Tidal effects on estuarine circulation in the Chesapeake Bay

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Abstract

The response of the Chesapeake Bay to river discharge under the influence and absence of tide is simulated with a numerical model. Four numerical experiments are examined: 1) response to river discharge only; 2) response to river discharge plus an ambient coastal current along the shelf outside the bay; 3) response to river discharge and tidal forcing; and 4) response to river discharge, tidal forcing, and ambient coastal current. The general salinity distribution in the four cases is similar to observations inside the bay. Observed features, such as low salinity in the western side of the bay, are consistent in the model results. Also, a typical estuarine circulation with seaward current in the upper layer and landward current in the lower layer is reproduced in the four cases. The two cases without tide produce stronger mean currents than the cases with tide. This is due to stronger friction in the cases with tide. Additionally, the tidal mixing induces lower salinity inside the bay than the non-tide cases. Differences among salinity distributions for the four cases appear mostly outside the bay in terms of the outflow plume structure. The two cases without tide produce a northward branch of the outflow plume, while the cases with tide produce an expected southward plume. Increased friction in the cases with tide induces large horizontal variations of exchange flows at the entrance to the bay and makes the outflow contact the bottom. Therefore, a tendency for a bottom-advected plume appears in the cases with tide, rather than a surface-advected plume, which develops in the cases without tide. Furthermore, the tidal current favors a salt balance between the horizontal and vertical advection of salinity around the plume and hinders the northward intrusion of the plume.

1. Introduction

Studies on the circulation and salinity distribution in the Chesapeake Bay have a long history (Pritchard, 1952). Using observations in the Chesapeake Bay, Pritchard (1954; 1956) constructed first, a two-dimensional framework for the dynamics of estuarine circulation. In such framework, the effects of tidal currents to the salt balance and to the momentum balance in the bay were recognized. Recent observations across the entrance to the bay revealed that the exchange flow has large horizontal variations (Valle-Levinson et al., 1998). This indicates the necessity to treat the estuarine circulation in the Chesapeake Bay in a three-dimensional framework. An analytical solution (Valle-Levinson et al., 2003) shows that not only bathymetry variation but also friction is crucial to determine the transverse structure of exchange flow across the bay entrance. This suggests that tidal currents play an important role in determining the nature of exchange flows at the entrance to the Chesapeake Bay.

Compared to observations, numerical modeling efforts in the Chesapeake Bay appear not to be sufficient. Valle-Levinson et al. (1996) used an idealized model to study the water exchange between an estuary and the coastal ocean. Spitz and Klinck (1998) simulated the tides in the Chesapeake Bay by assimilating data from tide gauges. Wang and Johnson (2000) developed a three dimensional hydrodynamic model for the Chesapeake Bay and drove the model using the realistic forcing from 1985 to 1994. Xu et al. (2002) demonstrated the improvements of model results by assimilating high-resolution salinity data. So far, several other three-dimensional numerical models are being developed (www.ccmp.chesapeake.org) and most of them are designed for simulations aimed to reproduce the system rather than to understand it.

In this study, we use a three-dimension model to examine the effects of tidal currents on the estuarine circulation in the Chesapeake Bay. Because of important effects of bathymetry in the bay (Valle-Levinson et al., 2003), we use a fine grid size (~400 m). Fresh water discharge, tidal forcing, and ambient coastal current outside the bay are used to drive the model. By inclusion and absence of tidal forcing and ambient coastal currents, their effects to the mean currents inside the bay and the plume outside the bay are examined. Although these results are targeted to the Chesapeake Bay, they should be applicable to any wide estuary.

After descriptions of the numerical model in section 2, we show the model results of the four cases in section 3. Analysis on the dynamics of the plume outside the bay is given in section 4, followed by a summary in section 5.

