The Axial Salinity Distribution in the Delaware Estuary and its Weak Response to River Discharge

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We use long term salinity and river discharge data from the Delaware estuary, U.S.A. to determine the mean axial salinity distribution and the salinity response to fresh water discharge. The Delaware is a weakly stratified estuary with a typical vertical salinity variation of only 1 psu. We find that over most of the estuarine salt intrusion length the mean axial distribution of salinity is surprisingly close to a linear decrease with axial distance. Using linear regression analysis, we find that the response of salinity to river discharge is surprisingly weak. The equivalent displacement of a given isohaline for a change in discharge of one standard deviation is only about 4 km, about half the amplitude of the $M_2$ tidal displacement. This implies that some powerful buffering agent exists to reduce the salinity response. We suggest two possible mechanisms for this agent: the action of vertical shear flow dispersion in a tidally stirred regime and the action of lateral shear coupled to strong lateral salinity gradients.

Introduction

Determining the axial salinity distribution in an estuary is a problem of long standing. Variance in this distribution will arise primarily from tidal advection and diffusion, wind-induced circulation, including that forced remotely by wind stress over the adjacent coastal ocean, and fresh water inflow by river discharge. In this paper we focus on the latter effect. In principle, the salinity in an estuary will range from that of the adjacent coastal ocean throughout its length for zero discharge to uniformly zero salinity for sufficiently high discharge. Nevertheless, we are particularly interested in the climatological mean salinity distribution and the response to river discharge within ordinary bounds.

We recently acquired long term salinity data from stations distributed along the axis of the Delaware estuary so as to cover most of the range over which salt intrudes from the coastal ocean. Together with long term river discharge data, these enable us to make good estimates of the mean axial salinity field and of the salinity response to river discharge.
Figure 1. Map of the Delaware estuary. The inset shows the drainage basin of the estuary in black. The large map shows the shoreline of the estuary. The axial coordinate $x$ is shown together with filled circles that indicate the locations of the salinity stations used.

The Delaware estuary occupies the weakly stratified or well mixed class of estuaries. It is a large coastal plain estuary adjoining the Middle Atlantic Bight shelf of the east coast of the U.S.A. The insert map of Figure 1 shows its general location and the land area it drains. In the main part of the figure we show the geometry of the estuary from its mouth at Cape May, New Jersey to its head of tide at Trenton, New Jersey. Here it adjoins the Delaware River that supplies 52% of its mean fresh water inflow (Ketchum, 1953). The length of the estuary is 215 km; its maximum breadth is 45 km and its mean depth 8 m.
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![Figure 2. Contours of salinity (psu) in a vertical section along the estuary axis, June 12 1990. The mouth is at the left where xL^-1 = 0. Triangular symbols at the lower margin indicate where CTD profiles were made. At Trenton x = L = 215 km.](image)

The $M_2$ tide is dominant and at the mouth its height amplitude is about 0.7 m and its volume flux about $1.5 \times 10^5$ m$^3$ s$^{-1}$ (Münchow et al., 1992). The ratio of $M_2$ tidal volume flux to mean total fresh water discharge ($650$ m$^3$ s$^{-1}$) is 226, this large value reflected in the weak vertical density stratification of only about 2 kg m$^{-3}$.

In Figure 1 we also show the axial coordinate $x$ that we use to define points along the estuary. At the mouth we set $x = 0$ and take positive values landward so that at the estuary head at Trenton $x = L = 215$ km. The midpoint, defined by $x/L = 0.5$, is also shown and, as we find below, roughly marks the landward end of the greater part of the observed salt intrusion.

