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Hydrodynamics of the Hudson River Estuary

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Abstract.—The Hudson River Estuary can be classified as a drowned river valley, partially mixed, tidally dominated estuary. Originally, it had a fjord-like morphology as a result of glacial scour which filled in over the past 3,000 years with river sediments. The hydrodynamics of the estuary are best described by the drivers of circulation, including the upstream river inflows, the oceanographic conditions at the downstream end, and meteorological conditions at the water surface and the response of the waters to these drivers in terms of tides and surges, currents, temperature, and salinity. Freshwater inflow is predominantly from the Mohawk and Upper Hudson rivers at Troy (average flow = 400 m^3/s, highest in April, lowest in August). At the downstream end at the Battery the dominant tidal constituent is the semidiurnal, principal lunar constituent (the M_2 tide), with an evident spring/neap cycle. The amplitude of the tide is highest at the Battery (67 cm), lower at West Point (38 cm), and higher again at Albany (69 cm), a function of friction, geometry, and wave reflection. Meteorological events can also raise the water surface elevation at the downstream end and propagate into the estuary. Freshwater pulses can raise the water level at the upstream end and propagate downstream. Tidal flows are typically about an order of magnitude greater than net flows. The typical tidal excursion in the Hudson River Estuary is 5–10 km, but it can be as high as 20 km. Temperature varies seasonally in response to atmospheric heating and cooling with a typical August maximum of 25°C and January-February minimum of 1°C. Power plants cause local heating. The salinity intrusion varies with the tide and amount of upstream freshwater input. The location of the salt front is between Yonkers and Tappan Zee in the spring and just south of Poughkeepsie in the summer. Vertical salinity stratification exists in the area of salt intrusion setting up an estuarine circulation. The effect of wind is limited due to a prevailing wind direction perpendicular to the main axis and the presence of cliffs and hills. Dispersive processes include shear dispersion and tidal trapping, resulting in an overall longitudinal dispersion coefficient of 3–270 m^2/s. The residence or flushing time in the freshwater reach is less than 40 d in the spring and about 200 d in the summer. In the area of salt intrusion, it is about 8 d.

Introduction

The Hudson River Estuary extends from Troy to the Battery at the southern tip of Manhattan Island, as shown in Figure 1. Strictly speaking, in order to meet the definition of an "estuary" a water body has to contain seawater "measurably diluted with freshwater derived from land drainage" (Pritchard 1967). The upper reaches of this estuary (from Troy to Poughkeepsie) are always fresh and therefore that part of the system is technically not an "estuary" but a "tidal river." However, in this paper we concern ourselves little with that distinction and simply refer to the whole water body as the Hudson River Estuary.

The Hudson River Estuary was created about 6,000 years ago when rising sea level flooded the lower portions of the Hudson River with ocean water. Originally, the estuary had a fjord-like morphology as a result of glacial scour. About 18,000 years ago, the Laurentide glacier retreated northward leaving behind a deep gouge in the bedrock that was filled with melt water. Substantial quantities of river sediments were brought into the gouge over the next 3,000 years, altering the morphology so that today the Hudson River Estuary can be classified as a drowned river valley (McHugh et al. 2004). Along its 247-km length (from Troy to the Battery), the geometry is extremely variable reflecting its geological past. The river contains wide shallow bays (e.g., Newburgh Bay), narrow deep channels (e.g., World's End), islands (e.g., Esopus Island), peninsulas (e.g., Croton Point), coves (e.g., Foundry Cove) and numerous other features (tidal flats, shoals, and rock outcroppings) that affect how water moves through the estuary. The Hudson River Estuary is a partially mixed estuary, where salt water and freshwater mix, resulting in a significant vertical density gradient. The currents in the estuary are predominantly driven by the tide. Meteorological events and freshwater inflows also play an important role in affecting the circulation.

The hydrodynamics of the Hudson River Estuary are of major importance to fish and fishery studies for many reasons. Most notably, in their early life stages, fish are planktonic and their movements are controlled by the ambient currents. Hydrodynamic processes lead to transport, dispersion, mixing, and flushing, important elements to consider when undertaking studies of the life histories of fishes and other aquatic organisms.