2. Numerical model

We use one of the community ocean models, Princeton Ocean Model (POM), as the basic model of our simulations. The POM is a three-dimensional primitive equation ocean model, which includes full thermodynamics and Level 2.5 Mellor-Yamada turbulence closure model (Blumberg and Mellor 1987; Mellor 1998). The model domain and topography are shown in Figure 1. The horizontal resolution is 1/240 degree in both the zonal and meridional directions. In the vertical, 11 sigma-levels are evenly arranged. The minimum water depth in the model domain is set as 3 m. The time step is 3 seconds for the external mode and 120 seconds for the internal mode. During the calculations, the vertical eddy viscosity and diffusivity are obtained with the Mellor-Yamada turbulence closure model, the horizontal eddy viscosity is calculated by the embedded Smagorinsky formula with a proportionality parameter of 0.1, and the horizontal eddy diffusivity is obtained using a given inverse Prandtl number (0.5).

At the surface, no wind stresses are imposed. The bottom stresses (t_x, t_y) are calculated using a quadratic friction law.

$$(t_x, t_y) = r C_z (u, v)\left(u^2 + v^2\right), \qquad (1)$$

where r is water density, u and v are zonal and meridional components of velocity. The bottom drag coefficient is calculated by the embedded formula in the POM (Mellor, 1998),

$$C_z = \frac{k^2}{\left[\ln\left(0.05H/z_0\right)\right]^2}, \qquad (2)$$

where $k = 0.4$ is the von Karman constant, H is the water depth, and z_0 is the roughness parameter that is set to 0.1 cm.

The forcing used to drive this model are river discharge, ambient coastal current and tidal forcing. A total of 2200 m^3/s of fresh water is introduced into the model domain from the Susquehanna (51% of total), Potomac (18%), James (14%), Rappahannock (4%), York (2%), and the other small rivers (11%)(see Figure 1 for the position of rivers). The river input is distributed to the corresponding grid points shown in Figure 1 by specifying the vertical velocity $w_s = -Q/(N\Delta x\Delta y)$ (Oey, 1996), where Q is river discharge (m^3/s), N is the number of grid points covered by the river head and Δx and Δy are the sizes of grid points in the east and north direction, respectively. The salinity of river discharge is assumed to be 0 psu.

The most dominant tidal constituent in the Chesapeake Bay is the M_2 (Browne and Fisher, 1988) and is introduced in the simulations with the oscillation of tidal currents along the eastern and southern boundaries. The harmonic constants of tidal currents along the eastern and southern boundaries are calculated in advance using a horizontal two-dimensional model in which the oscillation of sea level is imposed. The amplitude and phase of the M_2 tide along the eastern and southern boundaries are 46 cm and 210 degrees, which produce similar calculated charts to observations in the entire Chesapeake Bay (Browne and Fisher, 1988).

Prescription of the southward ambient current in the shelf was motivated by previous studies. Beardsley and Boicourt (1981) have reported a southward coastal current in the Middle Atlantic Bight. Epifanio and Garvine (2001) present the horizontal distribution of the coastal current south of Delaware Bay and infer the existence of a southward coastal current outside the Chesapeake Bay. In our simulations, we impose an ambient coastal current with a speed of 10 cm/s along the eastern boundary from 37.7N to 38N.

The open boundary conditions are specified in the same way as that in the subroutine BCOND in POM. In the case that includes both tidal currents and ambient current, linear superposition of the two currents at the open boundary is used.

In this study, we compare four cases. In the first case (Case 1), only river discharge is introduced; in the second case (Case 2), river discharge and ambient coastal current are introduced; in the third case (Case 3), river discharge and the M_2 tide are prescribed; in the fourth case (Case 4), river discharge, ambient coastal current, and the M_2 tide are introduced. All the simulations start from rest with an initial salinity of 32 psu at all grid points and run for 360 days. The results of Case 3 and Case 4 are saved hourly and the tidal components are removed with a tide killer filter (Hanawa and Mitsudera, 1985).