**Axial salinity distribution**

In the Delaware estuary lateral variations in salinity generally exceed the weak vertical variations (Wong et al., 1990), especially in the wide, seaward part of the estuary. But in this paper we discuss only the axial salinity distribution. In Figure 2 we present a typical vertical section of the axial salinity distribution from near the mouth at $x/L = 0$ to $x/L = 0.6$. We obtained the salinity contours shown by interpolating from vertical CTD (conductivity, temperature, depth) profiles we collected where indicated by the triangular symbols on June 12 1990 when the river discharge at Trenton was about 500 m$^3$ s$^{-1}$. The profiling required five hours to complete, so some distortion by tidal advection is present, a maximum of about 9 km in $x$ or 0.04 in $x/L$ at any one station. Nevertheless, the major features we wish to mark are clear: The vertical variations, about 1 psu, are much weaker than the axial ones, about 30 psu, consistent with the classification of the estuary as weakly stratified. At fixed depth the axial variation of salinity $S$ is smooth and monotonic from about 1 psu near $x/L = 0.5$ to about 30 psu near the mouth. The approach of $S$ to 1 psu is abrupt and lacks a long, tapered connection to the nearly fresh water regime landward.

Figure 1 shows the locations of the five long term salinity monitoring stations from which we used data. These stations were the only ones that met our requirements for the sampling rate to be high enough to avoid tidal aliasing, for record lengths of at least several months, and for sampling periods that included both low and high river discharge (less than 100 m$^3$ s$^{-1}$ and greater than 1000 m$^3$ s$^{-1}$, respectively). We obtained salinity samples
TABLE 1. Delaware estuary salinity statistics

<table>
<thead>
<tr>
<th>Site</th>
<th>x/L</th>
<th>Depth, m</th>
<th>Source</th>
<th>Data Days</th>
<th>Mean* Salinity, psu</th>
<th>SD, psu</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.03</td>
<td>20</td>
<td>UD</td>
<td>234</td>
<td>30.9 ± 0.1</td>
<td>0.8</td>
</tr>
<tr>
<td>2</td>
<td>0.11</td>
<td>4</td>
<td>NOS</td>
<td>153</td>
<td>25.1 ± 0.7</td>
<td>2.3</td>
</tr>
<tr>
<td>3</td>
<td>0.25</td>
<td>4</td>
<td>NOS</td>
<td>149</td>
<td>13.5 ± 0.9</td>
<td>2.9</td>
</tr>
<tr>
<td>4</td>
<td>0.27</td>
<td>0</td>
<td>USGS</td>
<td>4848</td>
<td>13.7 ± 0.4</td>
<td>4.4</td>
</tr>
<tr>
<td>5</td>
<td>0.40</td>
<td>0</td>
<td>USGS</td>
<td>4752</td>
<td>4.3 ± 0.3</td>
<td>3.3</td>
</tr>
</tbody>
</table>

*UD = University of Delaware, NOS = National Ocean Service, and USGS = U.S. Geological Survey. Values listed are the mean plus or minus the standard error.

at station 1 near the mouth from an Interocean S4 current meter we maintained at 20 m depth on a mooring line during 1986 and 1987. Conductivity and temperature sensors supplied samples at half-hour intervals from which we computed salinity. The National Ocean Service provided us with similar salinity time series at stations 2 and 3 at 4 m depth using Grundy current meters sampling at 20 minute intervals during 1984. The U.S. Geological Survey supplied daily mean near surface conductivity and temperature values at stations 4 and 5 computed from hourly samples. The latter time series were very long and covered the period from 1969 to 1987, although frequent data gaps were present.

Table 1 lists the x/L coordinate of each station, the sampling depth, and the amount of data in days, excluding data gaps. The five stations provided coverage over most of the salt intrusion length. We made no attempt to extrapolate these records to a common depth because, as Figure 2 shows, there is little vertical salinity variation anyway. Data amounts range from 149 days at station 3 to 4848 days (13 years) at station 4. These amounts were sufficient to estimate mean values with small standard errors, less than 1 psu, compared to the standard deviations (Table 1).

In Figure 3 we plot the mean value $\bar{S}$ for each of the five stations vs. x/L as an open circle with symmetrical brackets about each showing the standard deviation. The dashed line indicates the associated linear regression. This straight line fits the mean value distribution surprisingly well ($R^2 = 0.996$) and is given in psu by

$$\bar{S} = 32.9 - 72.5(x/L).$$

Thus, the mean axial salinity gradient is $d\bar{S}/d(x/L) = -72.5$ psu or $d\bar{S}/dx = -0.337$ psu km$^{-1}$. This simple fit predicts the landward end of the salt intrusion ($\bar{S} = 0$) at $x/L = 0.45$.

