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The variability of currents and sea level in the upper Delaware estuary

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ABSTRACT

The variability of currents and sea levels in the upper Delaware estuary are examined based on measurements from bottom mounted acoustic Doppler current profilers (ADCP) deployed at two sites (New Castle and Tinicum) from 18 March to 10 June 2003. New Castle is located 104 km from the mouth, and Tinicum is located another 32 km up-estuary. Supplemental data, including sea level at the mouth of the estuary, river discharge, and wind speed and direction, were also obtained from various federal agencies. The instantaneous current represents a superposition of variability driven by the tide, wind, and river discharge. Over the short (<36 hr) time scale, the tide is the dominant forcing mechanism, with M_2 being the principal tidal constituent. The amplitude of the M_2 tide increases from the mouth to the upper estuary and gives rise to a vigorous M_2 current of the order 80 cm s^{-1}. On time scales of 36 to 120 hr, the effect of wind drives a weak subtidal current with a standard deviation of 2 cm s^{-1} in the upper estuary. At time scales longer than 120 hr, the subtidal current variability, with a standard deviation of 6 cm s^{-1}, is dominated by the barotropic response of the upper estuary to variations in the river discharge. The upper estuary exhibits a strong down-estuary mean current of the order—15 cm s^{-1}. At Tinicum, river discharge accounts for more than half of the mean current, which is characterized by down-estuary flow throughout the water column. The magnitude of the river discharge-induced mean current is reduced at New Castle, in direct response to the down-estuary increase in the cross-sectional area. Tidally rectified current accounts for the remainder of the overall mean flow at Tinicum, and the effect of tidal rectification may be more important than river discharge in producing the mean flow at New Castle. There is no evidence of a baroclinic gravitational circulation, as the salt intrusion generally does not extend into the upper estuary.

1. Introduction

The physical exchange in an estuary may be forced by a wide range of mechanisms operating over vastly different time scales. Among the various mechanisms, the most energetic ones are often associated with the propagation of the astronomical tides from the adjacent continental shelf into the interior of the estuary. To first order, these tidal motions operate primarily at the semi-diurnal and/or diurnal time scales.

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At longer time scales of a few days the exchange process may be driven by winds (Weisberg, 1976; Weisberg and Sturges, 1976; Smith, 1977, 1978; Kjerfve *et al.,* 1978; Wang and Elliott, 1978; Wang, 1979a,b; Noble *et al.,* 1996) through the remote and local effects. The local wind effect represents the frictional coupling of wind over the surface area of the estuary. A wind blowing landward up the estuary often results in a set-up in the upper part of the estuary, whereas a wind blowing seaward down the estuary would produce a set-down in the upper estuary. The remote wind effect includes the action of the wind over the continental shelf directly adjacent to the estuary. Through coastal Ekman transport, a downwelling-favorable wind over the shelf would produce a rise in coastal sea level at the mouth of the estuary, thus forcing a rise in sea level in the interior of the estuary and a corresponding landward transport into the estuary. An upwelling-favorable wind would have the opposite effect in producing a drop in coastal sea level, thus forcing a drop in interior sea level and a seaward flow out of the estuary. In addition to this forced response, the coastal sea level at the mouth of the estuary may also include the down-shelf propagation of free waves originated from the continental shelf far away from the estuary (Wang, 1979a; Ou, 1981). Regardless of the exact mechanism, the remote wind effect can indirectly force the exchange via the sea level fluctuations at the mouth of the estuary.

At yet longer time scales the estuary may be forced by the river discharge that enters the system via one or more sources. Upon entering the estuary, the river discharge may produce a barotropic as well as a baroclinic response (Garvine, 1991). The barotropic current is produced in response to the volume input associated with the river discharge, and the baroclinic (gravitational, density driven) current is generated in response to the buoyancy input associated with the river discharge (Pritchard, 1952a, b). Given a river discharge, the barotropic and baroclinic currents may carry very different structures.

In addition to the river discharge, there are other processes that can generate currents on long time scales. For example, for a weakly nonlinear system, the nonlinear interaction of the first order tides may produce a second order, tidally rectified mean flow with a characteristic magnitude of $(\eta_0/h)U_0$, where η_0 is the amplitude of the first order tidal elevation, h is the water depth, and U_0 is the amplitude of the first order tidal current (Ianniello, 1977, 1979; Uncles and Jordan, 1980).

The relative importance of these mechanisms in forcing currents at tidal and subtidal frequencies may change significantly along the length of an estuary. This in turn has enormous consequences for the transport and distribution of waterborne material in estuarine waters. A large number of studies have been conducted during the past two decades to examine the physical exchange process in the lower reaches of major estuaries such as the Chesapeake (Valle-Levinson, 1995; Paraso and Valle-Levinson, 1996; Valle-Levinson and Lwiza, 1998; Valle-Levinson *et al.,* 1998) and Delaware Bays (Garvine, 1985, 1991; Garvine *et al.,* 1992, Münchow *et al.,* 1992; Wong, 1994; Wong and Münchow, 1995; Wong and Moses-Hall, 1998; Janzen and Wong, 2002). In contrast, relatively few studies have been conducted in the upper reaches of these estuaries, landward of the salt intrusion limit, where residual currents and mass transports driven by gradients in salinity are not present. Many of the world's most economically vital ports and industries are situated within the tidal river reach of river-estuarine systems, thus there is a general need to understand circulations specific to tidal rivers in order to address issues such as channel shoaling and contaminant transport. The present study examines a set of observational data collected in the upper Delaware estuary in order to elucidate the characteristics of the tidal and subtidal exchange associated with different forcing mechanisms.

2. Study area and data sources

The Delaware estuary (Fig. 1) is the second largest estuary along the east coast of the United States. It has a length of 215 km, from the mouth between Cape Henlopen, Delaware and Cape May, New Jersey to the head at Trenton, New Jersey. The head of the estuary is marked by falls that prevent any coastal disturbances such as tides from propagating farther upstream while allowing the Delaware River discharge to flow down the estuary. The mouth is about 18-km wide. From there the width of the estuary first increases to a maximum of 45 km in the lower estuary, and then decreases nearly exponentially with distance farther upstream. The mean depth of the estuary is about 10 m.

The Delaware River, with a mean discharge of 330 $m^3\ s^{-1}$ (Wong, 1995) accounts for about 60% of the total discharge. The Schuylkill River, entering the estuary at Philadelphia, Pennsylvania, accounts for another 15%. No other single source accounts for more than 1% of the total discharge (Sharp *et al.,* 1986). Münchow *et al.* (1992) have estimated the semi-diurnal tidal volume flux at the mouth to be $1.47 \times 10^5\ m^3\ s^{-1}$. The fact that the tidal volume flux is more than two orders of magnitude greater than the river discharge indicates that the Delaware is a weakly stratified estuary. Garvine *et al.* (1992) have examined the axial distribution of salinity in the estuary and found that the mean salt intrusion limit is located at about 97 km upstream of the mouth, some 10 km north of the point where the C&D Canal enters the estuary (Fig. 1). Wong (1994) found significant transverse variability in the buoyancy-driven currents in the lower estuary, with two branches of outflow along the Delaware and New Jersey shores and an inflow in the deeper part of the channel. In the middle reach of the estuary the presence of a two-layer gravitational circulation has been documented by Wong and Garvine (1984).

The primary source of data for the present study was from an 80-day survey conducted from 18 March to 10 June 2003. During that period two bottom mounted R D Instruments 600 kHz acoustic Doppler current profilers (ADCP) were deployed, one at Tinicum Island, and the other at New Castle (Fig. 1). At Tinicum the ADCP provided current measurements from 2 m above the bottom to the surface with a vertical resolution of 0.5 m. At New Castle the currents were available from 3 m above the bottom up to the surface with the same vertical resolution. In addition to currents, surface elevations at the two sites were derived from the pressure data provided by the ADCP. Optical backscatter was recorded at two points in the water column to investigate suspended-sediment concentrations and flux mechanisms (see Cook *et al.,* 2007). The focus of the present study, however, is on the tidal and subtidal variability in sea level and current in the upper Delaware estuary.

