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Transport of Hypoxic Waters: An Estuary-Subestuary Exchange

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Abstract

Hypoxic or anoxic conditions in the subpycnocline water of Chesapeake Bay persist throughout the summer. The effect on the dissolved oxygen concentration in the deep basin of the lower Rappahannock River, a subestuary on the western side of the bay, was studied with an observational program. The data indicate that in the lower portion of the water column the subtidal (or residual) current was directed into the subestuary most of the time. The mass fluxes of salt and dissolved oxygen into the subestuary through a point near the estuarine bottom at the river mouth were calculated for tidal and subtidal components respectively. From the analyses, we conclude that the mass exchange owing to the tidal component is at least an order of magnitude smaller than that resulting from the subtidal component. Characteristic differences in the properties of water transported into the subestuary were observed. On five occasions, each lasting about two days, during the one month period of field measurements, the imported water was characterized by low dissolved oxygen and high salinity, typically 3 to 4 mg l^{-1} lower dissolved oxygen and 2 to 3 psu higher salinity than at other times. The low dissolved oxygen, high salinity water masses were all accompanied by strong subtidal current and southwest wind. During periods of strong wind from the southwest quadrant, the density-driven current near the bottom was enhanced by the wind-driven circulation. Furthermore, the surface set-up favors the transport of the water from the deep portion of the bay resulting from (1) tilting of the pycnocline in the bay and (2) shoreline and bathymetric configurations around the estuary-subestuary junction.

29.1 Introduction

Problems of hypoxia and anoxia in estuarine and coastal waters have received increased attention in recent years (Schroeder, 1985). The depletion of dissolved oxygen (DO) generally results from a combination of biological and physical factors. It is most frequently observed in subpycnocline waters during summer when vertical mixing is weak and water temperature is high. In Chesapeake Bay, for example, summer anoxia has been observed in the deep

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bottom waters since the 1930s (Newcombe and Horne, 1938), and has become more widespread and of longer duration during recent times (Flemer et al., 1983).

All major subestuaries on the western side of Chesapeake Bay have a deep basin near their mouths (Fig. 1). Hypoxia has been observed frequently in the deep basins in all these subestuaries except in the James River, the southernmost subestuary. Kuo and Neilson (1987) made a comparative investigation of the bottom DO in the three Virginia subestuaries. They reported that hypoxia occurred most frequently in the deep water of the Rappahannock River, less frequently in the York River, and rarely in the James River, even though it received the heaviest wastewater loadings among the three subestuaries. This was attributed to the relatively strong gravitational circulation in the James River.
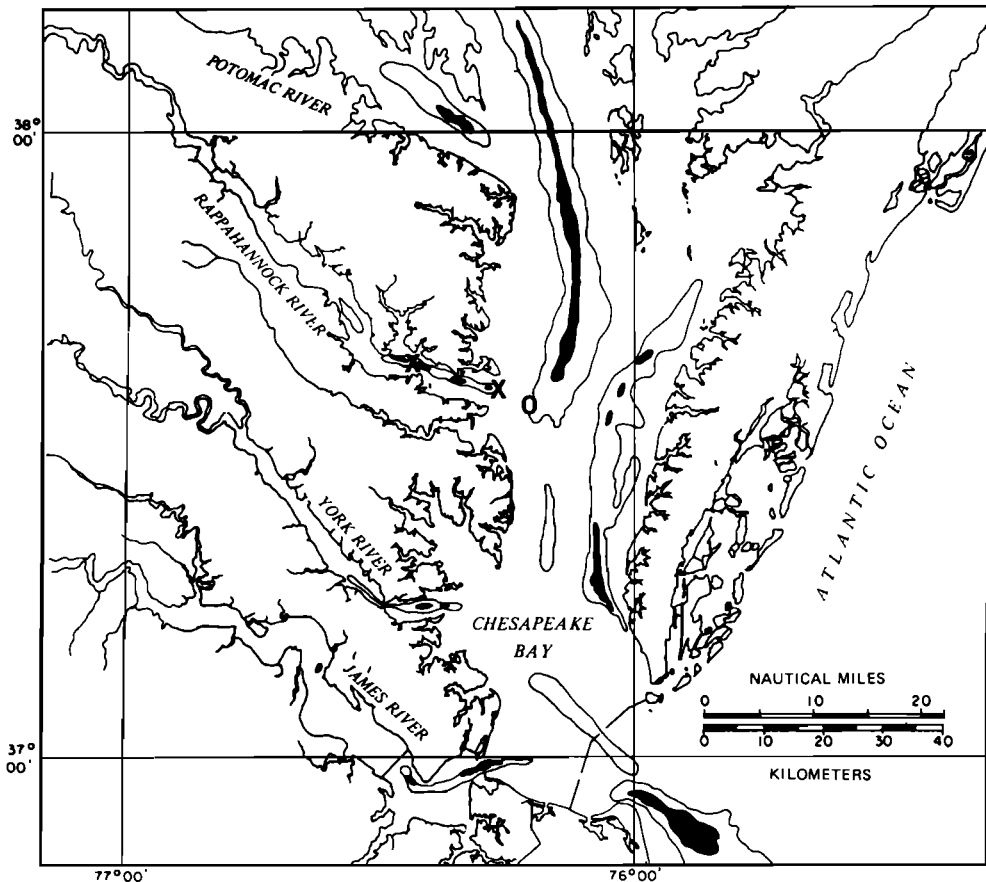