Before examining the mean current, the amplitudes and phases of M_2 tide and tidal current are shown to confirm the reproduction of tidal signals in the model. The amplitude of M_2 tide (Figure 2a) presents the same features as those from the observations (Browne and Fisher, 1988) and the results of data assimilation (Spitz and Klinck, 1998). The amplitude of M_2 tide is slightly lower along the western coast than along the eastern coast. Two regions with the lowest amplitude are found around 37.6N and 38.8N along the western coast. The phase of M_2 tide (Figure 2b) varies more than one tidal cycle from the mouth to the head of the bay, which is consistent with observations (Browne and Fisher, 1988). The M_2 tidal current (Figure 3a) is strong in the lower bay, also at the junction of the Potomac River and Chesapeake Bay, and in the upper bay. The features included by these distributions as well as the values of the amplitude of M_2 tidal current are consistent with observations (Browne and Fisher, 1988). The phase of M_2 tidal current (Figure 3b) increases from the mouth to the head of the bay, showing a similar distribution with the phase of tide. The effect of friction to the tidal current is manifested in Figure 3b. The smaller value of phase near the coast than in the central part of the Chesapeake Bay results from the strong friction in the shallow water. This feature can be also found in observations (Browne and Fisher, 1988). As for the vertical structure of tidal current (not shown here), the modeled tidal current at the entrance to the bay weakens visually near the bottom, which is consistent with observations by Valle-Levinson et al. (1998).

3. Model Results

3.1 Horizontal distribution of salinity and mean velocity

The mean velocity and the mean barotropic velocity in the model domain (Figure 4a) indicate that the motion reaches steady state at day ~200 although the salinity decreases slightly (Figure 4b). The baroclinic velocity, i.e., the difference between the two time series (mean and barotropic) of each case in Figure 4a, develops in the first 100 days and approximately reaches a steady state after day 200. Comparison of the velocity fields and salinity fields shows that the velocity in the entire domain and the salinity outside the bay have little difference on days 180, 240, 300, and 360. The decrease in salinity occurs mostly inside the bay. Here we choose the results of day 240 as the solution of each case because the mean salinity inside the bay at day 240 is close to the observed value (~18 psu, Austin, 2002).

The general distribution of salinity inside the bay in the four cases (Figure 5 & 6) is similar to observations (Pritchard, 1952). Observed features such as low salinity in the western side of the bay are consistent in the model results. The salinity of the upper bay is lower in the cases with tide (Figure 6) than in the cases without tide (Figure 5). The most significant differences in the salinity distributions among the four cases appear outside the bay. Case 1 produces a plume reaching 37.5N, far upstream (in the Kelvin wave propagation sense) from the mouth; the southward ambient coastal current in Case 2 largely corrects the northward plume but not completely; tidal forcing in Case 3 weakens greatly the northward plume; the combination of tidal forcing and ambient coastal current in Case 4 produces a realistic plume, as observed by Marmorino et al. (2000).

A typical estuarine circulation with seaward current in the upper layer and landward current in the lower layer (not shown here) is reproduced inside the bay in the four cases (Figure 7 & 8). The inclusion of tide makes the current concentrate in the deep channel inside the bay. The reason for this is explained in section 3.3. As with salinity, the most significant difference in the current fields is associated with the outflow plume outside the bay. The cases without tide produce a northward branch of the outflow plume (Figure 7), while the cases with tide produce an expected southward plume (Figure 8).

The influence of the ambient coastal current on modifying the northward plume is clear (Figure 7), as suggested by many previous plume simulations that a downstream ambient current can eliminate the upstream propagation of plume water on the shelf (Chapman and Lentz, 1994). The tidal currents, however, also constrain the development of an upstream plume as shown in Figure 8. A dynamical explanation on this phenomenon is given in Section 4.

3.2 Vertical structure of salinity and mean currents

The section along the longitudinal axis of the bay displays a typical distribution of salinity and current in a partially mixed estuary (Figure 9). There is little difference in the results inside the bay between the cases with and without the ambient coastal current. Therefore, the results of Case1 and Case3 are not shown from Figure 9 to Figure 11. Figure 9a shows that the salinity increases gradually from the head to the mouth of the bay and stratification is maintained in the vertical. The inclusion of tidal current decreases the salinity inside the bay and weakens the stratification (Figure 9b). The salinity distribution shown in Figure 9 is similar to observations in autumn and winter (Seitz, 1971). In spring and summer, however, the stratification in observations (Seitz, 1971) is stronger than the model results (Figure 9). The surface heating and large fresh water discharge in spring and summer could be response for the stronger stratification in observations.