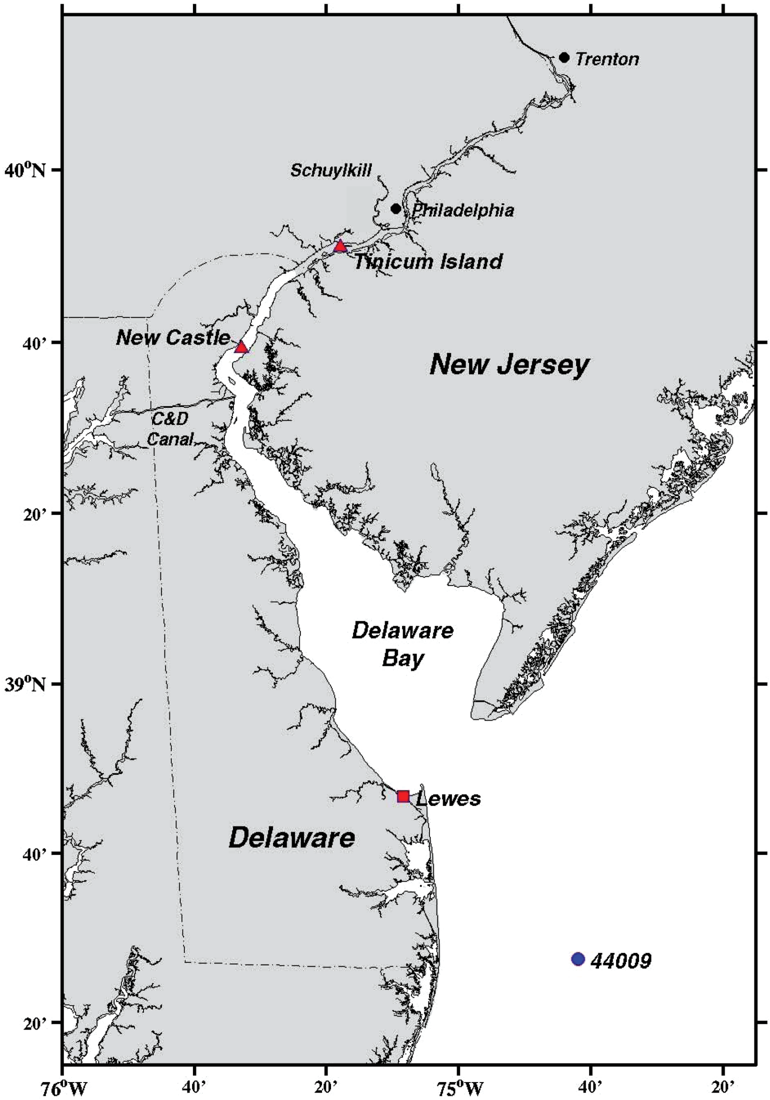

Figure 1. Location map of the Delaware estuary. Current and sea level measurements were collected at mooring sites at Tinicum and New Castle. Sea level data for a tide gauge at Lewes, wind data from environmental buoy 44009, and discharge data for the Delaware and Schuylkill rivers were acquired from various federal agencies (see text for sources).

The cross section at Tinicum has a width of about 900 m. The bottom bathymetry at this cross section is dominated by a 15-m deep, U-shaped channel flanked by narrow ($\approx$100 m) and shallow shoals. The ADCP was deployed slightly off the deepest part of the channel at a water depth of 12 m. The cross section at New Castle has a width of 2500 m. It has a more complex bathymetry, with a 15-m deep, 500-m wide channel flanked by extensive areas of flats with water depth of about 8 m, which then taper to shallow shoals along the shores. The ADCP at New Castle was deployed at the left edge (as one looks up the estuary) of the channel with a water depth of 9 m.

To supplement the data gathered at the two sites, sea surface heights near the mouth of the estuary were obtained from the National Ocean Service's tide gauge at Lewes, Delaware to provide information on coastal sea level fluctuations. Furthermore, wind speeds and directions from National Data Buoy Center (NBDC) Buoy 44009, located 30 km offshore of the mouth of the estuary, were obtained to provide wind conditions over the study area. Discharge data for the Delaware River at Trenton and the Schuylkill River at Philadelphia were obtained from the United States Geological Survey.

3. The tidal variability

Figure 2 shows time series of surface elevations from Tinicum, New Castle, and Lewes. It is apparent that the raw data contain variability at tidal as well as subtidal frequencies. To facilitate further analysis, the tidal and subtidal fluctuations in sea level are separated by passing the raw data time series through a low-pass Lanczos filter (Bloomfield, 1976) with a cut-off period of 36 hr. The tidal oscillations in sea level are then obtained by subtracting the low-passed time series from the raw data time series. To quantify the tidal variability in sea level, a least squares harmonic analysis was performed to obtain the amplitude and phase of the tidal constituents. The harmonic analysis shows that the tidal variability is dominated by the M_2 tide. The amplitude of M_2 is 58.3 cm at Lewes, and it increases to 74.6 cm at New Castle and 74.3 cm at Tinicum, reflecting the significance of the funnel-shaped geometry in amplifying the semi-diurnal tide that propagates from the ocean up the estuary (Parker, 1991). Next to the M_2 tide at Lewes there are two other semi-diurnal tides (N_2 and S_2) with amplitudes of 15.1 cm and 12.2 cm, respectively. The modulations between M_2 and the weaker N_2 and S_2 give rise to fortnightly and monthly spring-neap cycles. In addition to the semi-diurnal tides there are also diurnal tides in the Delaware estuary, with K_1 being the leading diurnal constituent. There is no appreciable increase in the amplitude of the K_1 tide from Lewes (9.4 cm) to Tinicum (10.0 cm).

Before examining the vertical profile of the tidal currents in the upper Delaware estuary, it is instructive to examine the depth-averaged currents at the two sites first. Figure 3 shows the raw data time series of depth-averaged current at New Castle and Tinicum. For the sake of brevity, only the major axis component of the current will be considered here. A positive current indicates water flowing up-estuary. A visual inspection suggests that the current is dominated by the semi-diurnal variability. The results of harmonic analysis indicate that the amplitude of the M_2 tidal current is 82.4 cm s^{-1} at New Castle and 83.2 cm s^{-1} at Tinicum. The weaker N_2 tidal current has amplitudes of 17.1 cm s^{-1} and 17.6 cm s^{-1} at New Castle and Tinicum, respectively. The diurnal tidal currents are much weaker than the semi-diurnal ones, and the amplitude of the K_1 current is only 5.4 cm s^{-1} at New Castle and 5.0 cm s^{-1} at Tinicum.