This paper describes the hydrodynamics of the Hudson River Estuary. It is written for the nonhydrodynamic professional and therefore attempts to describe many of the phenomena in nontechnical terms. The vocabulary of hydrodynamics is used where appropriate as an instructional aid. The paper starts with a description of the external drivers of the circulation at the upstream and downstream ends of the estuary and at the water surface. Then the responses of the estuary to these drivers are discussed, including tides and surges, currents, and temperature and salinity distributions. Finally, the processes that transport and mix salt, heat and any introduced pollutants are described.

Drivers of the Hydrodynamics

The hydrodynamics of the Hudson River Estuary are primarily driven by upstream inflows (mostly at Troy), oceanographic conditions at the downstream end (at the Battery), and meteorological conditions at the water surface. Following is a description of each of these drivers.

Upstream End

Upstream inflows are important mainly for

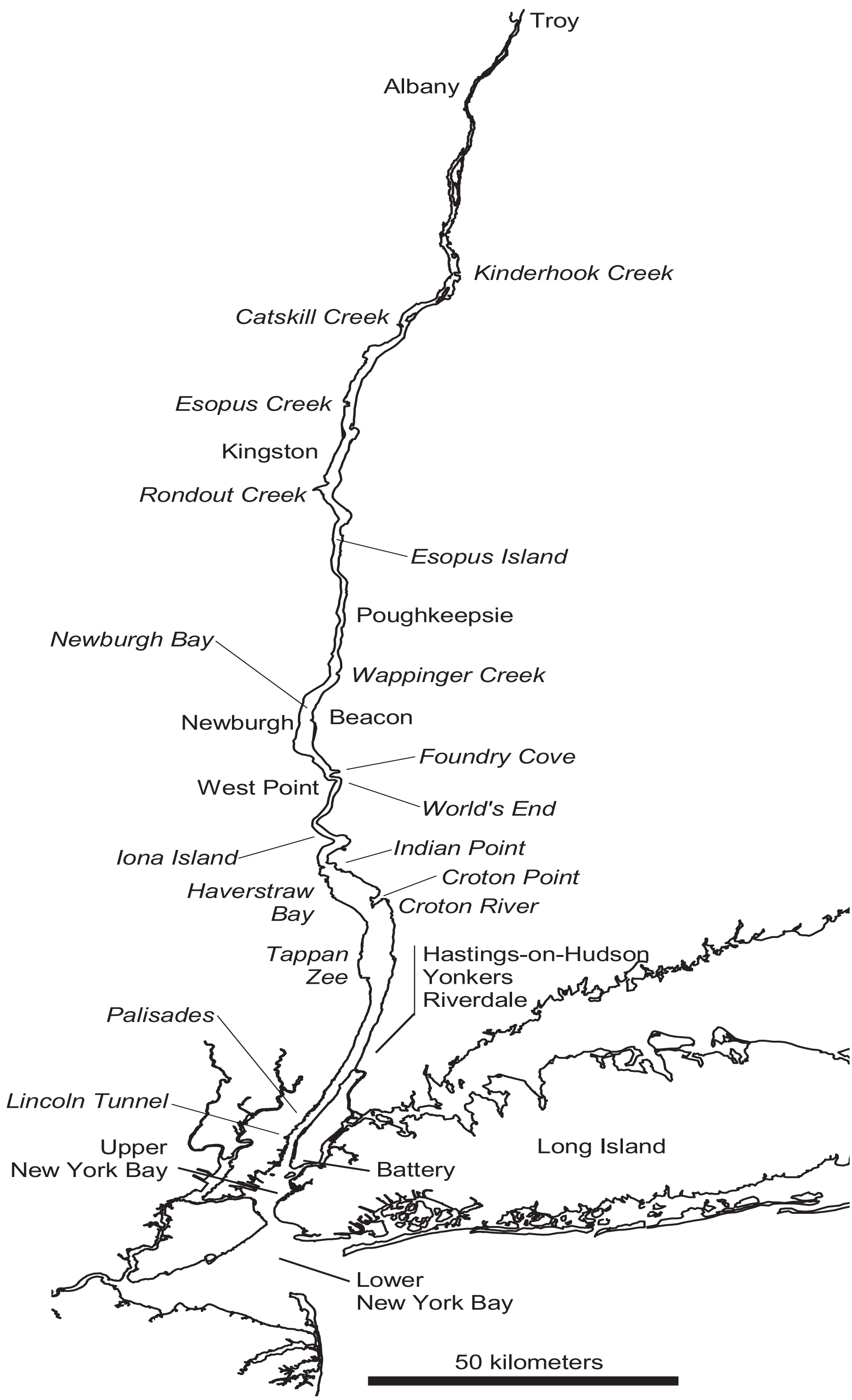

Figure 1. Overview map of the Hudson River Estuary.