Figure 1:. Lower Chesapeake Bay with its four major subestuaries (The shaded areas are deeper than 20 m, the depth contours are 13 m; **X** marks the station with time series measurements and **O** marks the slackwater survey station in the bay).

Since the variability in hypoxic conditions among the three Virginia subestuaries follows
a similar trend of decreasing bottom DO from south to north in the main stem of Chesapeake
Bay, it is possible that the variability may be due, at least in part, to the quality of the waters
imported into the mouths of the subestuaries from the bay. In this paper, the effect of the
hypoxic or anoxic conditions in the bay on water quality in the lower Rappahannock River is
investigated by examining the mass transport near the bottom at the river mouth. Time series
data from simultaneous measurements of DO, temperature, salinity and velocity at the bottom
of the river mouth were analyzed to demonstrate the primary factors dominating the estuary-
subestuary exchange.

29.2 Field observation

Field data for this study were collected during summer 1987. Data collection can be divided
into two major groups.

29.2.1 Slackwater surveys

A total of 13 slackwater surveys were conducted from June to September at slackwater before
ebb. During each survey, temperature, conductivity and DO were measured at 11 stations
along the river axis and one station off the river mouth in the main stem of the bay. The river
stations extended from the river mouth to 57.8 km upriver beyond the deep basin in the lower
subestuary.

Temperature and conductivity were measured with an Applied Micro System conductivity-
temperature-depth probe (CTD). Continuous vertical profiles, top to bottom, for these
variables were obtained at each station. DO was measured using an oxygen probe (Yellow
Springs Instruments) every meter from the surface to 15 m depth, then every 2 m until the
bottom was reached. Conductivity and temperature measurements were converted to salinity
as psu (UNESCO, 1983). Salinity, temperature and DO data for the river stations were
displayed as isopleths in the vertical-longitudinal plane for each survey.

Two examples of spatial distributions of DO and salinity are presented in Fig. 2. In
general, the surface DO concentrations were always high, with little variability along this
stretch of the subestuary. The DO concentrations in the bottom water were low in all surveys,
with a characteristic longitudinal pattern. The bottom DO concentrations decreased in the
upriver direction from the river mouth, reached a minimum and then increased as water depth
decreased. There was a striking difference in vertical salinity and DO distributions at the river
mouth between the two dates. The water column at the river mouth was moderately stratified
on August 10 and became nearly homogeneous on August 17 (Fig. 2a and 2b). Fig. 2b also
shows an apparent intrusion of high salinity water at mid-depth, thus implying probability of
reversed three layered circulation (Elliott, 1978). This created an unstable density stratification
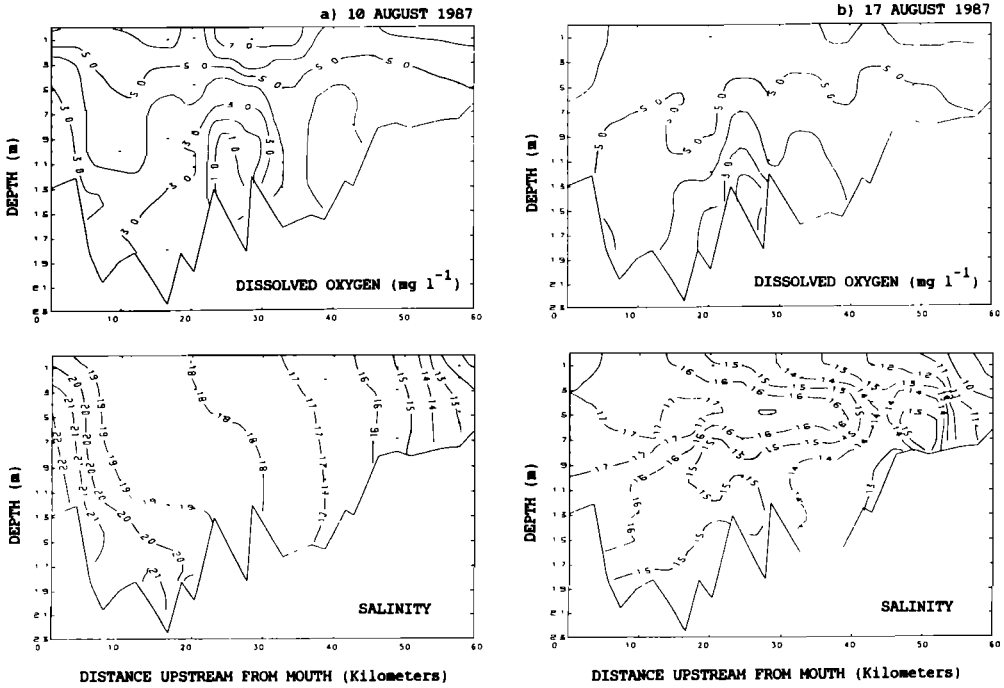in the lower portion of water column, and enhanced the vertical mixing.