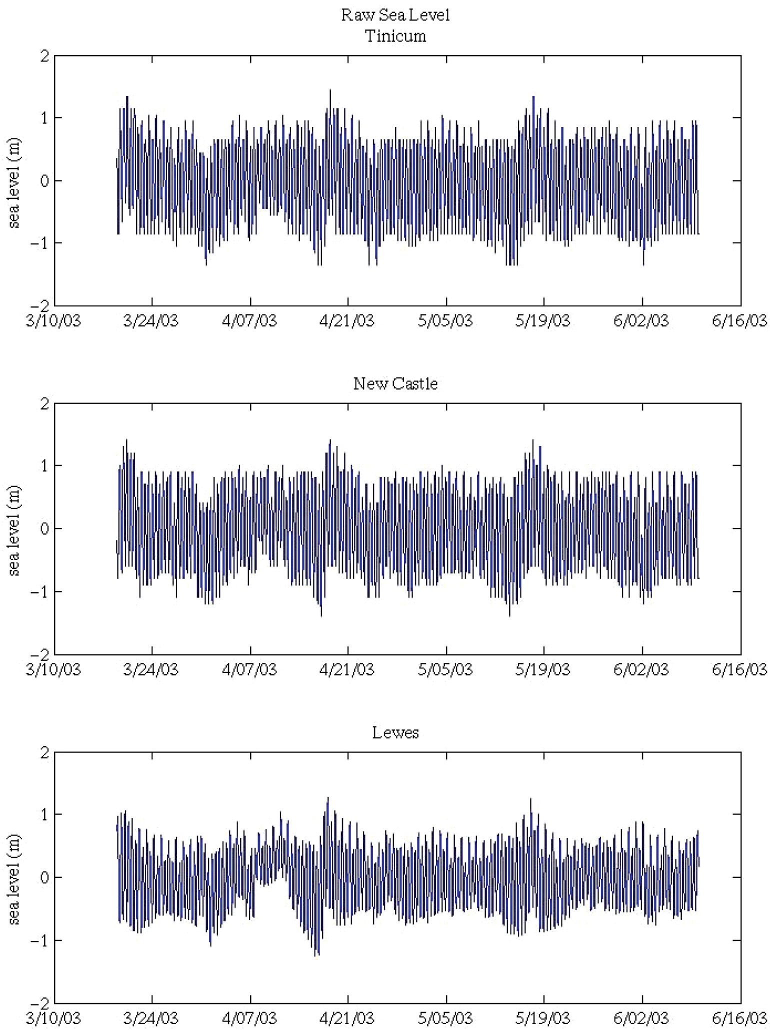

Figure 2. Time series of sea level data at Tinicum, New Castle, and Lewes. See Figure 1 for station locations and text for discussion.

The phase difference between tidal elevation and tidal current provides information on the characteristics of the tidal wave. The tidal wave is progressive when the elevation and current are in phase, and the tidal wave is standing when the two are in quadrature phase to each other. For a progressive tidal wave in a shallow estuary, the tidally averaged co-variation of surface elevation (η) and current (U) results in a positive Stokes transport $\langle(\eta + h)U\rangle$ up the estuary (in the direction of wave propagation). Here the angled brackets $\langle\,\rangle$ denote a temporal average over the tidal period. As a consequence of the positive Stokes transport, a negative Eulerian mean current is produced in order to maintain

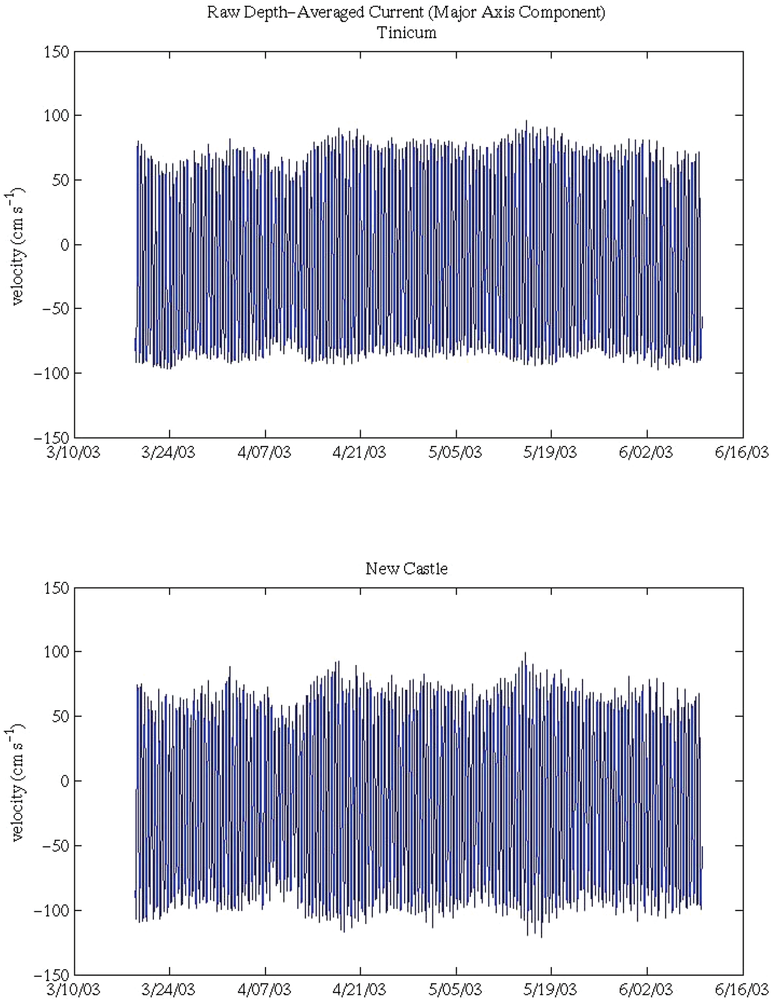

Figure 3. Time series of depth-averaged major axis component of current at Tinicum and New Castle. Positive current indicates up-estuary flow and negative current indicates down-estuary flow. Note the down-estuary mean flow averaged over the period of record.

a zero Lagrangian mean current, independent of river discharge. For a standing wave, the Stokes transport is zero due to the absence of tidally averaged co-variation in surface elevation and current. In the upper Delaware estuary, the tidal wave is neither purely progressive nor purely standing, as the phase difference between tidal elevation and tidal current is neither 0° nor 90°. However, the small phase difference between the M_2 tide and the M_2 current (10.6° at New Castle and 7.8° at Tinicum) indicates that the wave is mostly progressive in nature. This suggests the presence of up-estuary directed Stokes transport and a compensating down-estuary directed mean Eulerian current.

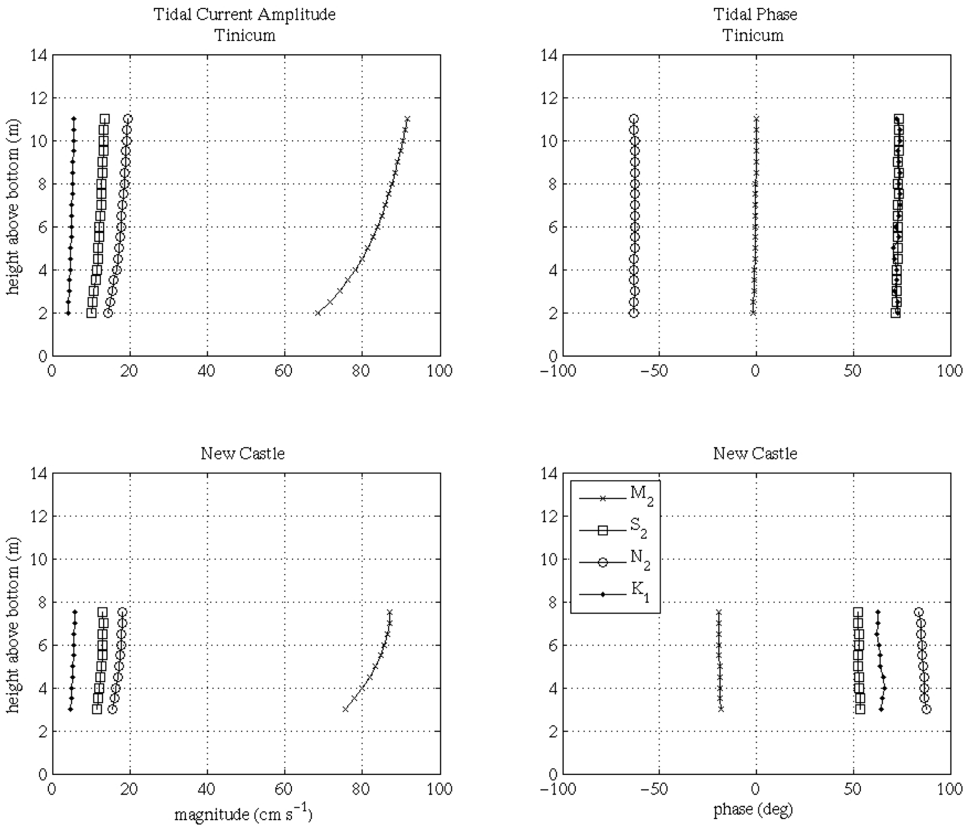

Figure 4. Vertical profiles of tidal current amplitude (left panels) and phase (right panels) for four leading tidal constituents at Tinicum and New Castle. See text for discussion.