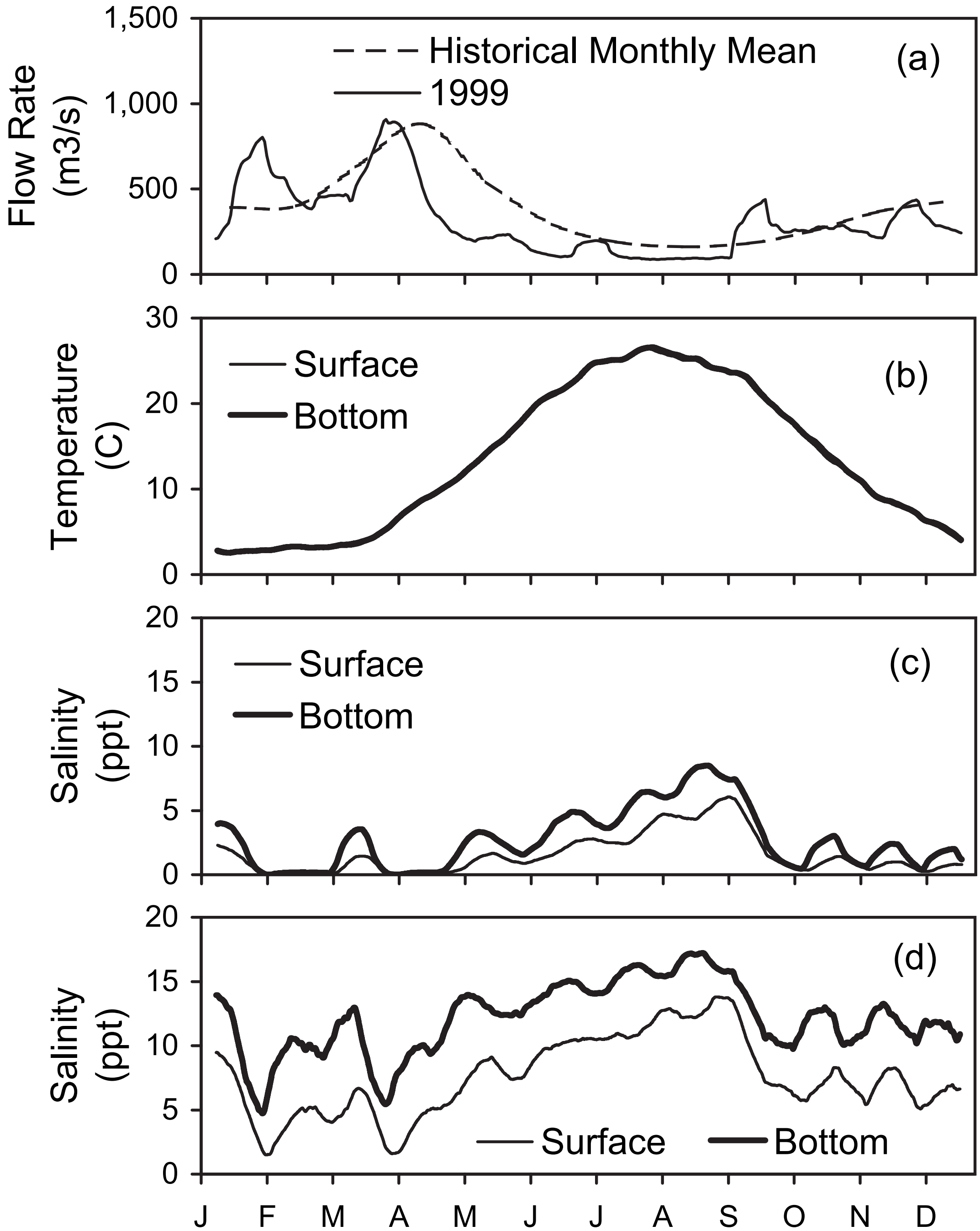

Figure 2. (a) Freshwater flow at Troy, New York (monthly means are based on the USGS Green Island gauge, period of record: 1946–2002), (b) surface and bottom temperature at Croton-on-Hudson, (c) surface and bottom salinity at Croton-on-Hudson, and (d) surface and bottom salinity at the Lincoln Tunnel. The temperature and salinity information are from the model of Blumberg et al. (1999) using forcing data from 1999. All data are "filtered" to remove oscillations occurring on time scales of less than 15 d.

two reasons. First, the added water increases the water surface elevation at the upstream end. This rise in elevation pushes water downstream (via the pressure gradient force). Second, the added water is fresh and the continuous input of freshwater prevents the salt water from flowing all the way up the estuary. Freshwater inflow is predominantly from the Mohawk and Upper Hudson rivers, which jointly enter the estuary from the north at Troy (average inflow of about 400 m^3/s). Other tributaries (see Figure 1 for locations of the inflow tributaries) include Rondout Creek (50 m^3/s), Kinderhook Creek (20 m^3/s), Esopus Creek (20 m^3/s), Catskill Creek (20 m^3/s), Croton River (10 m^3/s), and Wappinger Creek (10 m^3/s). Figure 2a shows the historical monthly average flows and the flows from 1999 at Troy (the other panels in the figure will be discussed subsequently). The inflow has a strong seasonal signal with highest flows in April and lowest flows in August. In any given year, the flow can vary significantly from the long-term mean. For example, in 1999, there were two peaks in the spring, one in late January and one in late March/early April. The summertime low flows in 1999 were considerably less than the long-term average. The freshwater inflows, the reader will learn later using Figures 2c and 2d, have a significant impact on the salinity distribution in the estuary. For a more detailed discussion on freshwater inputs to the Hudson River Estuary refer to Abood et al. (1992) and Wells and Young (1992).

Downstream End

The downstream end of the estuary is at the Battery where the Hudson River meets the upper part of the New York/New Jersey Harbor Estuary. Changes in the water surface elevation at this location are the most important drivers of water movement in the Hudson River Estuary. When the water surface elevation at the downstream end rises (e.g., high tide), water is pushed upstream into the estuary. Conversely, when it falls (e.g., low tide), water is pulled out of the estuary. The mechanism is the same as that responsible for the net downstream flow due to freshwater inputs at the upstream end (pressure gradient force).