Figure 2: Distribution of dissolved oxygen (mg l⁻¹) and salinity (psu) in the lower Rappahannock River. a) August 10, 1987 and b) August 17, 1987 (• marks the measuring location).

US EPA (Environmental Protection Agency) has a Chesapeake Bay monitoring program measuring water quality parameters throughout the bay at semi-monthly intervals. As indicated by the August 3-4, 1987 sampling, the bay was highly stratified and DO in the bottom water was severely depressed (Fig. 3). To demonstrate the time history of the vertical distribution of DO and salinity in the bay, the EPA program data and our slackwater survey data, both at the station off the Rappahannock River mouth (Fig. 1), were combined to construct constant value contours in the depth-time plane (Fig. 4). It shows that the bay was stratified throughout the period. Surface DO was always at or near saturation level while hypoxic conditions in the bottom water started to develop in May and persisted throughout the summer.
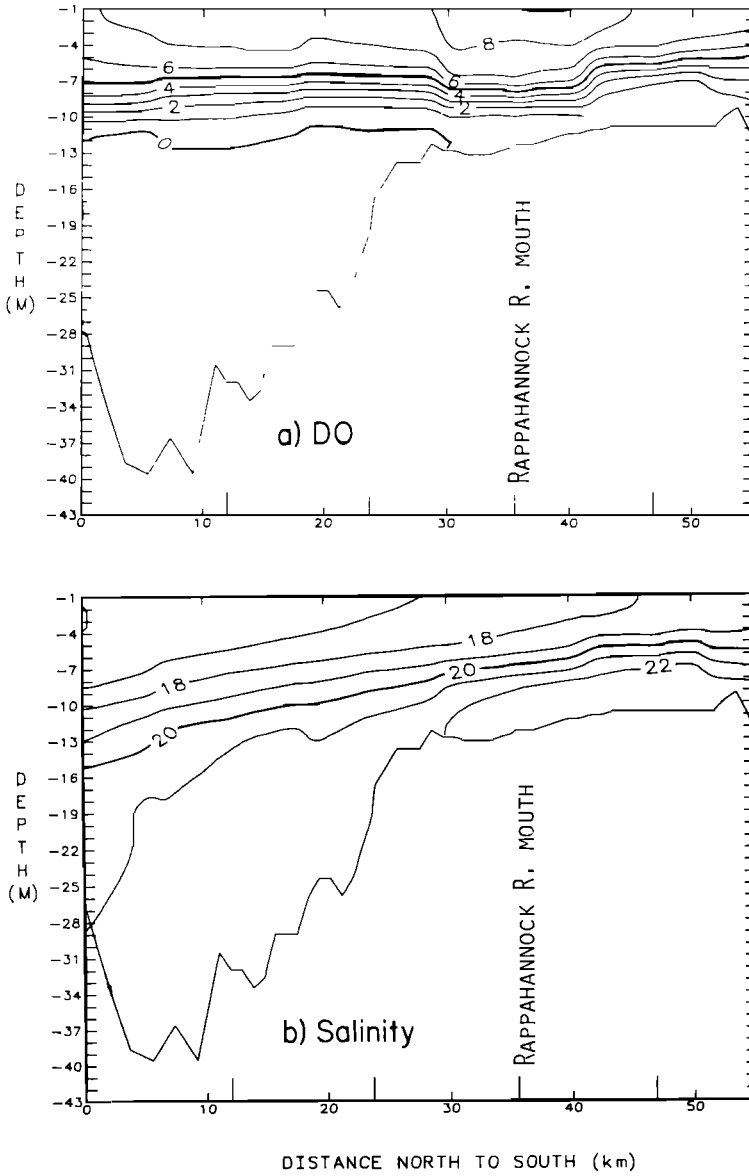
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Figure 3: Dissolved oxygen (mg l^{-1}) and salinity (psu) in the main stem of Chesapeake Bay on August 3-4, 1987 (The larger tick marks on abscissa indicate measuring stations).
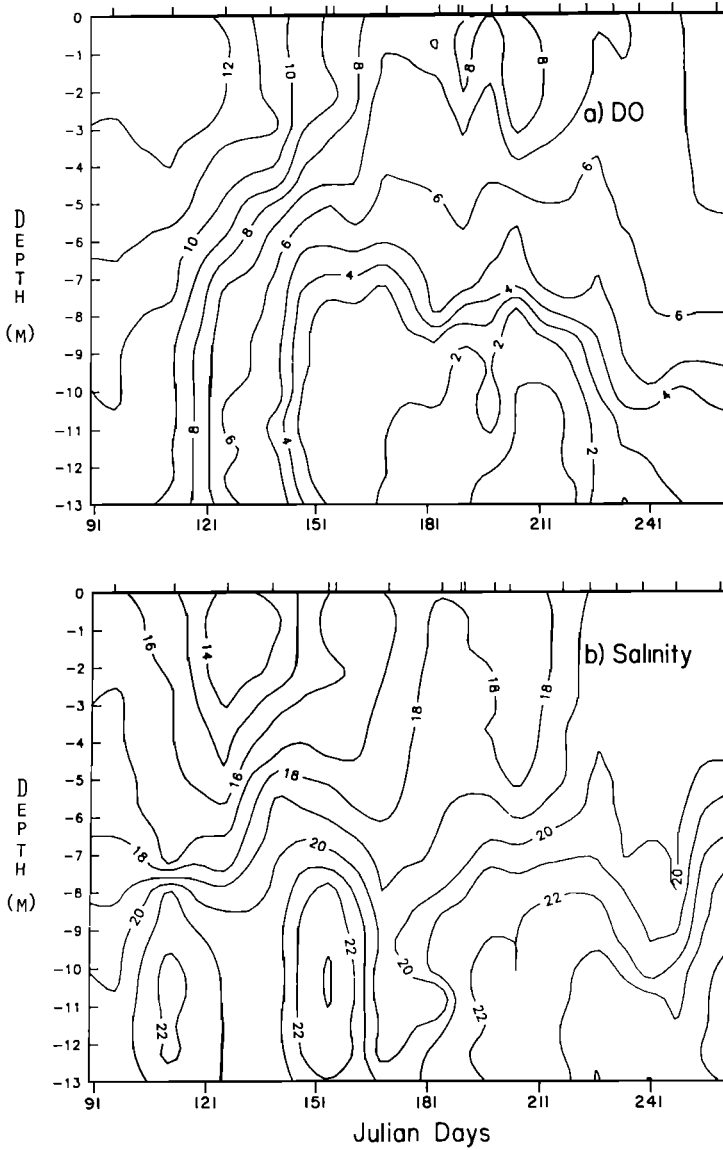
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Figure 4: Dissolved oxygen and salinity isopleths in depth-time plane (The tick marks at the top of the figure indicate sampling dates).