To further examine the structure of the tidal currents, vertical profiles of the amplitude and phase of the four leading tidal constituents are presented in Figure 4. The harmonic analysis shows that the M_2 current is the dominant constituent at all depths. The amplitude of the M_2 current undergoes significant variation with depth, with higher values near the surface and lower values near the bottom, consistent with the effect of bottom friction on tidal current. There is very little variation of the M_2 current phase with depth, suggesting that the turning of the tidal current from flood to ebb occurs nearly simultaneously from surface to bottom.

4. The subtidal variability

Figure 5 shows the subtidal sea level fluctuations at Lewes, New Castle, and Tinicum. It is apparent that the magnitude of the subtidal sea level is about a factor of 2 weaker than that of the astronomical tides. The subtidal sea level in the interior of an estuary may be

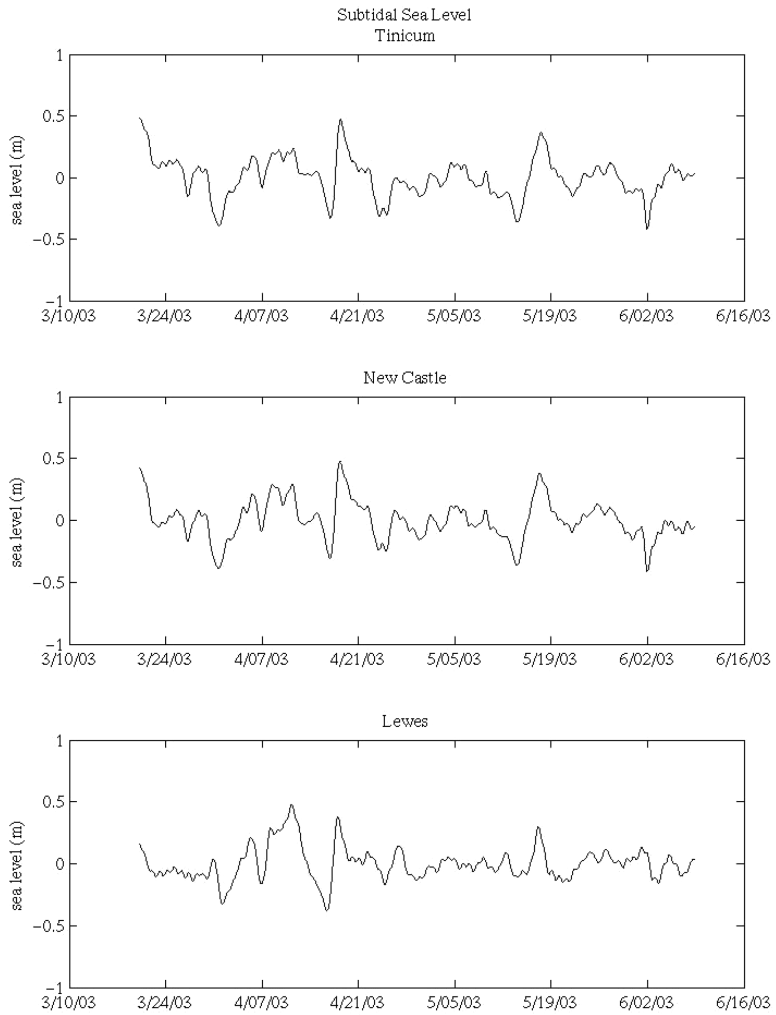

Figure 5. Time series of subtidal sea level at Tinicum, New Castle, and Lewes.

forced by a combination of the remote and local wind effect. Through the use of an analytical model, Garvine (1985) has shown that to lowest order, the remotely forced subtidal sea level in the interior of the estuary simply co-oscillates with the coastal sea level with no attenuation or phase lag. On the other hand, the local wind effect is primarily responsible for producing surface slopes along the estuary. The sea level in the upper Delaware estuary ($\eta_{\text{New Castle}}$ and η_{Tinicum}) can thus be represented as

$$\eta_{\text{New Castle}} = \text{remote wind effect} + \text{local wind effect} = \eta_{\text{Lewes}} + (\eta_{\text{New Castle}} - \eta_{\text{Lewes}}) \quad (1)$$

$$\eta_{\text{Tinicum}} = \text{remote wind effect} + \text{local wind effect} = \eta_{\text{Lewes}} + (\eta_{\text{Tinicum}} - \eta_{\text{Lewes}}). \quad (2)$$

Here the subtidal sea level at the mouth of the Delaware estuary, η_{Lewes}, is used to represent the remote wind effect. To examine the relationship between coastal sea level and wind from different directions, the correlation coefficient is computed between η_{Lewes} and wind components at 10°T intervals clockwise from north. The results show that the highest correlation ($r = -0.75$) occurs for wind centered around 30°T. The negative correlation indicates a drop in coastal sea level with a wind blowing toward the direction of 30°T. This direction is closely aligned with the orientation of the mouth of the Delaware estuary between Cape Henlopen and Cape May (Fig. 1). A wind blowing toward this direction corresponds to an upwelling-favorable wind, and a drop in coastal sea level is consistent with the across-shelf Ekman transport associated with this wind pattern.

Although coastal sea level responds preferentially to along-shelf wind measured at NBDC Buoy 44009, it is apparent that a sizable fraction of the coastal sea level is not correlated with the wind over the continental shelf adjacent to the Delaware estuary. The remainder of the coastal sea level fluctuation may come from the down-shelf (in the sense of Kelvin wave) propagation of free waves. These free waves may have been generated on the continental shelf far up-shelf from the Delaware estuary, and the coastal sea level fluctuation associated with these free waves may not be correlated with the winds measured over the inner shelf off Delaware.

To measure the relative importance of the remote and local wind effect in forcing the total subtidal sea level in the upper estuary, it is useful to examine the correlation between the interior sea level and coastal sea level. The correlation coefficient between η_{Lewes} and $\eta_{\text{New Castle}}$ is $r = 0.63$, and that between η_{Lewes} and η_{Tinicum} is $r = 0.56$. This indicates that, while the remote wind effect may account for more than half of the subtidal sea level variability in the upper estuary, there are still significant differences between the coastal and interior sea level (Fig. 5). To establish whether these sea level differences can be attributed to local wind forcing, the correlation between different wind components and sea level differences were computed. The sea surface slope along the estuary correlates negatively ($r = -0.5$) to wind blowing roughly toward the southeast (140°T) along the major axis of the lower and middle reaches of the estuary. The negative correlation coefficient indicates a local set-down occurs with down-estuary wind, consistent with the effect of local wind over the surface of the estuary.

Since the head of the estuary represents a closed physical boundary for the variability forced by both the remote and local wind effects, one would expect the subtidal current fluctuation to be linked to the subtidal sea level fluctuation via the continuity requirement of $AU = S\,\partial\eta/\partial t$, where A is the cross-sectional area at either New Castle or Tinicum, U is the current averaged over the cross-section, η is the surface elevation, S is the surface area of the upper estuary upstream of either New Castle or Tinicum, and t is time. Since the subtidal sea level at New Castle and Tinicum are highly correlated with each other ($r = 0.95$), it is not unreasonable to assume that the subtidal sea level farther up the estuary will be similar to those at New Castle and Tinicum. Assuming that the depth-averaged subtidal current is an adequate approximation for U, one would expect U to be well correlated with the time rate of change of surface elevation $\partial\eta/\partial t$. However, the correlation coefficient between the subtidal current U and the time rate of change of sea level $\partial\eta/\partial t$ is only $r = 0.37$ at New Castle, and the correlation coefficient decreases further to $r = 0.23$ at Tinicum.