For comparison we also show in Figure 3 the salinity interpolated at 4 m depth from the CTD section of Figure 2, shown here as a sequence of filled circles. At a given value of x/L this salinity exceeds $\bar{S}$, given by the regression, nearly everywhere, although the difference is not more than one standard deviation. Furthermore, the scale of the difference between the two lines in Figure 3, in terms of x/L at fixed $S$, corresponds to the scale of the tidal displacement, $\Delta x/L = 0.04$. This interpretation anticipates our central conclusion given in section 3, i.e., axial isohaline displacement only weakly responds to changes in river discharge so that much of the variance in displacement is produced by tidal advection.

Finally, in Figure 3 we show results from the estuary model one of us (McCarthy, 1991) developed and applied to the Delaware estuary. This model is analytical in formulation
The axial salinity distribution in the Delaware Estuary

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Figure 3. Salinity (psu) vs. scaled axial distance xL. The open circles show mean salinity at the five observation stations with brackets that denote the standard deviation. The straight, dashed line is the associated regression line. The filled circles show the salinity at 4 m depth interpolated from the CTD section of Figure 2. The solid line is the salinity from McCarthy's (1991) model.

and requires weak stratification. It includes both a time-dependent barotropic tidal flow and a tidally-averaged flow forced by the joint action of nonlinearities in the tidal flow, river discharge, and the gravitational circulation. The tidally-averaged density field is found as a function of x only and is coupled to the circulation, not prescribed. The density field solution depends on the axial diffusivity Kx, specified as a constant. McCarthy (1991) found that his model required that Kx equal or exceed a minimum value for a given set of model parameters in order to obtain a well behaved density field. In physical terms, sufficient landward diffusive buoyancy flux (proportional to Kx) is needed to balance the net seaward Lagrangian buoyancy transport produced by the river discharge, especially in the more landward reach of an estuary where tidal buoyancy transport becomes weak. The solid line in Figure 3 gives the equivalent tidally-averaged salinity field S(x/L) for minimum possible Kx. Clearly the model overestimates the salt intrusion length, as the salinity values are everywhere significantly higher than the regression line. For larger Kx values the salt intrusion length is longer still. We discuss the axial diffusivity further in the next section where we treat the salinity response to variations in river discharge.

Response to river discharge

The Delaware River discharge into the estuary at the head of tide at Trenton has long been gauged by the U.S. Geological Survey. They supplied us with daily mean discharge values that were continuous throughout the salinity station sampling period from 1969 through 1987. During this period this discharge, which we term R, had a mean value of $\overline{R} = 356$ m$^3$ s$^{-1}$, a standard deviation of 338 m$^3$ s$^{-1}$, a minimum of 54 m$^3$ s$^{-1}$ and a maximum of 3681 m$^3$ s$^{-1}$.

This continuous record permitted us readily to compute lagged correlations and linear regressions between R and S at the five stations, despite gaps in all five of the latter time series, by simply using properly corresponding pairs of values in time whenever a given
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TABLE 2. Lagged correlation between salinity and river discharge

<table>
<thead>
<tr>
<th>Site</th>
<th>S lag at maximum correlation, days</th>
<th>Correlation coefficient ± standard error</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>18</td>
<td>-0.55 ± 0.18</td>
</tr>
<tr>
<td>2</td>
<td>12</td>
<td>-0.48 ± 0.22</td>
</tr>
<tr>
<td>3</td>
<td>7</td>
<td>-0.60 ± 0.25</td>
</tr>
<tr>
<td>4</td>
<td>2</td>
<td>-0.60 ± 0.07</td>
</tr>
<tr>
<td>5</td>
<td>2</td>
<td>-0.60 ± 0.06</td>
</tr>
</tbody>
</table>

Salinity station produced data. In Table 2 we show the results for lagged correlation. The lag time corresponding to maximum correlation decreases in the landward direction, as one would expect, falling from 18 days at station 1 to 2 days at the landward end of the salt intrusion at stations 4 and 5. The correlation coefficients are negative, as expected, and range from about -0.5 to -0.6, significant values when compared to the standard error estimates shown.