29.2.2 Time series measurements

Time series measurements of temperature, conductivity and dissolved oxygen were taken at two stations, one at the river mouth and the other at 16.6 km upriver in the deep basin (Fig. 1). The data were measured and recorded every half hour with Hydrolab DataSounde 1. One sensor package was deployed at the river mouth near the bottom at 11.8 m depth, while at the upriver station two were deployed, one at mid-depth (7.5 m) and one at the bottom (16.2 m). The meter at the river mouth station was deployed from 5 August to 2 September. During this period, it was repeatedly serviced, calibrated and redeployed for periods ranging from 7 to 10 days. Battery life and fouling restricted the deployment period, thus creating gaps in the time series data. The meters at the upriver station were not deployed until 25 August because of availability.

Current velocities were measured with in-situ, self recording meters that were deployed with taut wire moorings at the river mouth station and the station 16.6 km upriver. Two types of current meters were used; Inter-Ocean S4 meters and modified Braincon Histogram meters. The S4 meter is an electromagnetic type current meter with solid state memory, some of which are also equipped with temperature, conductivity and pressure sensors. The meters were set to record the average values of all variables every 30 min. The Braincon meter measures current magnitude by a savonius rotor, with a vane attached for direction measurement. All three Braincon meters deployed functioned properly for a period of only a few days and then succumbed to electronic problems or fouling. This leaves available data for two depths at the river mouth station, and for three depths at the upriver station.

The velocity data from each current meter were analyzed to determine the principal axis of the flow at the meter location. The direction of the principal axis is defined as the direction along which the sum of the absolute values of velocity components from all data points is maximum. This direction was taken as the longitudinal axis of a coordinate system for further current analyses.

Three tide gauges were installed during this study at locations spanning the length of the subestuary. Water surface elevations were recorded every 6 min. These data were then converted to hourly surface elevations.