This low correlation between subtidal current and the subtidal sea level suggests that some other mechanism might play a significant role in forcing the current. Yet another explanation lies with the nature of wind forced variability in current and sea level. Since the subtidal current U is scaled by $\partial\eta/\partial t$, not by η, the magnitude of the subtidal current will decrease with decreasing frequency (or increasing time scale) even if the magnitude of the subtidal sea level remains the same. A 1-m rise in sea level over a time scale of 10 days would therefore yield a much weaker inflow compared to the same rise in sea level over a time scale of 2 days. This scaling argument suggests that the importance of remote wind in forcing subtidal current diminishes with increasing time scale. The presence of other mechanisms may overwhelm the effect of wind at these long time scales. It is interesting to explore this further by dividing the subtidal variability in current according to time scale. The subtidal current will therefore be represented as a linear superposition of current at intermediate (between 36 hr and 120 hr) and long (greater than 120 hr) time scales as

$$U\ (\mathrm{T} > 36\ \mathrm{hr}) = U\ (120\ \mathrm{hr} > \mathrm{T} > 36\ \mathrm{hr}) + U\ (\mathrm{T} > 120\ \mathrm{hr}) \tag{3}$$

Here T stands for time scale. U (T $>$ 36 hr) is the low-passed subtidal current obtained by passing the raw data into a Lanczos filter with a cut-off period of 36 hr. U (T $>$ 120 hr) is the low-low-passed current obtained by passing the raw data through the filter with a cut-off period of 120 hr, and U (120 hr $>$ T $>$ 36 hr) represents the band-passed data obtained by subtracting the low-low-passed data from the low-passed data (Fig. 6). The same decomposition can be done on the subtidal sea level η and the time rate of change of sea level $\partial\eta/\partial t$.

At time scales between 36 hr and 120 hr the band-passed current and $\partial\eta/\partial t$ are significantly correlated with each other at New Castle ($r = 0.84$) and Tinicum ($r = 0.73$). This suggests that wind is effective in forcing the subtidal current fluctuation over the intermediate subtidal time scales of 36 hr to 120 hr. At longer time scales of T $>$ 120 hr the current is poorly correlated with $\partial\eta/\partial t$, with $r = -0.13$ at New Castle and $r = -0.08$ at Tinicum. This suggests that some other mechanism is responsible for forcing the current variability at these long time scales. This is a matter of some importance, as there is stronger variability in the T $>$ 120 hr current than the 120 hr $>$ T $>$ 36 hr current. At Tinicum the standard deviation of U (T $>$ 120 hr) is 4.76 cm s^{-1}, and that of U (120 hr $>$ T $>$ 36 hr) is only 1.57 cm s^{-1}. At New Castle the standard deviation of U (T $>$ 120 hr) is 2.94 cm s^{-1}, and that of U (120 hr $>$ T $>$ 36 hr) is 1.67 cm s^{-1}.

River discharge represents one possible forcing mechanism at long time scales. Figure 7 shows the daily mean discharge of the Delaware River gauged at Trenton, New Jersey and that of the Schuylkill River gauged at Philadelphia, Pennsylvania. The discharge of the Schuylkill is roughly a quarter of that of the Delaware, and both of these rivers enter the

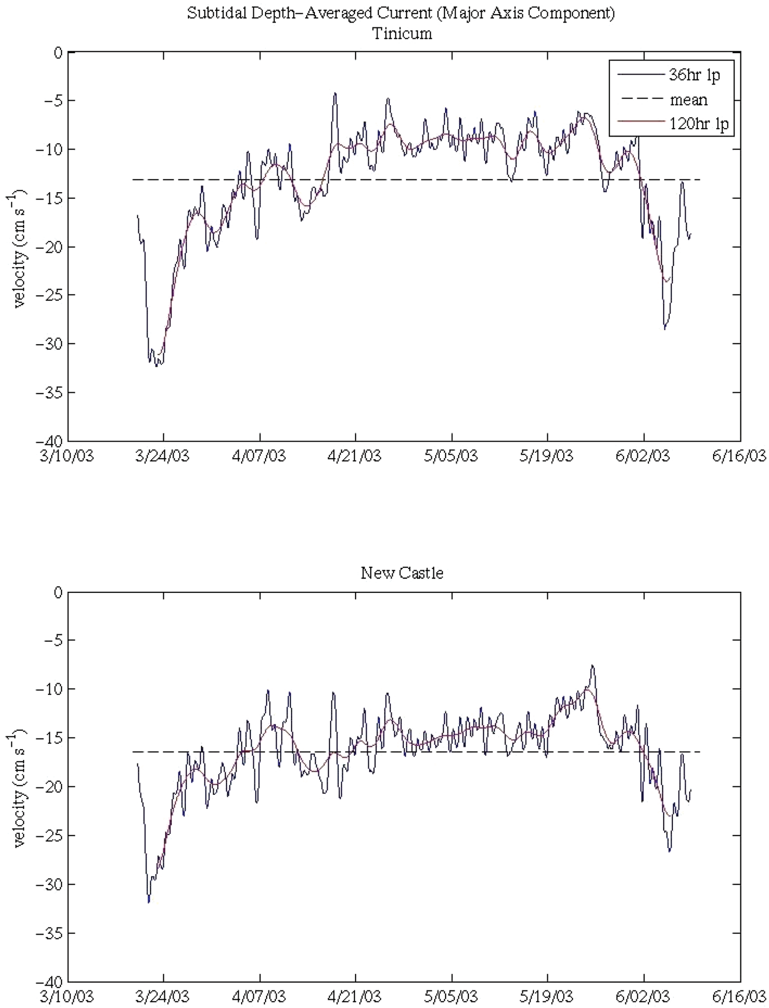

Figure 6. Time series of depth-averaged subtidal current with time scales longer than 36 hr and depth-averaged subtidal current with time scales longer than 120 hr at Tinicum and New Castle. The dashed line gives the mean current. Positive current indicates up-estuary flow and negative current indicates down-estuary flow.

Delaware estuary upstream of New Castle and Tinicum. The combined discharge from these two rivers (R) will therefore be used to assess its influence on the subtidal current. The vast majority of the variability in the combined river discharge operates over long time scales. The standard deviation of R (T $>$ 120 hr) is 423.0 $m^3\ s^{-1}$, but the standard deviation of R (120 hr $>$ T $>$ 36 hr) is only 33.6 $m^3\ s^{-1}$. The focus will, therefore, be placed on the effect of river discharge on the current at time scales longer than 120 hr.

The low frequency variation in the depth-averaged current, U (T $>$ 120 hr), follows the

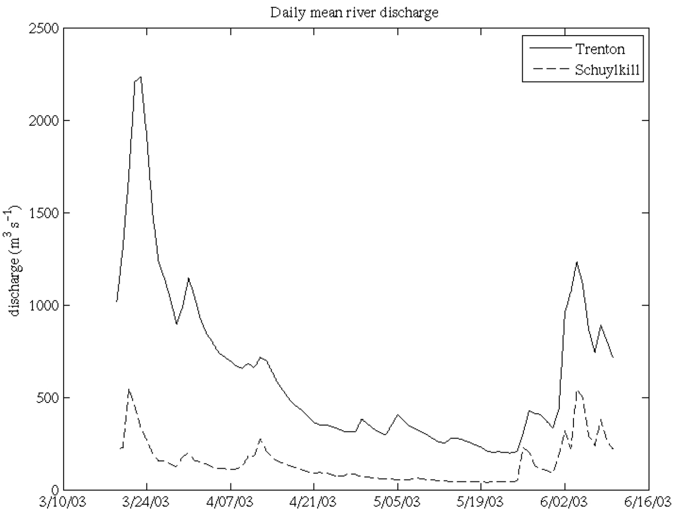

Figure 7. Daily mean discharge of the Delaware and Schuylkill rivers during the study period. Discharge peaks at the beginning and end of the time series produced significant subtidal currents at the Tinicum and New Castle ADCP sites (see Fig. 6).

variation in discharge, R (T $>$ 120 hr), very closely. The correlation coefficient between current and discharge is $r = -0.96$ at Tinicum and $r = -0.87$ at New Castle. The negative correlation indicates that an increase in the river discharge is associated with an increase in the negative current flowing down the estuary. To further explore the relationship between river discharge and depth-averaged current, a linear regression is conducted between the two (Fig. 8). At Tinicum the linear regression indicates a down-estuary current of -0.0011 cm s^{-1} per each m^3 s^{-1} of river discharge. At New Castle the regression also indicates that a river discharge produces a down-estuary current, but with a reduced magnitude of -0.00061 cm s^{-1} per each m^3 s^{-1} of river discharge. The decrease in the magnitude of the river-induced subtidal current fluctuation at New Castle is most likely a direct result of the increased cross-sectional area there relative to the situation at Tinicum.

