Estimating effective longitudinal dispersion in the Chesapeake Bay

Jay A. Austin

Center for Coastal Physical Oceanography, Old Dominion University, Norfolk, VA 23529, USA

Received 3 April 2003; accepted 20 January 2004

Abstract

An analysis of Environmental Protection Agency’s Chesapeake Bay Program hydrographic dataset shows that the bay responds coherently to variability in freshwater flux. Mean salinity and salinity stratification both respond to variability in freshwater flux on time scales of roughly 90 days. Stratification is also influenced by local wind forcing but on much shorter (4–5 day) time scales. The volume of available data allows the effective longitudinal dispersion coefficient to be estimated as a function of either time or space. Values for this dispersion coefficient vary between 200 and 1000 m² s⁻¹, with mean values around 650 m² s⁻¹. The spatially dependent structure has a maximum roughly 75 km from the head of the estuary, and decreases gradually towards the mouth. The temporally varying effective dispersion varies spatially as the inverse of the estuarine cross-section, and temporally as the cube root of the freshwater flux, and is at least qualitatively consistent with models of estuarine circulation and results of previous field studies. Estimates of the numerical values of the dispersion are useful for better understanding distributions of other tracers within the bay, as well as providing another metric against which numerical models should be measured.

© 2004 Elsevier Ltd. All rights reserved.

Keywords: Chesapeake Bay; estuary; dispersion; mixing; stratification; salinity

1. Introduction

There are a number of processes that determine the spatial distribution and seasonal variability of salinity (and other tracers) in estuaries. These processes are often simplified by casting the problem as a simple, one-dimensional advection–diffusion problem (Taylor, 1954; Fischer, 1976; Chatwin and Allen, 1985). The dispersive portion of the solution is an amalgamation of many disparate processes (Jay et al., 1997), which can effectively be modeled as a Fickian process; i.e. the flux of a property (in this case, salinity) due to dispersion is proportional to the gradient of the property. The magnitude of the proportionality between these two is of great interest as it determines the time scale upon which hydrographic properties such as salinity as well as pollutants, nutrients, and biota are transported upstream within estuaries.

Numerical values of dispersion on large scales in estuaries are relatively difficult to determine and interpret, for a number of different reasons. First, different processes are likely responsible for dispersion at different spatial and temporal scales (Geyer and Signell, 1992). Previous studies have typically studied this dispersion at much smaller spatial and temporal scales than those characterized by whole estuaries over long (i.e. seasonal) time scales of variation. For instance, studies which have used the mixing of dye tracer patches to estimate dispersion (i.e. Robert Chant, Rutgers University, unpublished results) estimate horizontal mixing on spatial scales small compared to the entire estuary, and on temporal scales small compared to seasonal variation. Second, moored observations of salinity over long time scales are relatively uncommon in estuaries due to the propensity for conductivity sensors to foul quickly due to biological activity. Third, repeated hydrographic surveys over time scales long compared to the characteristic time scale of mixing would provide a dataset capable of addressing this, but this entails significant
effort and expense, and are relatively uncommon. Finally, while horizontal dispersion is a primary agent for determining property distributions within estuaries, it is an intrinsically noisy process whose influence is felt only over long time scales, and hence large volumes of data, taken over spans of time long compared to the characteristic time scales of the estuary, are required to make statistically significant estimates of the mixing coefficients. For these reasons, there are very few opportunities to make estimates of the dispersion in large estuaries such as the Chesapeake Bay. An exception to the last two of these difficulties is the Environmental Protection Agency’s Chesapeake Bay Program hydrographic dataset (Environmental Protection Agency, 1993).