On a global scale, the ocean's surface elevation oscillates as a result of the balance between gravitational attraction and centrifugal forces on the ocean water in the Earth, Moon, and Sun system. There are 399 "tidal constituents" (individual components that make up the overall tide) that are used to describe the tides on earth (Doodson 1922). Each constituent represents a periodic change or variation in the relative positions of the Earth, Moon and Sun and each has a unique period. The amplitude and phase associated which each constituent varies from place to place, however. Most constituents have very small amplitudes, and the observed tide is dominated by only a few of them.

The dominant tidal constituent in the Hudson River Estuary is the semidiurnal, principal lunar constituent (the M_2 tide), which has a period of 12.42 h. It therefore produces approximately two "high tides" and two "low tides" per day. The semidiurnal, principal solar component (the S_2 tide) also produces two "high tides" and two "low tides" per day and has a period of 12 h. This constituent is smaller than the lunar component. It works to amplify or reduce the amplitude of the lunar component, depending on whether it is in or out of phase with it. The result is an oscillation in the observed tidal amplitude with a period of 14.8 d. So, about every 15 d, the lunar and solar components are in phase and the observed tidal amplitude is highest. This is called the spring tide. And about every 15 d, the two components are out of phase and the observed tidal amplitude is lowest, which is called the neap tide. Therefore, this oscillation is commonly referred to as the spring/neap (spring here has no relation to the season) tidal cycle. In the Hudson River Estuary, the N_2 constituent (larger lunar elliptic) with a period of 12.66 h is also important. The M_2 - N_2 combination produces an oscillation that repeats every 27.6 d. This combination modulates the amplitude of the spring/neap cycle, causing it to be larger and smaller. Also observed in the Hudson River Estuary, although of lesser importance, are the K_1 (solilunar) and O_1 (diurnal lunar) constituents. The water surface elevation at the Battery during the spring of 1998 is shown in Figure 3e. The twice daily oscillation is due to the M_2 tide. The spring/neap component of the tidal system is also evident in the water surface oscillations. The amplitude of water level oscillation is lowest during the neap tide of 4 April 1998 and is highest during spring tide about 7.5 d (about midway in the spring/neap cycle) earlier, on 27 March 1998.