Wind data from Norfolk and Richmond airports were examined. Norfolk is located 65 km to the south and Richmond is located 80 km to the west of the Rappahannock River mouth. Both sets of wind data exhibit similar wind direction and magnitude except during periods of weak wind with variable direction. The wind data in Norfolk is presented in Fig. 5a and used for later analyses. Fig. 5 also displays the time series data of water surface elevation, longitudinal velocity components, salinity and DO at the river mouth station.
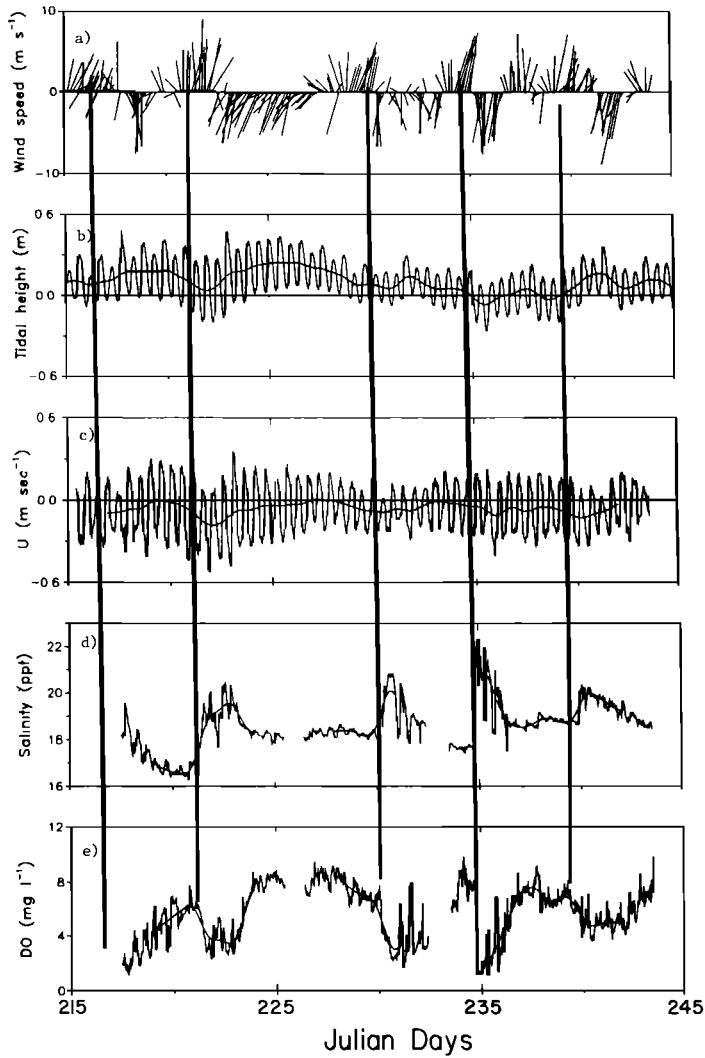
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Figure 5: Time series data. a) Wind data at Norfolk Airport (y-axis points to the north), b) Filtered and unfiltered surface elevation at river mouth and c-e) Filtered and unfiltered time series data near the bottom at the river mouth (positive velocity is in ebb direction). The vertical lines indicate the events of high salinity, low DO water intrusion.

29.3 Data analyses and mass flux calculation

All time series data were subjected to a low-pass filter with a cut-off frequency of $(36 \text{ hr})^{-1}$. The filter was a modification of the low-pass filter designed by Godin (1972). The filtering process essentially eliminates the diurnal and semi-diurnal tidal constituents and fluctuations of higher frequencies. The resulting filtered series were considered as subtidal components. The difference between the measured time series and a subtidal component is considered as the tidal component for the following discussion. That is

$$q(t) = <q>(t) + Q(t)$$

where

q = total component, or the measured time series data,
<q> = subtidal component, or the filtered time series data,
Q = tidal component,
t = time.

Fig. 5 also presents the subtidal components of surface elevation, longitudinal current, salinity and DO. The subtidal current (or Eulerian residual current) was directed into the subestuary along the bottom at all times (Fig. 5c). The average value over the period of measurement is -6.3 cm s^{-1}. The average values of the longitudinal velocity components at the upriver station were 2.9, -3.0 and -3.7 cm s^{-1} at depths of 1.2, 10.0 and 18.7 m, respectively. This agrees with the classical estuarine circulation of seaward flow in the upper layer and landward flow in the lower layer.

Since the hypoxic conditions exist only at the lower portion of the water column, both in the bay and subestuary, the influence of the bay on the subestuary may be investigated by considering the mass transport through the lower portion of the water volume at the subestuary mouth. Quantitative estimates of mass transport should be obtained by spatial integration of the product of velocity and concentration over the portion of the cross-section through which the transport is landward.

$$M(t) = -\iint_A u(y,z,t) \, c(y,z,t) \, dy \, dz$$

where

M = mass transport rate into the subestuary,
u = longitudinal velocity component (positive is out of subestuary),
c = concentration of dissolved substance,
A = the portion of the cross-section in which the transport is
   into the subestuary,
y, z = the spatial coordinates in the cross-section.