From the standpoint of the river discharge, the head of the estuary is not a physical barrier, as the discharge of the Delaware River can flow over the falls and then into the estuary. The upper estuary should, therefore, not be treated as a semi-enclosed basin when one considers the effect of river discharge on the current. The barotropic response of the estuary to the river discharge should generate a down-estuary flowing volume flux that equals the river discharge. Since the river discharge represents a through-flow situation, the continuity requirement for the upper estuary between New Castle and Tinicum becomes $(A_1U_1 - A_2U_2) = S_{12}\, \partial\eta/\partial t$. Here A_1 is the cross-sectional area at New Castle, U_1 is the sectionally averaged current at New Castle, A_2 is the cross-sectional area at Tinicum, U_2 is the sectionally averaged current at Tinicum, S_{12} is the surface area between New Castle

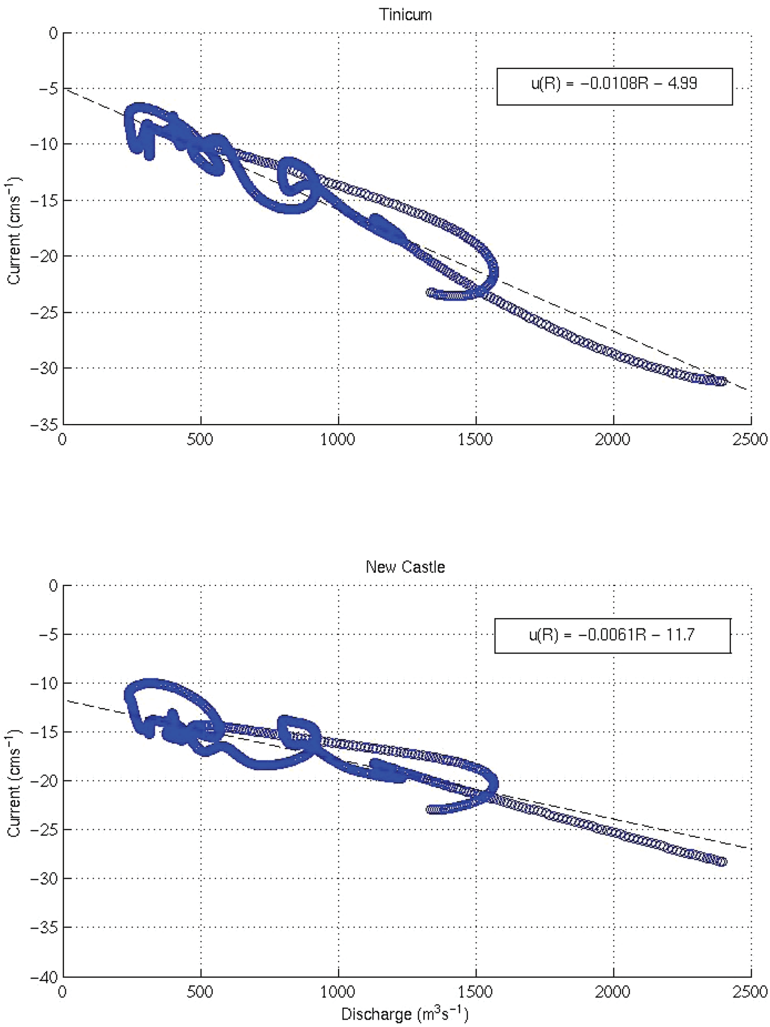

Figure 8. Scatter plot of the depth-averaged subtidal current versus the river discharge at time scales longer than 120 hr. The dashed line represents the linear regression between the two. Top panel is for Tinicum and lower panel is for New Castle.

and Tinicum, η is the surface elevation there, and t is time. For a through-flow situation, it is possible for the river discharge to generate significant current, but as long as $A_1U_1 \approx A_2U_2 \approx$ R, the river discharge does not necessarily have to generate a change in surface elevation. This is perhaps why U and $\partial\eta/\partial t$ are so poorly correlated at time scales longer than 120 hr. Of course the situation can be further complicated if there is a baroclinic response to the river discharge. Garvine *et al.* (1992) have shown that the mean salt intrusion limit extends only 97 km upstream of the estuary's mouth. New Castle is located

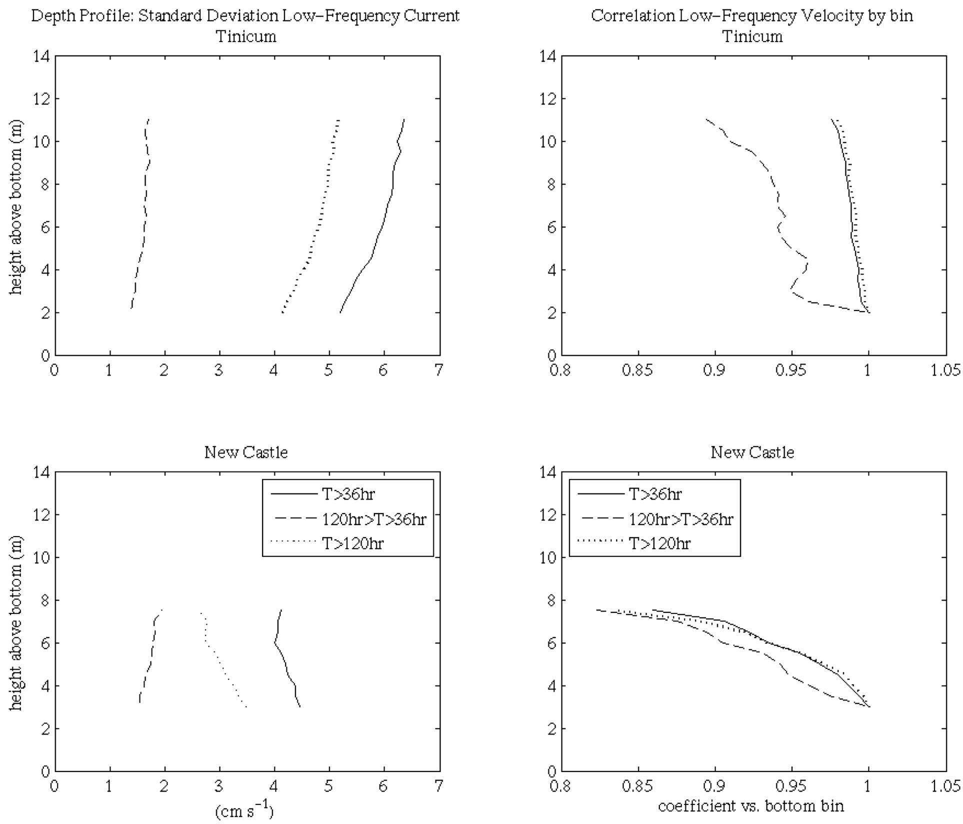

Figure 9. Left panels show vertical profiles of the standard deviation of the subtidal current over time scales of T > 36 hr, 120 hr > T > 36 hr, and T > 120 hr at Tinicum (top) and New Castle (bottom). Right panels show the correlation between currents at different depths and current near the bottom.

104 km upstream of the mouth, and Tinicum is located another 32 km up-estuary from New Castle. While it is possible for the salt intrusion limit to extend beyond New Castle under low discharge conditions, the salt intrusion almost never reaches Tinicum. The study period of March–June 2003 corresponds to the spring freshet with above average river discharge (Cook *et al.,* 2007). It is not likely for the salt to intrude into the upper estuary in any significant manner, so the potential for a strong baroclinic response to river discharge is low.