The Environmental Protection Agency’s Chesapeake Bay Program (EPA CBP) has been collecting hydrographic data at fixed locations around the entire Chesapeake Bay and its tributaries since 1985. This dataset provides a unique opportunity to examine seasonal and estuary-wide scale variation in the hydrography not possible using surveys of lesser spatial or temporal extent. The frequency of surveys (roughly every three weeks) limits the study to time scales considerably longer than the interval between surveys. The characteristic time scale of adjustment of the bay, set by the mean ocean–bay exchange, appears to be on the order of 90 days (Austin, 2002) so the EPA CBP data provide us with an opportunity to look at the large-scale dispersion within the bay. The goal of this paper is to determine a range of values for the effective horizontal dispersion in the bay, which is ultimately responsible for salt (and other tracers) reaching the upper portions of the bay. In addition, the dispersion is shown to depend, albeit weakly, on the instantaneous rate of freshwater flux, a dependence that has been observed in past studies (Paulson, 1970; Garvine et al., 1992).

The analysis outlined here is descriptive, and does not address the specific mechanisms responsible for this dispersion, in that the observations lack velocity measurements, the data used here are not appropriate for the sort of detailed analysis which would be necessary to discern the mechanisms behind the dispersion. However, a better quantitative knowledge of the dispersive characteristics of a particular estuary will be useful for developing and testing hypotheses about various mixing mechanisms. This is also the only example known to the author of dispersion estimates for the Chesapeake Bay, and one of the very few examples of consideration of either temporal or spatial variability of dispersion within an estuary, largely due to the large volumes of data necessary to make those sorts of estimates. A knowledge of the numerical magnitude of the mixing, plus some indication of the spatial and temporal variability of the dispersion may be useful for developing a better understanding of biological and chemical distributions within the bay. Of course, this sort of simple model is only applicable to conservative tracers, and models of biochemical fields such as nutrients or oxygen would require source and sink terms which may be on the same order as the dispersion. Perhaps most importantly, the effective dispersion and exchange rate, as well as the dependence of the dispersion on the freshwater flux, can be viewed as intrinsic properties of an estuary, and therefore represent at least as important a benchmark against which numerical models should be measured as, for instance, instantaneous or time-averaged distributions of salinity.

The paper will continue as follows: Section 2 is a description of the data, Section 3 discusses the salinity structure and variability within the bay, and uses this salinity data to make estimates of the effective longitudinal dispersion the bay. This is followed by a short summary.

2. Data

The Environmental Protection Agency’s (EPA) Chesapeake Bay Program (CBP) (Environmental Protection Agency, 1993), (http://www.chesapeakebay.net/data/index.htm) has collected a large amount of data in the Chesapeake Bay and its numerous tributaries starting in 1984 (the data used for this paper start at the beginning of 1985). These data include temperature and salinity along with a host of biological and chemical parameters. The data are used mainly to quantify distributions of nutrients and other biological parameters in the bay (e.g. Boynton et al., 1995; Dauer and Alden, 1995), though the hydrographic data have been used in some studies as well (Gibson and Najjar, 2000; Austin, 2002). The 37 distinct stations in the main stem of the bay are roughly eveny distributed around the bay (Fig. 1A). As of early 2002, this dataset consisted of over 11,000 individual hydrographic casts. For most of the analysis in this paper, a subset of 21 of these stations which lies along the center of the bay will be used, as indicated in Fig. 1A by stars. Of these data, there are 235 instances in which salinity was measured at 18 or more of the 21 central stations within 5 days. Each of these represents a relatively synoptic survey of the bay, judging by the 90-day time scale of temporal variation estimated by Austin (2002). The data will be somewhat corrupted by tidal variation, but it is assumed that tidal variability is on scales sufficiently small compared to the total variability of the bay, so that they are averaged over in the course of a survey, which typically take several tidal cycles to complete. Therefore, the tides may add noise, but they do not add a bias. The spring–neap cycle plays an important role in mixing processes in estuaries (e.g. Bowen, 1999), and may be manifested in these surveys, but it is assumed that with a
17-year record they are averaged into the estimates of the salinity distribution, and once again appear only as noise.