Besides the astronomical tidal forcing, meteorological events, such as storm surges due to strong persistent onshore wind, often raise the water surface elevation at the downstream boundary. The pronounced increase in water surface elevation on 21 March 1998 was caused by a coastal low pressure system. Strong winds from the East (Figure 3b) produced a surge of water that propagated into lower and upper New York Bay and raised the water surface elevation for several days at the Battery (Figure 3e), and further upstream into the estuary as will be discussed later.

The salinity at the Battery depends on the amount of freshwater input upstream. The salinity reflects the mixture of water from upstream freshwater inflows (mostly at Troy, 0 parts of salt per thousand parts of water [ppt]) and the offshore saltier waters of the New York Bight (about 33 ppt). The saline waters near the Battery are the primary source of salt to the estuary. During low flow conditions, surface and bottom salinities can be as high as 20 and 28 ppt, respectively. During periods of high freshwater inputs, the surface and bottom salinities are much lower, about 5 and 20 ppt, respectively.

Water Surface

The water surface itself is an important boundary of the estuary because it is where atmospheric heating (e.g., solar radiation) and cooling (e.g., latent heat of evaporation) occur. In addition, wind over the water surface can drag water and modify the currents. There are two factors that limit the effect of wind on the Hudson River Estuary, compared to other estuaries. First, the prevailing wind direction is from the west, whereas the estuary is oriented north-south. This implies that the wind does not blow along the main axis, which would lead to a maximum effect of the wind. Second, the estuary is relatively narrow, and cliffs (e.g., Palisades) and hills often shelter and reduce the wind at the water surface. However, winds can have a strong impact on limited straight sections of the river, if they blow in the right direction. Hunkins (1981) showed a weak effect of wind on surface currents at Yonkers. In an isolated event (16 February 1967) Busby and Darmer (1970) found that strong northerly (along the main axis) winds with gusts up to 55 mi/h prevented the occurrence of high tide at Poughkeepsie. The effect of wind is expected to be largest in the open, shallow areas of the estuary (Tappan Zee/Haverstraw Bay, Newburgh Bay).

Water Level

The drivers, discussed above, affect the hydrodynamics of the estuary. How and why