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In this study, there are measurements of velocity, salinity and DO at only one point in the lower layer of the transect at the river mouth. No information on the spatial variability of mass transport may be extracted. However, the temporal variability may be investigated, at least qualitatively. The mass flux, or the mass transport per unit area, at the current meter location may be calculated as:

$$m(t) = - u(t)c(t)$$

where the concentration c may be salinity or DO (or dissolved oxygen deficit). If it is assumed that spatial distributions of u and c over the area A do not vary significantly with time, temporal variability of m should be the same as that of M.

The net flux of mass may be decomposed into two components, i.e.,

$$\overline{m} = - \overline{<u><c>} - \overline{UC}$$

where the overbars designate the temporal average over the study period. The first term on the right hand side is the net transport by the subtidal component and the second term is the net transport by the tidal component. The data collected near the bottom of the river mouth were used to calculate the net mass fluxes of salt and DO, contributed by total, tidal and subtidal components respectively. The results are listed in Table 1, together with the average values of each component of longitudinal velocity, salinity and DO. The root mean square values of tidal current, and tidal components of salinity and DO fluctuation are also presented in Table 1. It shows that, even though the tidal current is much stronger than the average current, the salinity and DO fluctuations at tidal time scales are much smaller than their respective average values. Furthermore, Fig. 5 indicates that the salinity and DO fluctuations at time scales longer than tidal are much larger than those at tidal or shorter time scales.

Table 1. Average values of measured quantities and mass fluxes

	Total Component	Subtidal Component	Tidal Component	Root Mean Square of Tidal Component
(a) Average over 596 hours				
u (cm s^{-1})	-6.31	-6.28	-0.03	16.6
s (psu)	18.5	18.5	0.00	0.26
- u × s	119	118	1	-
(b) Average over 333 hours				
u (cm s^{-1})	-7.93	-7.90	0.03	17.4
DO (mg l^{-1})	5.63	5.61	0.02	0.72
- u × DO	37.9	41.1	-3.19	-

The net mass fluxes of salt and DO by the tidal component amount to only 1% and 8% respectively of those by the subtidal components (Table 1). Therefore, we need to consider only the subtidal transport in studying the characteristics of the estuary-subestuary exchanges of salt and DO in this case.

29.4 Temporal variability of mass exchanges

The subtidal current was variable with a dominant time scale of several days (Fig. 5c). Since the freshwater discharge into the Rappahannock River was fairly constant over the two months prior to and during the field measurements, the variability may be attributed to meteorological forcing, primarily the wind. The wind events not only exert variations in wind stress over the water surface, but also cause variations in salinity structure in the bay near the river mouth. The salinity structure in the bay may also vary in response to changes in discharge from the Susquehanna River, the major freshwater contributor of the bay. However, the time scale of variations resulting from this change should be much longer than several days. The magnitude, and its amplitude of variation, of the observed subtidal current are consistent with the values derived from the theoretical model of Hansen and Rattray (1965). They have derived an equation describing the vertical distribution of steady-state estuarine circulation. In the equation, the circulation consists of three modes: barotropic circulation forced by freshwater discharge, baroclinic circulation driven by density gradient and the wind-driven circulation. Based on the observed longitudinal salinity gradient, the model estimates that, under the condition of 5 m s^{-1} wind speed,

$$u_g / u_w = 0.32$$

where u_g and u_w are, respectively, the maximum density driven and wind-driven velocities in the lower layer of the water column. Application of the formula proposed by Hansen and Rattray (1966) gives

$$u_g = -3.3 \text{ cm s}^{-1}$$
$$u_w = -10.2 \text{ cm s}^{-1}$$

These values are consistent with the subtidal current presented in Fig. 5c.

Further evidence of wind-induced variability may be discerned by considering Figs. 5a, 5b and 5c. Increased landward (or westward) current near the river bottom (Fig. 5c) always occurred when a strong wind was blowing from the southwest quadrant. The wind drove surface water out of the river, and thus lowered the surface elevation (Fig. 5b) and caused a setup in the bay that drove the bottom water into the river.

There were characteristic differences in the properties of water masses transported into the river over different periods of time. On five occasions during the month of measurements, the

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incoming water masses were characterized by low DO and high salinity, typically 3 to 4 mg l^{-1} lower DO and 2 to 3 psu higher salinity than those of other periods. The import of low DO, high salinity water masses was always accompanied by increased landward current and southwest wind. Comparing Figs. 5d and 5e with Figs. 3 and 4, it is evident that the low DO, high salinity water masses were derived from the lower portion of the water column in the bay. In the absence of strong southwest wind, the shallow sill off the river mouth blocks the bottom water in the bay from entering the river, and therefore, the imported water is derived from the surface layer in the bay, with higher DO and lower salinity.