The mean currents along the longitudinal axis of the bay show a typical estuarine circulation with a seaward current in the upper layer and a landward current in the lower layer (Figures 9c). The typical speed of currents is 10~20 cm/s and the level of no motion is located roughly at 5 m. The inclusion of tidal forcing weakens slightly the landward current in the lower layer, deepens the level of no motion, and intensifies the seaward current in the upper layer (Figure 9d). It must be noted that the intensified seaward current in the upper layer in Case 4 (Figure 9d) is caused by the concentration of current in the central deep channel

(Figure 10). This concentration of mean outflow in the deep channel responds to the variation of vertical eddy viscosities and will be discussed in Section 3.3. In Figure 10, the landward current is stronger in Case 2 than in Case 4, indicating that tidal forcing weakens the landward current inside the Chesapeake Bay. The lifting of the level of no motion above the channel in Case 2 induces two cores of the seaward current in the surface layer. As the tidal forcing is included, the lifting of the level of no motion vanishes and the seaward current concentrates above the channel.

In order to further evaluate the vertical structure of mean current and salinity, a cross-section at the entrance (see Figure 1 for its position) is chosen to correspond with that of observations by Valle-Levinson et al (1998) and Reyer-Hernandez (2001). There are two channels in the section. The Chesapeake Channel is located at southern side and the North Channel at the northern side. The shallow area between these two channels is called Six-Meters Shoal. The salinity is low in the southern side and high in the northern side of the entrance (Figure 11a). The stratification is stronger in the Chesapeake Channel than in the North Channel. These features are consistent with the observations (Valle-Levinson and Lwiza, 1997; Reyer-Hernandez, 2001). As the tidal forcing is included, the stratification weakens (Figure 11b). The strong tidal current (Figure 3a) causes more mixing at the entrance to the Chesapeake Bay.

The vertical structure of the current across the section corresponds to a two-layered current in the main channel: outflow in the upper layer and inflow in the lower layer (Figure 11c). Inflow appears only near the bottom of the two channels. The inclusion of tidal forcing transforms the level of no motion from being flat to being sloped (Figure 11d). This indicates that tidal forcing produces larger horizontal variations of the mean current. Consequently, the outflow shifts to the southern part of the Chesapeake Channel and to the North Channel and a weak inflow appears between the two channels, i.e., above the Six-Meters Shoal. This distributions reproduced by the case with tide are remarkably similar to those observed by Valle-Levinson et al. (1998). In turn, the current along the section at the entrance is southward in the upper layer and northward in the lower layer (Figure 11e). The strength of southward and northward currents is sensitive to tidal forcing. In the case with tide, the northward current in the lower layer weakens largely and the southward current dominates the section (Figure 11f).

In addition to river discharge and tidal forcing, wind forcing also influences the subtidal currents observed at the entrance. However, the essential characteristics of the currents observed across the section at the bay entrance are that the outflow in the surface layer concentrates in the southern part of the Chesapeake Channel and in the North Channel, the inflow concentrates in the lower layer of the two channels, the level of no motion in Chesapeake Channel slopes steeply, and the current above the Six-Meters Shoal reflects a weak inflow. All these features are consistent with the results of Case 4 (Figure 11d) rather than those of Case 2 (Figure 11c). Furthermore, the observations show that the southward current in the surface layer dominates the current along the section at the entrance (Velle-Levinson et al., 1998). This is also consistent with the results of Case 4 (Figure 11f) rather than those of Case 2 (Figure 11e). Therefore, tidal forcing is necessary to reproduce the mean currents at the entrance to the Chesapeake Bay.

3.3 Tidal effects on the vertical eddy viscosity

According to Kasai et al. (2002), friction plays a preponderant role in the transverse variation of exchange flows in estuaries. Valle-Levinson et al. (2003) use this finding to show that the horizontal variation of exchange flows at the Chesapeake Bay entrance is explained by the increase of Ekman number. Vertical eddy viscosity varies largely with the inclusion of tidal forcing (Figure 12). Compared to Case 2 (Figure 12a), the vertical eddy viscosity increases significantly near the bottom inside the bay and the area around the entrance in Case 4 (Figure 12b). At a transverse section in the middle of the bay, the vertical eddy viscosity decreases slightly near the interface between the seaward and landward currents in Case 4 (Figure 12d). This might be due to decreased vertical shears related to Case 4 relative to Case

2. At the transverse section of the bay entrance, the vertical eddy viscosity increases significantly in the entire section in Case 4 (Figure 12f).