The Chesapeake Bay has many rivers supplying it with fresh water, with an average total freshwater flux of approximately 2280 m$^3$ s$^{-1}$. The Susquehanna and the Potomac, with 48% and 16%, respectively, of the total freshwater input (Schubel and Pritchard, 1986), make up nearly two thirds of this input (Fig. 2A). Daily values of the flux from these two rivers are measured by the United States Geological Survey River Gaging program (stations # 01578310, at Conwingo Dam, near the...
mouth of the Susquehanna River, and # 01646500 near Washington, DC; data available at http://water.usgs.gov). A variety of other rivers (in order, the James (13%), Rappahannock (3%), York (3%), and the Patuxent, Choptank, and Nanticoke (1% each)) make up the remainder of the gauged flux; small, ungauged streams contribute 10% and 6% from the western and eastern shore, respectively (Schubel and Pritchard, 1986). The time-varying flow out of the Susquehanna and Potomac are correlated at a level of 0.7 (linear correlation coefficient), and most of the sub-watershed fluxes are highly correlated with each other. The decorrelation time scale of freshwater events (i.e. the lag at which the autocorrelation is 0.5) is around 10 days, so the events are relatively short-lived. This is significant in that it suggests that steady-state models of salinity distribution will not accurately reflect actual salinity distributions, since the time scale of adjustment is much longer than the time scale of the forcing.

Meteorological data are available at several stations around the Chesapeake Bay, courtesy of the National Data Buoy Center (NDBC, http://seaboard.ndbc.noaa.gov). Hourly winds from Thomas Point Light (TPLM2) at the northern end of the bay and from Chesapeake Light (CHLV2) just outside the mouth of the bay are used.

Fig. 2. (A) The total estimated freshwater input to the bay. (B) The integrated freshwater input to the bay integrated with a 90-day relaxation scale, as described in Austin (2002). (C) The temporal mode of the 1st EOF of the vertically-averaged salinity structure of the Bay. (D) The temporal mode of the 1st EOF of the salinity stratification distribution in the Bay.
3. Results

3.1. Salinity distribution and variability

The salinity in the bay, averaged over the top 12 m (Fig. 3A, B) over the time period 1985–2001, shows, as expected, nearly oceanic salinities near the mouth (around 29, just inside the mouth of the bay) and freshwater at the head, where the Susquehanna enters the bay. Laterally, fresher water appears on the western side of the bay and saltier water on the eastern side, consistent with Pritchard’s (1952) analysis of mean bay salinities. This is due largely to rotational effects and presumably also to the predominance of larger watersheds on the western coast of the bay (about 7% of the total freshwater input to the bay is from the Eastern Shore). The distribution of mean salinity as a function of along-axis distance from the head of the bay, using the 21-station subset discussed earlier, is indicated by a solid line in Fig. 3B. This suggests that there are strong horizontal gradients in the top 75 km of the bay, and a gentle but uniform gradient over the next 175 km. The up-estuary extent of the salinity intrusion can move as far south as 75 km from the head of the estuary.

Empirical Orthogonal Functions (EOFs, Emery and Thompson, 1997, Chapter 4.3) are used to ascertain the most significant coherent mode of variation in the bay. The largest coherent component of variation (Fig. 3C, D, the first EOF) explains 65% of the total variation in vertically averaged salinity. It is highly spatially coherent, with roughly constant value throughout the bay, except near the head and the mouth, where it is somewhat smaller. This is a consequence of the fact that the salinity at the head of the estuary is almost always zero, and the salinity at the mouth is always close to oceanic values. The second EOF (not shown) represents a mode of variation where the head and the mouth of the bay are negatively correlated. This mode

---

*Fig. 3. (A and B) The mean distribution of salinity in the Chesapeake Bay over the 17-year period of the EPA surveys. (A) contours; (B) along-estuary dependence. The ‘x’s represent the mean values along the main channel of the bay; the dashed lines represent the mean plus and minus the product of the first spatial EOF and the standard deviation of the first temporal mode EOF. (C and D) The spatial distribution of the first EOF. ‘x’s represent the first EOF along main channel.*
only accounts for 5% of the total variance and will not be studied. The rest of the modes are by definition even smaller.