29.5 Mechanisms of wind induced mass exchange

As indicated in Fig. 5, the intrusion of the low DO, high salinity water into the Rappahannock River has a time scale of 2 to 3 days. This agrees with the time scale of the effect of local wind forcing in Chesapeake Bay. Wang and Elliott (1978) observed that the subtidal sea level fluctuations in Chesapeake Bay had spectral peaks at periods of 20, 5 and 2.5 days. Their analyses associated the 2.5 day fluctuations with the local wind forcing, while the lower frequency fluctuations were related to non-local wind forcing through the interaction with coastal sea. With local wind as forcing function, there are two possible mechanisms that contribute to the transport of the subpycnocline water of the bay into the Rappahannock River. For simplicity, we may look at the wind-driven circulation around the river mouth in two-dimensional space, one in the horizontal plane and the other in the vertical plane.

29.5.1 Horizontal plane

Fig. 6 depicts the forcing of bottom current around the river mouth. The left panel is for the wind from the southwest. Surface set-up in two transects across the bay is considered (Fig. 6b). The surface current in both transects is toward the east because of direct wind stress and Coriolis effect. The boundary on the east causes the water surface to rise and a return westward flow along the bottom. There is a difference in the degree of surface set-down on the western boundary between sections AA and BB. Owing to the surface out-flow from the river, the set-down at section AA will be less than that at section BB. This results in a barotropic flow away from the river mouth along the western boundary of the bay (Fig. 6c). Therefore, the bottom water transported into the river is mainly derived from the main stem of the bay where hypoxic conditions exist in the summer. On the other hand, if wind is blowing from the northeast quadrant the surface set-up at the river mouth will be less than that at section BB. This results in an along shore barotropic flow toward the river mouth. If the density-driven gravitational circulation at the bottom of the river mouth is stronger than the wind-driven current, the bay bottom water will still be transported into the river. However, this bottom water is derived from near shore regions to the north and south of the river mouth. Because of the shallow depths in these regions, the imported water mass has high DO and low salinity.
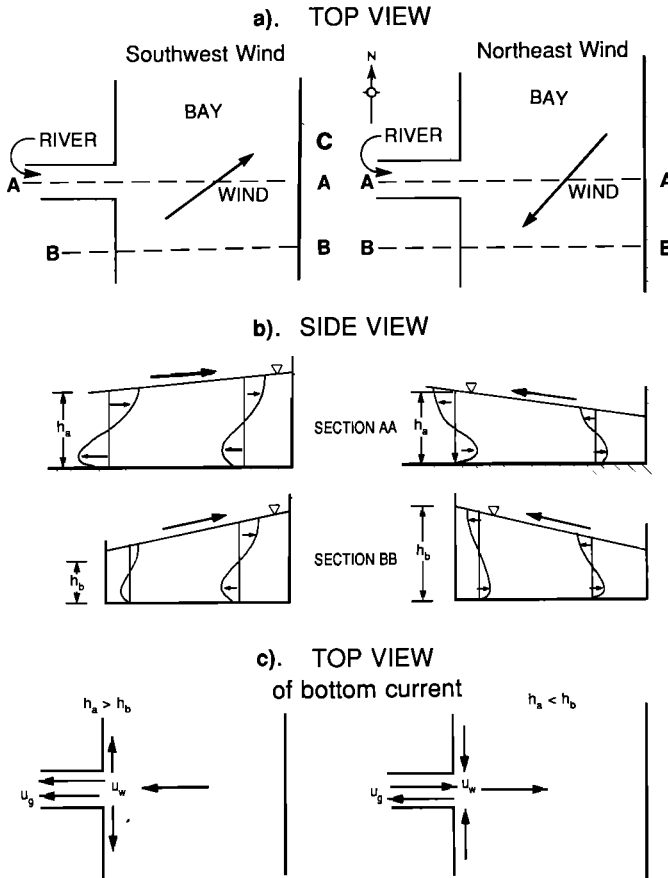
Figure 6: Directions of wind-driven currents around an estuary-subestuary junction.

29.5.2 Vertical plane

It was observed that the bay water is highly stratified, both in salinity and DO in summer months. Fig. 7 depicts the water surface and pycnocline in a transect across the bay (Section AA of Fig. 6). In the absence of wind, the pycnocline is nearly horizontal and is located at about the same height as the sill near the river entrance. Most of the water transported into the river by gravitational circulation is derived from the bay at a height above the pycnocline. In the time of strong southwest wind (Fig. 7b), the surface set-up and westward return flow near the bottom force the pycnocline to tilt upward on the western side of the bay. If the tilt is large enough to raise the pycnocline above the sill, it not only forces the subpycnocline water into the river, but also strengthens the bottom flow into the river. Therefore, strong inflow of low DO, high salinity water was observed (Fig. 5). Similar argument has been