The weakening of landward flows in Case 4 (Figures 9&10) is a natural response to the increase of vertical eddy viscosity near the bottom with the inclusion of tidal forcing. The response of transverse structure of the exchange flows in the middle bay (Figure 10) and in the entrance (Figure 11) to the vertical eddy viscosity is consistent with the analytical solution given by Valle-Levinson et al. (2003). In the middle bay, the apparent larger friction in Case2 lifts the interface of exchange flows upward, i.e., induces larger horizontal variation of exchange flows (Figure 10a) relative to Case 4. At the entrance, larger friction in Case 4 increases the tilt of the interface of exchange flows (Figure 11d), which also induces a larger horizontal variation of exchange flows.

4. The plume outside of the bay

As shown in Figures 6 and 7, Case 1 produces a northward plume outside the Chesapeake Bay. The ambient coastal current in Case 2 largely corrects the northward plume but not completely. Tidal forcing in Case 3 weakens the northward plume but cannot eliminate it. The combination of ambient coastal current and tidal forcing in Case 4 produces a realistic plume. Case 1 seems to represent an extreme case of surface-advected plume and Case 4 illustrates a bottom-advected plume. These concepts are explained next.

4.1 Surface-advected plume versus bottom-advected plume

Yankovsky and Chapman (1997) show two extreme cases of plume: the bottom-advected plume and the surface-advected plume. In the bottom-advected plume, the plume water occupies the entire water column and turns right immediately after it flows out the estuary mouth. The outflow associated with the bottom-advected plume contacts the bottom. In the surface-advected plume, the plume water stays on the surface layer forming a thin layer above the ambient denser water. The plume typically forms a bulge near the estuary mouth within which an anticyclonic flow is generated. The outflow associated with the surface-advected plume is limited to the surface layer and has little contact with the bottom.

In the model results presented here, the plume in the cases without tide (Figure 5) is consistent with a surface-advected plume, while the plume in the cases with tide (Figure 6) emulates a bottom-advected plume. The vertical distribution of salinity along a section across the plume outside the bay confirms this inference (Figure 13). In the cases without tide, the isohalines lean toward the front of plume (Figure 13a), which is the response in a surface-advected plume (Yankovsky and Chapman, 1997). The ambient coastal current does not change this basic structure (Figure 13b). In the cases with tide, the salinity is vertically well mixed (Figure 13c & 13d) and consistent with the isohalines in a bottom-advected plume (Yankovsky and Chapman, 1997). Returning to the vertical distribution of exchange flows at the bay entrance (Figure 11), the outflow in Case 4 shows more contact area with the bottom than in Case 2. This again confirms the inference that tidal forcing transforms the outflow of Chesapeake Bay water from a surface-advected plume to a bottom-advected plume. It is possible that during very high fresh water outflow and weak tidal current, the bay plume behaves as surface-advected.

According to Chapman and Lentz (1994), the upstream movement of a surface-advected plume can be explained as a self-advected process. Applying their explanation to the northward movement of the plume outside the bay, we need a positive acceleration of northward velocity and a negative temporal variation of salinity near the plume front. To confirm these two points, the terms in the momentum equation are analyzed for the northward velocity and the terms in the conservation equation for salinity are evaluated.

4.2 Dynamic analysis of northward velocity and salinity

The momentum equation of northward velocity v can be expressed as

$$\frac{\partial v}{\partial t} + ADV = PRE + COR + VDIF \, , \tag{3}$$

where, ADV denotes advection terms, PRE denotes pressure gradient, COR denotes Coriolis force, $VDIF$ denotes the internal stress divergence related to vertical eddy viscosity. The stresses related to horizontal viscosity are small enough to be negligible.

In Figures 14 and 15, we present the surface distributions of each term in (3) plus salinity contours of 31.5 from each of the four cases, which approximately represent the plume front (Figure 13). The dynamic terms are shown on day 100 and the salinity contours on days 90, 120, and 150. The plume front in Case 1 moves northward with a speed of 0.7 km/day. Behind the plume front, i.e., south of the plume front, there is an area with positive acceleration close to the coast (Figure 14a). The plume front in Case 3 also moves northward. Its speed is smaller than that in Case 1. The positive acceleration of northward velocity behind the plume is weaker in Case 3 than in Case 1. In Case 2 and Case 4, positive acceleration of northward velocity behind the plume weakens greatly (Figure 14b&d).