To further explore the characteristics of the subtidal currents at different time scales, Figure 9 shows the vertical profile of the standard deviations of currents with time scales of T > 36 hr, 120 hr > T > 36 hr, and T > 120 hr. At Tinicum it is clear that the majority of the subtidal current variability resides in the T > 120 hr band. This current, largely induced by the river discharge, is much stronger than the wind-induced current operating over the time scale of 36 hr to 120 hr. The subtidal current fluctuations at all time scales show decreasing magnitudes with decreasing distance away from the bottom, consistent with the effect of bottom friction. Figure 9 also shows that the subtidal currents are well correlated throughout the water column, with $r > 0.9$ between currents near the bottom and those at other depths. All evidence thus points to a barotropic current with in-phase fluctuations from surface to bottom. At New Castle the magnitude of the river-induced current fluctuation is reduced due to the increase in the cross-sectional area there, but it is still stronger than the wind-induced current. The wind-induced current at New Castle shows a decrease in magnitude with depth, similar to the situation in Tinicum. However, the river-induced current fluctuation shows a slight decrease in magnitude toward the surface, although the cause of this anomalous behavior is unclear.

5. The mean flow

One of the most prominent features of the subtidal currents is the strong mean flow at both observational sites. At Tinicum the depth-averaged mean current is $Um = -13.41$ cm s^{-1}, and that at New Castle is $Um = -16.47$ cm s^{-1}. Part of the mean flow may be induced by the river discharge, but other mechanisms can also play a role in forcing a mean current. The linear regression shown in Figure 8 can be used to estimate the magnitude of the river-induced mean current as well as the magnitude of the mean current from nonriver mechanisms such as tidal rectification. For example, the regression line for the Tinicum data has a slope of -0.0011 cm s^{-1} per m^3 s^{-1} and an intercept of -4.99 cm s^{-1}. The intercept is an estimate of the current when the river discharge is zero, so it represents the mean current generated by mechanisms other than the river (Um-nr). On the other hand, the mean current induced by the river discharge (Um-r) can be estimated by multiplying the slope to the mean river discharge (680 m^3 s^{-1}), yielding a value of -7.48 cm s^{-1}. The sum of these two estimates, $Um\text{-}r + Um\text{-}nr = -12.47$ cm s^{-1} comes within 1 cm s^{-1} of the observed depth-averaged mean current. The fact that Um-r $>$ Um-nr suggests that the river discharge is larger than the nonriver mechanisms in forcing the mean flow at Tinicum. One likely candidate for the nonriver mean flow is a tidally rectified current, in other words, a seaward Eulerian current induced by Stokes wave transport during flood tide. Based on the work of Ianniello (1977, 1979), the magnitude of the tidally rectified mean current should be of the order $(\eta_0/h)U_0$, where η_0 is the amplitude of the first order tidal elevation, h is the water depth, and U_0 is the amplitude of the first order tidal current. For the dominant M_2 tide at Tinicum, η_0 is 0.74 m, h is 10 m, and U_0 is 83.2 cm s^{-1}, so the tidally rectified mean current should be of the order -6 cm s^{-1}. It should be cautioned that this is only an order of magnitude estimate, not an exact computation of the tidally rectified mean current. Nevertheless, this result suggests that tidal rectification is a strong candidate for the part of the mean current forced by nonriver sources.

The same procedure can be repeated for New Castle to obtain a river-induced mean current of -4.14 cm s^{-1} and a nonriver mean current of -12.47 cm s^{-1}. Once again the sum of these two estimates comes within 1 cm s^{-1} of the observed depth-averaged mean current at New Castle. At New Castle the effect of the river discharge on the mean current is reduced due to the increase in cross-sectional area relative to the situation farther up-estuary at Tinicum. As a result, the river-induced mean current is weaker than the nonriver, tidally rectified mean current at New Castle. The significance of this current lies in its potential to sweep dissolved substances and fine particulate matter from the upper to lower estuary. New Castle falls near the head of the estuarine turbidity maximum, a trapping zone for suspended sediments and particle-borne contaminants (Cook *et al.,* 2007). In conjunction with river discharge, the tidally rectified current drives a down-estuary flux of fine sediment from the upper estuary to the lower estuary and its fringing wetland coast.

To examine the vertical profiles of the mean current, linear regression is performed between the river discharge and the current at each bin in order to estimate the river and nonriver-induced mean current as a function of depth (Fig. 10). At Tinicum both the river and nonriver-induced mean currents are negative at all depths, indicating a unidirectional flow down-estuary at all depths. Both the river and nonriver-induced mean currents show a reduction in magnitude with depth. It is important to note that the ADCP provides the vertical mean flow profile at only one point along the cross section of the upper Delaware estuary at either Tinicum or New Castle. The fact that the vertical profiles show surface to bottom outflow does not mean that the residual current flows down-estuary over the entire cross section.

6. Origin of tidally rectified residual flows in estuaries

Li and O'Donnell (2005) have developed an analytic model to examine the tidally rectified residual circulation in tidally dominated estuaries of different lengths and lateral depth variations. They found that the response of the estuary depends on an important parameter $\delta = 4\ L/\lambda$, which is the ratio between the length of the estuary (L) and one quarter of the tidal wavelength ($\lambda = (gh)^{1/2}/\omega$, where g is the acceleration of gravity, h is depth, and ω is tidal frequency). For an estuary with a δ value smaller than 0.6–0.7 (short estuary), the rectified current at the mouth shows inward (up estuary) transport in deep water and outward (down estuary) transport in shallow water. The situation is just the opposite for an estuary with a δ value greater than 0.6–0.7 (long estuary), where up-estuary flow at the mouth occurs in shallow water and down estuary flow occurs over deep water. Li and O'Donnell (2005) argued that the parameter δ is important because tidal characteristics are different between long and short estuaries, as tide behaves like a progressive wave in a long estuary and like a standing wave in a short estuary. They showed that the lateral structure of the tidally rectified residual current gives inflow over the shoals and outflow over deep water when the tidal wave is close to a progressive wave. The opposite occurs when the tidal wave is close to a standing wave.

For the case of the Delaware estuary with a mean depth of $h = 10$ m and a length of $L = 215$ km, the length ratio for the dominant semi-diurnal tide is about 2, so the

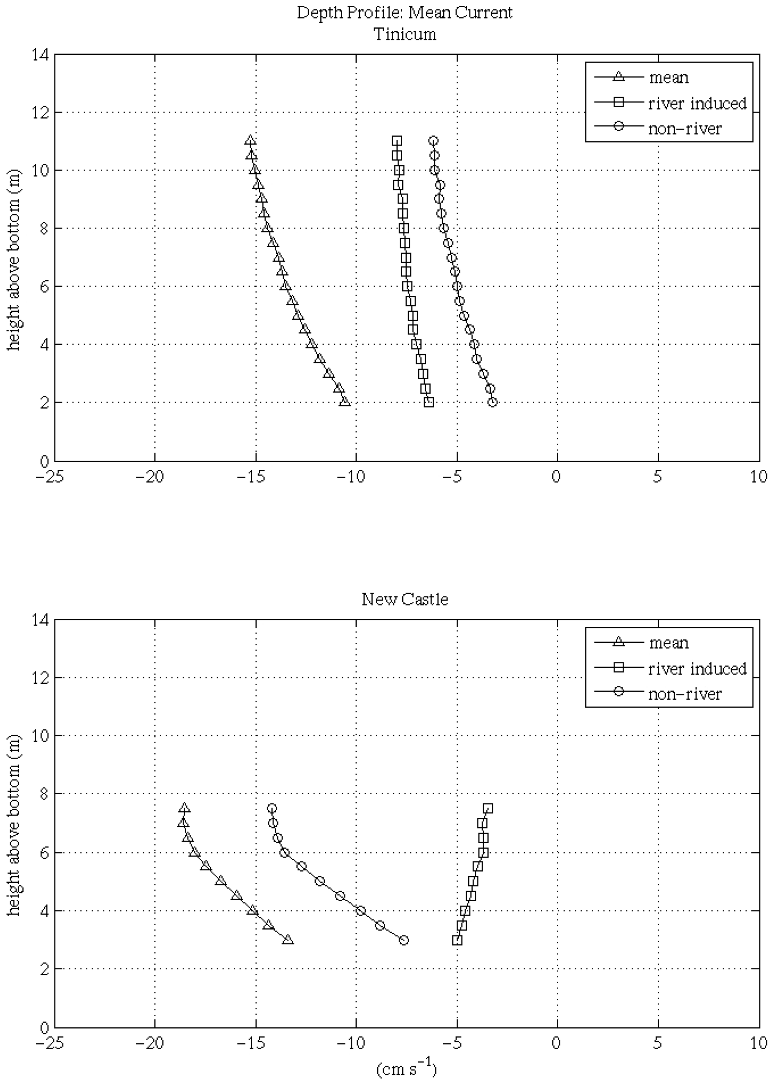

Figure 10. Vertical profiles of the observed mean current and the estimated mean current induced by river discharge and nonriver mechanisms. The top panel is for Tinicum and the bottom panel is for New Castle. See text for discussion.