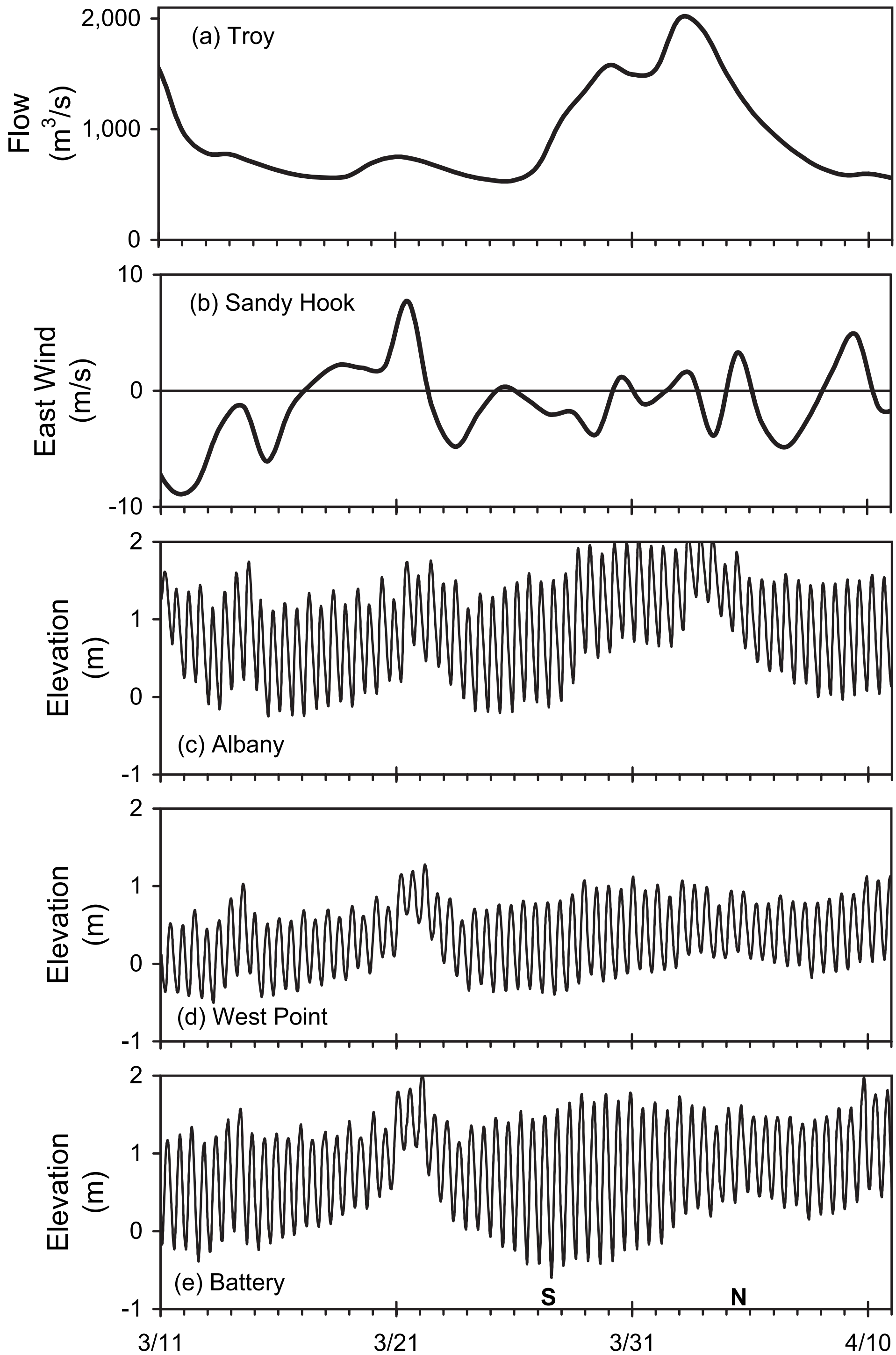

Figure 3. (a) Daily freshwater flow rate at Troy, (b) mean daily easterly wind component at Sandy Hook, and (c-e) water surface elevation at Albany, West Point, and the Battery in the spring of 1998. "S" and "N" labels in panel (e) mark spring and neap tides referred to in the text, respectively.

the hydrodynamics respond to the drivers is the subject of the remainder of this paper, starting in this section with the water levels. The most obvious feature of the water surface elevation of the Hudson River Estuary is its periodic oscillations, due to tides (solely due to the moon and sun). Water levels also respond to surges (due to winds and atmospheric pressure effects). Changes in water surface elevation at the downstream end propagate all the way up to Troy. The tidal peak moves up river at about 25–30 km/h with high tide at Albany occurring about 9–10 h after high tide at the Battery. Freshwater input, especially at the most upstream end, can also affect the water surface elevation (Darmer 1970).

Spatial Variability

While the M_2 tide is the major source of energy which drives the circulation in the region, the N_2, S_2, K_1, and O_1 tides are all significant and contribute to the diurnal, fortnightly, and monthly variations in the magnitude of the tides, currents and mixing. The amplitudes of these tidal constituents (M_2, N_2, S_2, K_1, and O_1) vary, at the Battery (67 cm, 16 cm, 13 cm, 10 cm, 5 cm), West Point (38 cm, 10 cm, 4 cm, 8 cm, 4 cm) and Albany (69 cm, 11 cm, 10 cm, 13 cm, 7 cm). The amplitude of the overall tide is thus highest at the Battery (Figure 3e), lower at West Point (Figure 3 d), and higher again at Albany (Figure 3c). This spatial pattern in amplitude is due to the interplay of three main factors, including friction, geometry, and wave reflection.