The advection terms (Figure 14e-h) are generally small except for the bay entrance in the cases with tide and the northeastern corner where the ambient coastal current is imposed. In the cases with tide, tidally averaged advection terms correspond to the generation of tide-induced residual current, which is large around the entrance of the bay.

The combination of pressure gradient and Coriolis force (Figure 14i-l) is responsible for the positive acceleration of northward velocity behind the plume front. This is particularly clear in Case 1 (Figure 14i). Figure 15 indicates that the positive acceleration of northward velocity is caused by the pressure gradient. In Case 1 (Figure 15e), an area with positive pressure gradient exists from the bay entrance to the plume. The ambient coastal current in Case 2 brings negative pressure gradient ahead of the plume front, i.e., north of the plume front. Tidal forcing in Case 3 weakens the positive pressure gradient behind the plume front.

Internal stresses (Figure 15a-d) vary greatly with the inclusion of tidal forcing. Behind the plume front, the internal stresses show negative values and therefore act to prevent the northward movement of the plume. In Case 1 (Figure 15a), negative internal stresses are very weak and allow northward propagation. In Case 3 (Figure 15c), stresses increase greatly. Therefore, tidal forcing increases the internal stresses in the plume front area and hinders northward plume propagation.

Further analysis may be derived from the salinity equation, which can be expressed as

$$\frac{\partial S}{\partial t} + ADVS = VDIFS \; , \qquad (4)$$

where, S denotes salinity, $ADVS$ denotes advection, and $VDIFS$ denotes vertical diffusivity. Similarly to the momentum equation (3), the horizontal diffusivity is negligible.

The temporal variation of salinity around the plume front is negative in all the four cases (Figure 16), indicating the decrease of salinity. The magnitude of negative temporal variation of salinity around the plume front in Case 1 and Case 2 is large while that in Case 3 and Case 4 is small. Figure 16 indicates that the imbalance between advection and vertical diffusivity causes the decrease of salinity.

The advection of salinity in Figure 16 is then separated into two components: horizontal and vertical (Figure 17). Around the plume front, horizontal advection is positive and vertical advection is negative. This suggests that vertical advection compensates the salinity loss caused by horizontal advection. In Case 1 and Case 2, the horizontal advection is not balanced with the vertical advection. However, in Case 3 and Case 4, the advections in the two directions are nearly balanced. Therefore, the self-advected process of a surface-advected plume suggested by Chapman and Lentz (1994) is confirmed in the cases without tide. In Case1, the northward pressure gradient behind the plume front is not balanced by the other terms and produces a positive acceleration of northward velocity along the coast north of the entrance. Meanwhile, the horizontal advection of salinity is not balanced by the other processes and causes the decrease of salinity. These two effects make the plume propagate northward. On the other hand, the inclusion of tidal forcing increases friction, decreases the northward pressure gradient around the plume, and makes the horizontal and vertical advection of salinity to be balanced. Therefore, the northward plume outside the bay is severely hampered.

5. Conclusions

Tidal forcing affects not only estuarine circulation and the salinity distribution inside the Chesapeake Bay but also the river plume outside the bay. Increased mixing by tidal currents induces lower salinity inside the bay and weakens stratification. Inside the bay, the seaward flow in the surface layer concentrates in the deep channel and the landward flow in the bottom layer weakens. Additionally, tidal forcing induces large horizontal variation of exchange flows at the bay entrance and makes the outflow plume contact the bottom. Consequently, a bottom-advected plume is formed in the presence of tidal forcing. Within the plume outside the bay, tidal currents increase friction, and make horizontal and vertical advection of salinity to be balanced. These two effects, favored by tidal forcing, hinder the upstream movement of the plume. In general, because the transverse structure of exchange flows in estuaries is sensitive to friction (Valle-Levinson et al., 2003), which tidal forcing influences greatly, it is necessary to simulate estuarine circulation in the presence of tidal forcing. This is not only true in the Chesapeake Bay, but also in most estuaries.

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