In Austin (2002), the mean bay salinity was shown to be a function of river input and bay—ocean exchange, so that the spatial mean salinity could be estimated by:

\[ S_{\text{mean}}(t) = S_a + \left(1 - \int_{-\infty}^{t'} e^{\frac{(Q + F/V)(t' - t)}{\tau}} F(t') \, dt' \right), \]  

(1)

where \( S_a \) is the ambient ocean salinity, \( F(t) \) is the freshwater influx, \( Q \) the mean ocean—bay exchange rate, \( F \) the mean freshwater flux, and \( V \) the volume of the bay. The quantity \( V(Q + F)^{-1} \) represents the ocean—bay exchange time, determined in Austin (2002) to be approximately 90 days. The integral portion \( \int_{-\infty}^{t'} e^{\frac{(Q + F/V)(t' - t)}{\tau}} F(t') \, dt' \), referred to as the freshwater index, represents the recent history of freshwater events, with more recent events having more weight, and is shown in Fig. 2B.

The temporal component of the first EOF (Fig. 2C) is strongly correlated (0.89) with the freshwater index. This is consistent with the fact that this freshwater index is highly correlated with the mean bay salinity. Since the first EOF is uniform in the central portion of the bay, variations in the salinity measured anywhere in the central portion of the bay (away from the mouth or head) are going to be a reasonable proxy for salinity variations on seasonal and longer time scales anywhere in the central bay. By adding or subtracting the first spatial EOF (scaled by the standard deviation of the first temporal mode) to the spatial mean distribution (Fig. 3, dashed lines), it is clear that the primary variation in bay salinity gradients occurs at the head of the bay. During periods following strong freshwater fluxes, the bay salinity drops uniformly over the central portions of the bay. Likewise, during dry periods, the gradient is strongest at the head of the bay, with the gradient over the central regions of the bay being approximately constant.

The salinity difference between 1 m and 10 m is used as a proxy for vertical stratification. The mean stratification (Fig. 4A, B) is nearly constant throughout the bay, with slightly lower stratification at mid-bay. The spatial portion of the first EOF (Fig. 4C, D) is uniform, suggesting that the vertical stratification changes uniformly. The first temporal mode (Fig. 2D) is correlated (0.61), once again, with the freshwater index, as was the variation in mean salinity. High stratification tends to be associated with recent freshwater events. However, the time series (Fig. 2D) is noisy compared to that of the mean salinity (Fig. 2C). This is due to the fact that stratification can be locally influenced on relatively short time scales by wind forcing, while the vertical mean salinity remains unchanged during wind mixing events.

The portion of the stratification that is not explained by the freshwater index is significantly correlated (0.50) with the average wind speed over roughly four days prior to the survey, as measured at Thomas Point Light (NDBC CMAN TPLM2), at the northern end of the bay. The correlation is statistically significant (at the 95% level) but not large. It is worth noting that the wind is measured at the northern extent of the bay, and may not be representative of the winds over the whole of the bay. For instance, the correlation between wind speed measured at Thomas Point light and at Chesapeake Light (NDBC CMAN CHLV2), at the mouth of the bay, is only 0.44. Also, the surveys often took up to five days, and hence are not synoptic relative to the scales of the forcing proposed here. Nevertheless, this is consistent with the results of Goodrich et al. (1987) who demonstrated the relationship of the wind field to the breakdown of vertical stratification, using moored data. In summary, the stratification varies due to freshwater forcing on long (~90 day) time scales, and local wind forcing on short (~4 day) time scales.