Delaware should behave like a long estuary in response to tidal forcing. The fact that the semi-diurnal tidal elevation and tidal current are almost in phase at both Tinicum and New Castle shows that the tidal wave is close to a progressive wave in the study area. The work of Li and O'Donnell (2005) thus suggests that the tidally rectified residual flow should be directed down the Delaware estuary over the deep channel and up the estuary over the shallow shoals flanking the channel. This pattern is consistent with the observed nonriver-induced mean outflow over the channel at Tinicum.

The situation at New Castle is more complicated because the ADCP was located at the western edge of the channel. The study of Li and O'Donnell (2005) suggests relatively weak residual current in this transition zone between outflow over the channel and inflow over the shoals, but the observed nonriver-induced mean flow there is strong and decidedly down estuary. The width of the Delaware estuary increases by more than a factor of 2 from Tinicum to New Castle. It is possible that one has to consider the Coriolis effect on the tidally rectified mean flow when the estuary becomes wider. Winant (2008) has developed an analytical model to describe the three-dimensional Eulerian and Lagrangian mean current in basins where the weakly nonlinear viscous flow is forced by the tide under the f-plane approximation. He found that the Coriolis effect can drastically change the residual circulation pattern under weak to moderate friction, with inflow on the right side of the basin as one looks up the estuary and a return outflow along the opposite bank. Under strong friction the pattern of the tidally rectified mean flow is consistent with that of Li and O'Donnell (2005). The effect of friction on the current is evident in the Delaware estuary in that the magnitudes of the currents (at both tidal and subtidal frequencies) decrease with decreasing distance away from the bottom. However, the effect of friction is only moderate because there is no evidence of a significant attenuation of sea level and current between the mouth and the upper estuary. The observed strong nonriver-induced outflow at the left edge of the channel at New Castle is thus consistent with the tidally rectified mean flow under the effect of rotation.

There are a number of factors that can affect the generation of the tidally rectified current. These include the amplitude of the tidal current, the depth ratio (amplitude of tidal elevation over depth), and the progressive or standing nature of the tidal wave. However, there are no significant differences in these factors between New Castle and Tinicum, yet the tidally rectified current at New Castle is a factor of 2 stronger than that at Tinicum. One possible explanation lies in the sharp bend in the estuary near New Castle which tends to enhanced velocity gradients. Ianniello (1977, 1979) has indicated that the advective terms such as $u\ \partial u/\partial x$ in the momentum equation is important for the generation of tidally rectified currents. Comparing with the relatively straight coastline near Tinicum, the sharp bend in the estuary near New Castle thus favors the generation of stronger rectified current. It is important to note, however, that the vertical current profile was measured at only one point along the estuary's cross section, so the differences between the two sites could be influenced by the placement of the ADCP along different parts of the cross sections and the exact nature of the lateral distribution of the rectified current across the estuary.

The tidally rectified mean current is barotropic in nature. The mean flow associated with the river input (Um-r) may have both a barotropic (u_R) and a baroclinic (u_ρ) component. The baroclinic circulation depends upon the existence of the horizontal density gradient. Officer (1976) has examined the vertical profiles of the barotropic and baroclinic mean currents that arise from the river input and showed that the baroclinic component has a two-layer structure with a surface outflow and a bottom inflow, and the magnitude of the current is proportional to the longitudinal density gradient. Wong (1994) has extended Officer's (1976) analysis by examining the lateral variability in the river-induced baroclinic circulation and found up-estuary flow over the channel and down-estuary flow over the shallow areas along the shores. The observed pattern of the river-induced mean flow shows no evidence of a reversal with depth or an up-estuary inflow over the channel. The absence of a gravitational circulation in the upper Delaware estuary is consistent with the fact that the salt intrusion limit usually falls seaward of the New Castle area during periods of elevated river discharge.

7. Summary

The tidal and subtidal variability in the upper Delaware estuary has been examined based on a set of current and sea level measurements collected during March–June 2003. At short (semi-diurnal and diurnal) time scales, the astronomical tides within the Delaware estuary are forced by the tide on the continental shelf off the mouth of the estuary. Results obtained from harmonic analysis show that the tidal variability is dominated by the M_2. The amplitude of the M_2 tide increases by about 20% from the mouth to the upper estuary, reflecting the influence of the decreasing cross-sectional area with distance up the estuary. The M_2 tide is near progressive in nature, with a small phase lag between the tidal current and tidal elevation. This suggests the generation of an up-estuary Stokes transport and a down-estuary, tidally rectified, residual Eulerian current in compensation.

At longer time scales the subtidal sea level variability in the upper estuary is forced primarily by a combination of the remote and local wind effect. The remote wind effect forces the interior of the estuary through the impingement of the costal sea level at the mouth of the estuary. This coastal sea level fluctuation is largely forced by the across-shelf Ekman transport associated with the upwelling/downwelling winds over the continental shelf adjacent to the estuary. However, the down-shelf propagation of free waves may also play a role in determining the coastal sea level off the mouth of the Delaware estuary. The local wind effect represents the frictional coupling of the wind along the major axis of the estuary. It is responsible for producing surface slopes along the estuary. The wind-forced sea level is well correlated with the current at time scales of 36 to 120 hr, indicating the importance of wind on the transport process over these intermediate time scales. The wind-forced current shows coherent, in-phase fluctuation throughout the water column.

At time scales longer than 120 hr the current fluctuation is primarily forced by the variation in the river discharge. The river-induced current fluctuation over these long time scales can be more than a factor of 3 stronger than that produced by the wind over the intermediate time scales. One of the most prominent features of the subtidal current time series is the strong mean flow. The present study shows that river discharge is the primary mechanism for generating the observed mean current at Tinicum, 60% of the total current. All evidence indicates that the river-induced current represents a barotropic response to river discharge, resulting in a unidirectional flow down-estuary from surface to bottom. In addition to the river discharge, tidal rectification appears to produce the remainder (40%) of the Eulerian mean current measured at Tinicum.

At New Castle, river discharge remains an important factor in forcing the mean flow and the subtidal current fluctuation over time scales longer than 120 hr. However, in contrast to Tinicum, the seaward Eulerian mean current induced by Stokes Drift dominates the mean flow (75% of the total current). River discharge produces a weaker current at New Castle, because the discharge must pass through a larger cross-sectional area relative to the channel at Tinicum. Despite the fact that New Castle is only a short distance upstream of the mean salt intrusion limit, there is no evidence of a baroclinic response to river discharge, and the river-induced mean current is down-estuary at all depths in the water column.

To sum up, the physical exchange process in the upper Delaware estuary is forced by the combined effect of the tide, the wind, and the river discharge. Of the three, the first-order tide generates the strongest exchange. Importantly, the semi-diurnal tide generates a significant rectified mean flow which rivals or even exceeds the barotropic mean flow induced by the river discharge. A number of factors, including the strong tidal current, the large ratio of tidal amplitude versus depth, and the near-progressive nature of the tidal wave, combined to favor the generation of the rectified mean current, which has major implications for transporting dissolved materials and particulates from the upper to lower estuary. In comparison, wind is by far the weakest among the three mechanisms in producing subtidal currents in the upper Delaware estuary. The remote effect of wind forces significant subtidal variability in sea level, but the wind-forced subtidal current decreases with decreasing frequency so it cannot compete with either the tidally rectified current or the river-induced barotropic current at long time scales. Note that the measurements of longitudinal flow presented here represent only two points in the estuary, and that the data provide no information about the lateral variability in current. A more comprehensive survey is required to resolve the three-dimensional variability in the upper Delaware estuary.

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