Friction.—The energy of the tidal wave is dissipated by friction, which works to reduce the amplitude with distance upstream into the estuary.

Geometry.—The cross sectional area decreases with distance upstream. To conserve energy, the amplitude of the wave increases. This tends to cause an increase in amplitude with distance upstream.

Wave reflection.—The tidal wave traveling upstream from the Battery is reflected at the dam at Troy and travels back downstream. The resulting water surface elevation is then the sum of the upstream propagating wave and the downstream propagating reflected wave. These waves have the same period, but since they travel in different directions, they can amplify or reduce each other's signals (this varies in space and time).

Temporal Variability

There are a number of variations in the observed water surface elevations that occur on time scales longer than the semidiurnal and diurnal tides. These can be attributed to the spring/neap tidal cycle and surges, already discussed previously. Fluctuations in water surface elevation due to storm surges can easily propagate along the entire estuary, just like the tide does.

Another factor influencing the water surface elevation is freshwater inflow. Especially at the upstream end, the water surface elevation is significantly affected by the freshwater input. The water surface elevation at Albany is elevated around 31 March 1998 (Figure 3c), due to the high freshwater input (Figure 3a). The fluctuations in water surface elevation introduced by changes in inflow propagate in the downstream direction (opposite to tide and storm surges) and can be seen at West Point (Figure 3 d). They are hardly noticeable at the Battery, however.

Currents

Tides and currents are intimately connected. Spatial gradients in water surface elevations cause currents (water flows from higher to lower elevation with some modification due to the earth's rotation) and spatially nonuniform currents cause changes in water surface elevation (water "piles up" where currents converge). Therefore, much of the oscillatory nature of the tides is also reflected in the currents.

Timing of Water Level and Current Fluctuations

In the vicinity of the Battery, the tide is produced by a progressive wave which propagates in from the Atlantic Ocean. The time of maximum flood currents occurs at the same time as high tide and the time of maximum ebb currents occurs at the same time as low water. The times of the two slack (time when the current is not moving) waters of a tidal cycle occur at the times that the water surface is at its tidal mean level. This situation changes markedly as you move upstream. For example, near the George Washington Bridge, maximum flood occurs about 30 min before high tide and maximum ebb occurs about 30 min before low tide. This time shift increases even more farther upstream. As Albany is approached, the tide wave takes on the characteristics of a standing wave. The time of slack water occurs much closer to high water and low water. Maximum ebb currents begin to occur nearly 3 h before low water and maximum flood currents begin to occur about 3 h before high water.

Temporal Pattern

The current variability within the tidal cycle near Indian Point is shown in Figure 4. The tidal cycle reversals are most obvious. Bottom currents are generally slower than the surface currents, due to the effect of friction acting of the water column at the bottom. The only exception occurs when strong winds oppose the direction of flow, causing the surface currents to slow down. Then the highest currents are located at middepth. As with the water surface elevations, the spring/neap cycle manifests itself by varying the amplitude of the oscillations (Figure 4). The spring tide on 27 March 1998 is accompanied by the largest currents, while the neap tide on 4 April 1998 has the lowest currents, within the time period shown.

Tidal versus Net Flow

The previous section illustrated that the observed flow direction oscillates from upstream to downstream in response to the tidal forcing. Since the flow is driven primarily by the tidal forcing, it is called "tidal flow." The "net flow" is defined as the long-term average flow at a given point. Practically, the net flow is difficult to measure because it involves a very large fluctuating component and a small mean value. The net flow rate has to be in the downstream direction and equal to the magnitude of freshwater flow entering the estuary upstream of that point (as well as precipitation minus evaporation and any other inputs), which is on the order of 400 m^3/s (mean flow at Troy). It should be noted that this only has to be true for the total net flow rate over the entire cross section. As discussed below, net bottom currents are typically in the upstream direction in estuaries.