### 3.2. The effective horizontal dispersion

The relative abundance of salinity data in the EPA dataset allows crude estimates to be made of the distribution of the effective dispersion \( K_H \) both in space and time. The simple conceptual model (Taylor, 1954; Chatwin and Allen, 1985):

\[
\frac{\partial}{\partial t} (A(x)S(x, t)) + \frac{\partial}{\partial x} (A(x)U(x, t)S(x, t)) = \frac{\partial}{\partial x} \left( A(x)K_H \frac{\partial S}{\partial x} \right),
\]  

(2)

where \( x \) is the along-estuary coordinate with \( x = 0 \) at the head of the estuary, \( A(x) \) is the cross-sectional area of the estuary, and \( U(x, t) \) the estimated down-estuary advection, represents a balance between local salinity fluctuation, advection, and horizontal dispersion. Many variations have been made on this model. A common simplification is to disregard the time-varying term, which in this case can be significant. Banas et al. (submitted for publication) in a study of Willapa Bay, WA, as well as modeling approaches (MacCready, 1999) include a baroclinic dispersion term, which is impossible in this case due to the lack of simultaneous velocity measurements. The bay has strong vertical shear in its mean circulation (Goodrich and Blumberg, 1991), with classical estuarine flow out of the estuary at the surface and into the estuary at the bottom. In this case, dispersion contributions from shear processes are merged into the generic dispersion term.

The advective term is estimated using \( U(x, t) = F(t)A(x)^{-1} \), where \( F(t) \) is the estimated freshwater input. North of the Potomac River, the flux is estimated to be...
1.18 times the measured flow of the Susquehanna, and below the Potomac, 1.6 times the sum of the Potomac and Susquehanna. The scaling factors are included so that the fluxes are approximately representative of the upper and lower bay, respectively. As the measured fluxes out of the large watersheds tend to be highly correlated (i.e. the correlation between the time series of freshwater flux from the Potomac and Susquehanna is 0.7), the results are not particularly sensitive to how the freshwater flux is implemented, as long as the total input of water is roughly correct. The cross-sectional area $A(x)$ was approximated using a truncated bathymetry (taken from various NOAA sources) of the bay (Fig. 1B), which disregards sub-estuaries entering the bay. $A(x)$ was then determined for a given latitude by simply integrating the bathymetry across the truncated bay. Since the temporal scale of the sampling is much longer than typical scales of surface fluctuations, $A(x)$ is assumed to be constant in time.

Fig. 4. (A and B) The mean distribution of salinity stratification in the Chesapeake Bay over the 17-year period of the EPA surveys. (A) contours; (B) along-estuary dependence. 'x's represent the mean values along the main channel of the bay. The dashed lines represent the mean plus and minus the product of the first spatial EOF and the standard deviation of the first temporal mode EOF. (C and D) The spatial distribution of the first EOF. 'x's represent first EOF along main channel.

The dispersive term (the right-hand side of Eq. (2)) represents all of the processes that can transport salinity upstream and downstream, which are modeled as behaving in a Fickian manner. In order to estimate $K_H(x,t)$, Eq. (2) can be integrated from the head of the estuary ($x=0$) to a longitudinal position $x=X$ within the estuary:

$$
\int_0^X A(x) \frac{\partial S}{\partial t} \, dx + F(t)S(x) \big|_{x=X} - F(t)S(x) \big|_{x=0} = A(x)K_H(x,t) \frac{\partial S}{\partial x} \big|_{x=X} - A(x)K_H(x,t) \frac{\partial S}{\partial x} \big|_{x=0}.
$$

(3)

Two boundary conditions at the head of the estuary ($x=0$) can be applied to this:

$$
S(x=0) = 0,
$$

(4)

i.e. the water at the head of the estuary is always fresh, and
\[ K_H S_x(x = 0) = 0, \quad (5) \]

i.e. the diffusive flux of salt through the head of the estuary is zero. This leaves, evaluating (3) at \( x = X \):

\[
\int_0^X A(x) \frac{\partial S}{\partial t} dx + F(t) S(X) = A(x) K_H(X, t) \frac{\partial S}{\partial x}. \quad (6)
\]