These levels of correlation between $S$ and $R$ encouraged us to compute the linear regression properties at each of the five salinity stations. We present these values in Table 3 in the form of the slope $dS/dR$ (psu/$10^3$ m$^3$ s$^{-1}$) for the same lag time as in Table 2, the lag with maximum correlation. The values are all negative so that increased discharge is associated with decreased salinity (at fixed $x$), as expected. The magnitudes of the regression slopes are all large compared to the standard error estimates that follow them in the table, but are much smaller than expected. For example, at station 2 where the mean salinity is 25.1 (Table 1) the regression predicts that a change in $R$ of $1000$ m$^3$ s$^{-1}$ (about three times the standard deviation) corresponds to a change in $S$ of only 2.35 psu.

We can transform the regression results into an equivalent isohaline axial displacement change with $R$ using

$$\frac{dx_s}{dR} = -\frac{dS/dR}{dS/dx}$$

where $x_s$ is the axial coordinate of an isohaline of salinity $S$ and $dS/dx = -0.337$ psu km$^{-1}$, as in section 2. Values of $dx_s/dR$ (km/$10^3$ m$^3$ s$^{-1}$) appear in the third column of Table 3. All are negative, indicating seaward displacement for increasing discharge. To give a clearer sense of the isohaline displacement potential of river discharge we give in the fourth column $\Delta x_R$, the absolute value of the displacement that would result from a change in $R$ of one standard deviation ($338$ m$^3$ s$^{-1}$). These values are surprisingly low, typically only 4 km, or in terms of $\Delta x_R/L$, only about 0.02. For comparison we list in the last column the $M_2$ tidal current particle excursion or displacement amplitude $\Delta x_T$ at each station computed from the local $M_2$ surface current amplitude obtained from McCarthy (1991). The tidal displacement is about 9 km, roughly twice $\Delta x_R$.

Thus, we reach the central point of this paper. Within its ordinary range of river discharge variation the Delaware estuary responds weakly in terms of axial isohaline displacement. In particular, discharge changes of one standard deviation are associated with displacements only about half those produced by tidal advection. This implies the existence of a powerful buffering agent that counters the seaward advection of salt associated with increased river discharge. We discuss this agent next.
Figure 4. The response of isohaline position $x_s L^{-1}$ for the 13-7 psu isohaline vs. $R$, the Delaware River discharge at Trenton. The solid, straight line shows the linear regression from the observations, the dash-dot line the response from McCarthy’s (1991) model for fixed axial diffusion, and the dashed line the response of the same model for minimum diffusion at each value of $R$.

### Table 3. Salinity response to river discharge

<table>
<thead>
<tr>
<th>Site</th>
<th>$dS/dR$, psu/10³ m³ s⁻¹</th>
<th>$dx_s/dR$, km/10³ m³ s⁻¹</th>
<th>$\Delta x_p$, km</th>
<th>$\Delta x_r$, km</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>$-1.54 \pm 0.15$</td>
<td>$-4.6$</td>
<td>1.5</td>
<td>8.5</td>
</tr>
<tr>
<td>2</td>
<td>$-2.35 \pm 0.35$</td>
<td>$-7.0$</td>
<td>2.4</td>
<td>8.5</td>
</tr>
<tr>
<td>3</td>
<td>$-3.89 \pm 0.43$</td>
<td>$-11.5$</td>
<td>3.9</td>
<td>8.8</td>
</tr>
<tr>
<td>4</td>
<td>$-7.69 \pm 0.15$</td>
<td>$-22.8$</td>
<td>7.7</td>
<td>8.8</td>
</tr>
<tr>
<td>5</td>
<td>$-5.69 \pm 0.11$</td>
<td>$-16.9$</td>
<td>5.7</td>
<td>9.4</td>
</tr>
</tbody>
</table>