proposed to explain the observed intrusion of hypoxic water onto the shallow flanks of the bay (Tyler, 1984; Malone et al., 1986). On the other hand, if strong wind is blowing from the northeast, the pycnocline will be depressed below the sill on the western side of the bay (Fig. 7c). The water transported into the river will be that above the pycnocline, thus with high DO, low salinity.
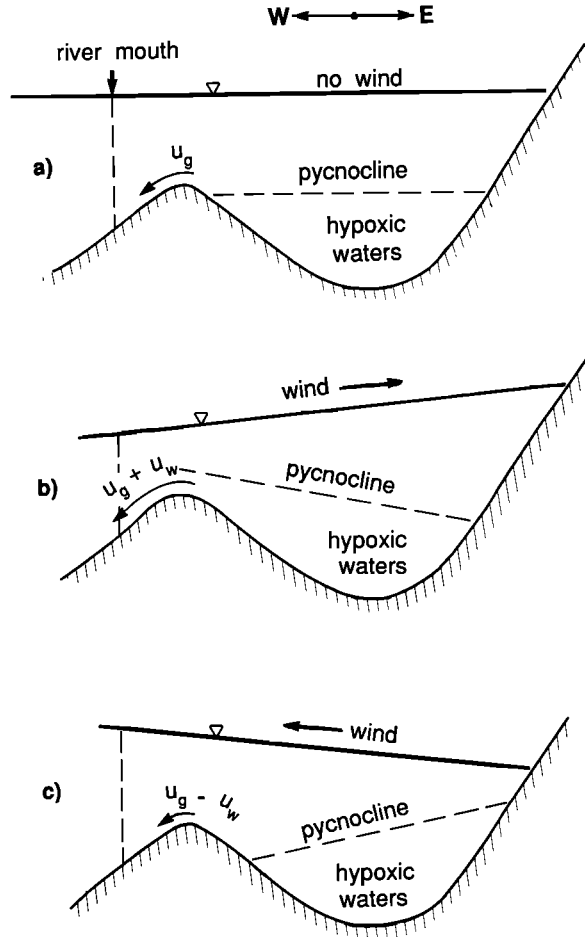

Figure 7: The tilting of pycnocline in a cross-bay section in response to wind.

29.6 Discussion

In this study, we have focused on the import of waters into the subestuary through the lower layer of the water column. Analyses of observed data indicate that transports of water, DO and salinity by subtidal components are an order of magnitude larger than those by tidal

components of water movements. Since we have measurements at only one point in the water column no quantitative estimate of the amount of mass transport may be made. However, the temporal variability of mass transport may be assessed assuming similarity in spatial distribution over time.

Two distinct characteristic water masses were observed being transported into the subestuary, the low DO, high salinity subpycnocline water and the high DO, low salinity surface water. Through simultaneous comparison of wind data with surface elevation, subtidal current, salinity and DO, we can conclude that wind is the dominant factor influencing the properties of imported waters. The hypoxic water was transported into the subestuary as episodic events when strong wind was blowing from the southwest. During these periods, the pycnocline in the western side of the bay rose above the sill at the river mouth and subpycnocline water in the bay 'rushed' into the river along the river bottom. The landward flow of gravitational circulation was further enhanced by the wind-driven current and increased longitudinal salinity gradient.

Other studies also demonstrated the temporal variabilities of bottom-water intrusions over the entrance sills of estuaries or subestuaries. Cannon et al. (1990) reported that the onset of the intrusions into Puget Sound were associated with the combined effects of neap tides and the horizontal density gradient across the sill. The regular spring-neap cycle was the result of mixing effects over the sill; however, the episodic occurrence of bottom-water intrusion in winter was attributed to wind events on the Pacific Coast. Other causes have been proposed for the variability of the density gradient across the sills of fjords (e.g., Gade and Edward, 1980). These include seasonal variation in density and upwelling outside of the entrance sill. For a fjord estuary, bottom-water intrusion is considered to play a dominant role in the replacement of water below sill depth, and the flushing of contaminants. On the other hand, the intrusions of bottom water from Chesapeake Bay into its subestuaries could mean the intrusion of hypoxic or anoxic water in the summer season, adversely affecting the water quality in subestuaries. Sanford (1988) also reported that intrusion of lower-layer water from the bay into the Choptank River occurred as an episodic, wind-driven internal surge rather than as a slow, steady flow. Since the Choptank River is on the eastern side of Chesapeake Bay, the intrusion events occurred when strong wind was blowing from the northeast.

The episodic nature of subpycnocline-water intrusions often results in the reversal of the longitudinal salinity gradient in the lowest reach of the subestuary. This would pose some difficulties in specifying boundary conditions for numerical modelling of subestuaries. A priori knowledge of conditions in the main estuary is required, or the model domain would need to be extended to include portions of the main estuary.

Acknowledgements

29.7 References

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