This is equivalent to the formulation of Officer (1976), equation (5-99), except for the addition of a salt storage fluctuation term, the first term on the LHS of Eq. (6). In this case, the storage fluctuation term can be significant and will be retained. At this stage, each of the terms save \( K_H(x, t) \) can be evaluated by differencing vertical mean salinity data in either space or time, with care taken that the resulting spatial and temporal derivatives end up on a consistent space–time grid. To evaluate the data, the 21-station subset of stations (Fig. 1A) was taken to approximate the along-axis section of the bay. Vertical averages over the top 12 m were used for the individual values of \( S(x, t) \) for a given location and time. The results are not sensitive to this choice of averaging depths greater than 8 m, and only the four stations nearest the head of the estuary are shallower than 12 m. This was done so that horizontal differences between stations were representative of the horizontal gradients and not unduly influenced by changes in depth between stations. The 12 m isobath is included in Fig. 1A.

In total, 235 surveys fit the criteria that 18 of the 21 stations must be occupied in a 5-day time period. Therefore, the left-hand side of Eq. (6) can be represented as a 234 × 20 matrix, as can the \( \partial S/\partial x \) term on the right-hand side. At this stage, \( K_H(x, t) \) can be determined using a least squares method, either as a function of space or of time, by assuming that \( K_H(x, t) \) is constant in the other dimension. \( K_H(x, t) \) is a function of both but determining the space–time structure of this value in any statistically significant sense, given the noisy, limited, irregularly spaced data is difficult. Two different approaches will be taken: one in which \( K_H(x, t) \) is assumed constant in time to produce a spatially dependent estimate of \( \hat{K}_H(x) \), and one in which \( K_H(x, t) \) is assumed constant in space to produce a temporally dependent estimate \( \hat{K}_H(t) \).

3.2.1. Spatial distribution

A least-squares method was used to estimate the optimal \( \hat{K}_H(x) \) at each location. The distribution (Fig. 5A) is uneven, but there does appear to be an along-estuary increase in \( \hat{K}_H(x) \) in the first 75 km, and a decrease in \( \hat{K}_H(x) \) throughout the rest of the estuary. The structure in the upper bay (i.e. apparently

![Fig. 5. (A) Estimates of \( \hat{K}_H \) as a function of along-estuary position. The Potomac River enters the Bay in the bin at roughly 170 km (in parentheses). (B) Estimates of \( \hat{K}_H \) as a function of the inverse of the cross-sectional area. 'x’s represent first five bins, stars are remaining bins. The Potomac bin is in parentheses.](image-url)
increasing $\tilde{K}_H(x)$ with downstream position) over the first six stations (north of Thomas Point, Fig. 1A) may be due to two artifacts of the analysis: one, the model does not take into account the fact that the upstream extent of the salinity intrusion may move south, frequently as far as the fourth station from the head, and it is not clear that the simple salinity balance outlined by Eq. (6) applies over the region upstream of the intrusion. Second, the first four stations are all shallower than 12 m, which is the depth to which the vertical average is being taken, and may bias the vertical mean salinities in this region. Seaward of the 75 km the dispersion gradually decreases towards the bay mouth. The value of the dispersion is inversely proportional to the cross-sectional area of the estuary (Fig. 5B). This is contrary to the model of MacCready (1999), who postulates that the width of the estuary determines the largest tidally-driven eddy which can form, and hence the magnitude of the dispersion. The fact that the dispersion actually decreases where the estuary is wide suggests that the dispersion is not likely caused by lateral variability, but instead by the intensity of either the area-averaged along-estuary tidal currents or exchange flow, which, by continuity, vary inversely with the cross-sectional area of the estuary. The large value roughly 170 km from the head of the estuary is the grid cell into which the Potomac flows and the freshwater flux applied in this bin may not be representative. The values range between about 300 and 2000 m$^2$/s$^1$, with a mean value of 950 m$^2$/s$^1$. This estimate is consistent in magnitude with other estimates and with other literature references to observations of longitudinal variation of dispersion, though Paulson (1969), in a study of the Delaware Estuary (using far less data), observed no discernible longitudinal variation in $\tilde{K}_H(x)$.  

