show locations of dye releases. release 1 and 3 occurred at slack before ebb on May 5th and May 22nd respectively. Release 2 and 4 occurred at during early flood on May 9th and May 24th. Dashed lines are the 10m isobath, and closed solid lines are 15m isobath. Upper right insert depicts channel cross-section at mooring array and locations of moorings.

**Figure 2** A) Thick line shows mean daily discharge based on 56 year record at Green Island. Thin line shows discharge at Green Island in 2002. Grey line depicts duration of mooring deployment and black dashes depict the duration of each dye experiment. B) Tidal range at the Battery (thick line) and subtidal surface to bottom salinity difference from mooring 4 (thin line). Thick black dashes depict duration of each dye experiment.

**Figure 3** Upper panel) Near surface (green) and near bottom (blue) currents from mooring site 4. Flooding current is positive. Lower panel. Near surface (green) and near bottom (blue) salinity from mooring site 4. See figure 1 for locations of moorings.

**Figure 4**. Along channel sections across dye patch on May 5th Salinity (red contour), velocity profile (thick black line), and dye concentration (filled contours). The dye is contoured at 5, 10,20,50 100,200,500,800 parts per 10-11. The thicker salinity contours are the 5,10 and 15 psu isohalines. Velocity profiles are shown for every other CTD/dye profile. Dashed line is zero velocity and vector depicts near surface velocity. Panels are taken during max ebb, early flood peak flood and late flood. Times are in EDT. The injection occurred at km 25.

**Figure 5** Profiles of salinity, dye concentration, along-channel velocity and Richardson numbers during first tidal cycle following the May 5th (neap tide) injection. Profiles are obtained from ship-board observations. Flooding current is positive. Horizontal dashed lines depict the top of the boundary layer. Vertical dashed lines in velocity profiles is at 0 m/s. Vertical dashed line in Richardson number plots is at 0.25.

**Figure 6** Cross-channel sections across center of dye patch during the flood tide on May 5th. Color is dye concentration and salinity is red contour.

**Figure 7** Mean salinity of core of dye patch during the neap tide experiments (A) and during transitional experiment (B). Line is best fit to ebb and flooding freshening rate. For the neap tide experiments plus signs are for the May 5th data and dots are for May 9th data.
**Figure 8a** Profiles of salinity, dye concentration, along-channel velocity and Richardson numbers during ebb following the May 22nd (transitional) injection. Profiles are obtained from ship-board observations. Flooding current is positive. Vertical dashed lines in velocity profiles are at 0m/s. Vertical dashed line in Richardson number plots are at 0.25.

**Figure 8b** Profiles of salinity, dye concentration, along-channel velocity and Richardson numbers during flood following the May 22nd (transitional) injection. Profiles are obtained from ship-board observations. Flooding current is positive. Horizontal dashed lines depict the top of the boundary layer. Vertical dashed lines in velocity profiles are at 0m/s. Vertical dashed line in Richardson number plots are at 0.25.

**Figure 9.** Height of bottom boundary during the May 5th neap tide experiment (plus signs) and during the May 22nd transitional experiment (dots). Line is best fit to flood tide boundary layer height during flood.

**Figure 10.** Estimates of bottom friction velocity during the first (upper panel) and third (lower panel) Dye experiments. Thick solid line is for site 4, dashed line for site 3 and thin line for site 2.

**Figure 11** Conceptual model of tidal straining. Thin solid lines show 4 isohalines and dashed line depicts velocity profile. Velocity is 0 at the bottom, increases linearly to Δu at δ and is constant above. The boundary layer is mixed over height δ_M and stratified at higher elevation at across layer of thickness dδ_s. Vertical shear advects point b upstream Δu/2 faster than point a and causes boundary layer height to increase by dδ_s when it catches up to point a.

**Figure 12** Left panel: Vertical structure of dye observed (thin colored lines) and hyperbolic tangent. Right panel vertical structure of dye, eddy diffusivity and vertical flux of dye based on analysis presented in section 5.1.3.

**Figure 13** upper panel: Bottom friction velocity during flood tide from site 4. Middle panel: Surface to bottom density difference from site 4. Lower panel: Friction velocity squared times stratification, which is proportional to vertical salt flux (see on analysis in section 4.1.3). Grey bars indicate time of each dye experiment.

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