### Discussion

The agent we suggest is the increase in effective axial diffusivity $K_x$ with increasing river discharge $R$. This agent is suggested by the results we show in Figure 4. There we plot vs. $R$ the axial coordinate $x_s/L$ of the 13-7 isohaline, the same salinity as the mean value at station 4. The solid, straight line shows the observational result from the linear regression slope (Table 3) with $x_s/L = 0.27$ (Table 1) for $R = \overline{R} = 356$ m³ s⁻¹. Again we see the weak response of $x_s/L$ to $R$ variations, here with $R$ ranging from 100 to 900 m³ s⁻¹. In sharp contrast, the dash-dot curve shows the response of the same isohaline from McCarthy’s (1991) model when $K_x$ is held fixed as $R$ varies. For this result $K_x = 2960$ m² s⁻¹, the minimum value possible when $R = 900$ m³ s⁻¹. As $R$ decreases the model predicts that $x_s/L$ increases rapidly until it reaches all the way to the estuary head for $R \approx 200$ m³ s⁻¹. Evidently far too much diffusion is being applied for the lower discharge values so that the landward intrusion of salt becomes progressively much greater than the observations show.
If, instead, the model is used with the minimum possible $K_x$ for each value of $R$, the nearly linear response results for $x_s/L$ shown in Figure 4 by the dashed line. In that case $K_x$ decreases nearly linearly with $R$ from 2960 m$^3$s$^{-1}$ to 2090 m$^3$s$^{-1}$ at $R = 200$ m$^3$s$^{-1}$. The slope of $x_s/L$ vs. $R$ for this line is $-0.183/10^3$ m$^3$s$^{-1}$ compared to $-0.106/10^3$ m$^3$s$^{-1}$ for the slope of the regression line based on observations. Thus, even for this case the isohaline response to discharge changes is too large, but is much closer to the observed behavior than the case with constant $K_x$. Better agreement could have resulted if the model had used spatially variable $K_x$, but this would have made it very difficult to retain analytical solutions.

Galperin and Mellor (1990) applied a three-dimensional, time dependent numerical model to the Delaware estuary that employed a turbulence closure submodel that generated its own effective diffusivities without the need to impose them a priori. Nevertheless, their results for salinity response to discharge changes were much higher than we showed above from observations. Their monthly mean axial isohaline positions changed landward about 50 km between April and October 1984. The monthly mean values of $R$ for those two months were 107 and 960 m$^3$s$^{-1}$, respectively. Thus, their model isohaline response was $dx_s/dR \approx -60$ km/10$^3$ m$^3$s$^{-1}$, about 5 times greater than the average of the observed values we showed in Table 3.

Paulson (1970) employed a simple one-dimensional, steady state model for salinity in the upper Delaware estuary near Philadelphia. He found it critical to increase $K_x$ significantly as $R$ increased in order to simulate salinity observations. There are several plausible mechanisms that may produce the covariance of $K_x$ with $R$. Here we suggest two possibilities: the action of vertical shear flow dispersion in a tidally stirred regime and the action of lateral shear coupled to strong lateral salinity gradients. Linden and Simpson (1986) recently studied the former mechanism through laboratory experiments. They found that increasing the baroclinic flow in a vertically stirred regime produced effectively greater axial diffusivity. Smith (1980) studied the latter mechanism but omitted the effects of Coriolis acceleration. Our data, however, are not useful for evaluating which of these two mechanisms dominates. It seems important to us to establish quantitatively, nonetheless, how the increase of $K_x$ with $R$ is accomplished in order to improve modeling capability for application to weakly stratified estuaries, such as the Delaware, and perhaps to more stratified ones as well.

It is also important to estuarine ecology to understand the general process by which the axial salinity response to river discharge is weakened. A wide variety of estuarine species have limited tolerance to salinity change. Perhaps many species have evolved in part to take advantage of this weakened response.

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References