3.2.2. Temporal distribution

Independent of the spatial analysis, the time-varying structure $\tilde{K}_H(t)$ can be determined through a similar least-squares method. Each estimate represents the spatially-uniform value of $\tilde{K}_H(t)$ that best fits the data from a given survey. Most estimates fall within the range 200–1300 m$^2$/s$^1$ (Fig. 6A). The mean value over all of the data is 650 ± 320 m$^2$/s$^1$. The river flow out of the Susquehanna is weakly correlated with the time series of dispersion, with a correlation of 0.5 (Fig. 6B), significant at the 95% level. (Correlating it with the combined Susquehanna/Potomac flow has similar results.) The solid line in Fig. 6B has slope $\tilde{K}_H \sim F^{1/3}$ for comparison, and is not a fit to the data.

Several papers have speculated on the relationship between river flux and various stratification and exchange parameters. In contrast to the results of Garvine et al. (1992), Oey (1984), and Monismith et al. (2002), no clear relationship was found in the central portion of the bay between the along-estuary gradient and the river flux. However, we do observe an increase in the vertical salinity during periods of high river flux, which may act to make the buoyant exchange more effective.

Without corresponding velocity data, it is impossible to separate the buoyancy-driven exchange from diffusive processes, as is done in, for example, Banas et al. (submitted for publication). It is likely that the dispersive term here is a function of both of these, but without an estimate of the magnitude of the true diffusive process, it is impossible to judge the relative contributions of it and the baroclinic exchange term.

However, the fact that the dispersion increases with river flux is consistent with previous studies which were capable of separating these two terms.

4. Summary

Seventeen years of hydrographic data taken along the main stem of the Chesapeake Bay are used to analyze the mean salinity structure, stratification structure, and infer trends in the effective longitudinal dispersion of...
properties. The mean salinity structure is a strong function of variations in freshwater flux, and the bay responds coherently to episodic inputs of freshwater. Stratification is increased during these inputs, responding on a time scale of roughly 90 days, but is decreased following wind events, on time scales of 4–5 days.

Using these same data and a simplified, integrated form of an advection–diffusion equation, the effective horizontal dispersion is inferred both as a function of space and time. The effective horizontal dispersion appeared to be inversely proportional to the cross-sectional area of the estuary, and a $F^{1/3}$ dependence of dispersion on the freshwater flux was observed at all but the lowest flux rates. Both of these results are consistent with buoyant exchange dominating the up-estuary transport of salt. Although the limited data available make it difficult, if not impossible, to separate this effective dispersion into its components, namely the buoyant exchange and truly diffusive processes, the results outlined in this paper present investigators with a framework within which to work. Long-term moored time series, such as those discussed in Banas et al. (submitted for publication) would be appropriate for this sort of calculation. Perhaps more importantly, these basic estimates of effective dispersion and some information about their spatial and temporal variability will provide a difficult but essential test for numerical models of estuarine circulation.

Acknowledgements

During the writing of this manuscript, the author was sponsored by The Center for Coastal Physical Oceanography, Old Dominion University. Data were provided in the public domain for this work by the Environmental Protection Agency, Chesapeake Bay Program, the United States Geological Survey, and the National Data Buoy Center. Arnoldo Valle-Levinson provided the gridded bathymetry. The author would also like to thank Parker MacCready, Jamie Pringle and Neil Banas for their input, as well as two anonymous reviewers, whose comments have resulted in a substantially improved manuscript over the original.

References