The time series of currents at Indian Point (Figure 4) demonstrates that the tidal flows are significantly larger than the net flow. At that location (average depth = 12 m, width = 1,300 m), a current of 50 cm/s corresponds to a flow rate of 7,700 m^3/s. Busby and Darmer (1970) measured tidal flows of 4,800 m^3/s at Poughkeepsie. de Vries and Weiss (1999) measured tidal flows of 11,000 m^3/s in Haverstraw Bay. Thus, it is evident that tidal flows in the Hudson River Estu-

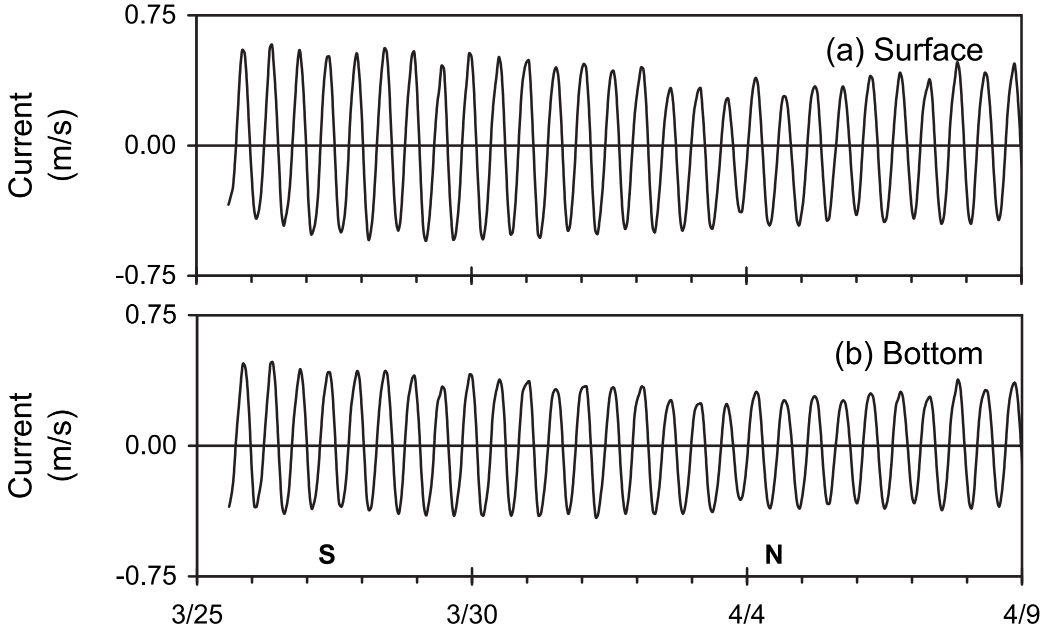

Figure 4. Time series of alongshore currents in the middle of the estuary near Indian Point during 1998 at (a) surface and (b) bottom. Positive is upstream. "S" and "N" labels in panel (b) mark spring/neap tides referred to in the text, respectively.

ary are typically about an order of magnitude greater than the net flows. Since the net flow is small compared to the tidal flow, low frequency variations in the tidal flow (e.g., storm surges) can lead to temporary reversal of net flow. Busby and Darmer (1970) found net upstream movement for several days at a time at Poughkeepsie.

Spatial pattern

Spatially, the currents vary across and along the estuary due to geometry effects. Typically currents are higher over the deeper center channel and lower on the shallower side banks. However, such an idealized picture is rarely observed in the Hudson River Estuary because the geometry can be very complex. Figure 5 shows lateral profiles of near surface currents by Indian Point at four different times in the tidal cycle. During times of strong ebb (Figure 5a, b), flow is in the downstream direction with typical currents of 50–100 cm/s. During the slack before flood period (Figure 5c), the flow is in the process of switching from downstream to upstream. At that time, the current is actually directed in opposite directions at several cross sections. The current is upstream on the west side of the river and downstream on the east side. Two hours later (Figure 5 d), the flood currents are upstream throughout the Indian Point area, with typical currents of less than 50 cm/s.

The fact that the tide generates currents that flow around bending regions of the Hudson River Estuary complicates the circulation even further. Georgas and Blumberg (2004) demonstrate that water level is slightly higher on a bend's outer bank during both flood and ebb than it is on the bend's inner bank. A transverse circulation cell is setup that is directed towards the outer bank at the surface and toward the inner bank at the bottom. In the presence of stratification, the transverse circulation tends to produce upwelling of salt at the inner part of

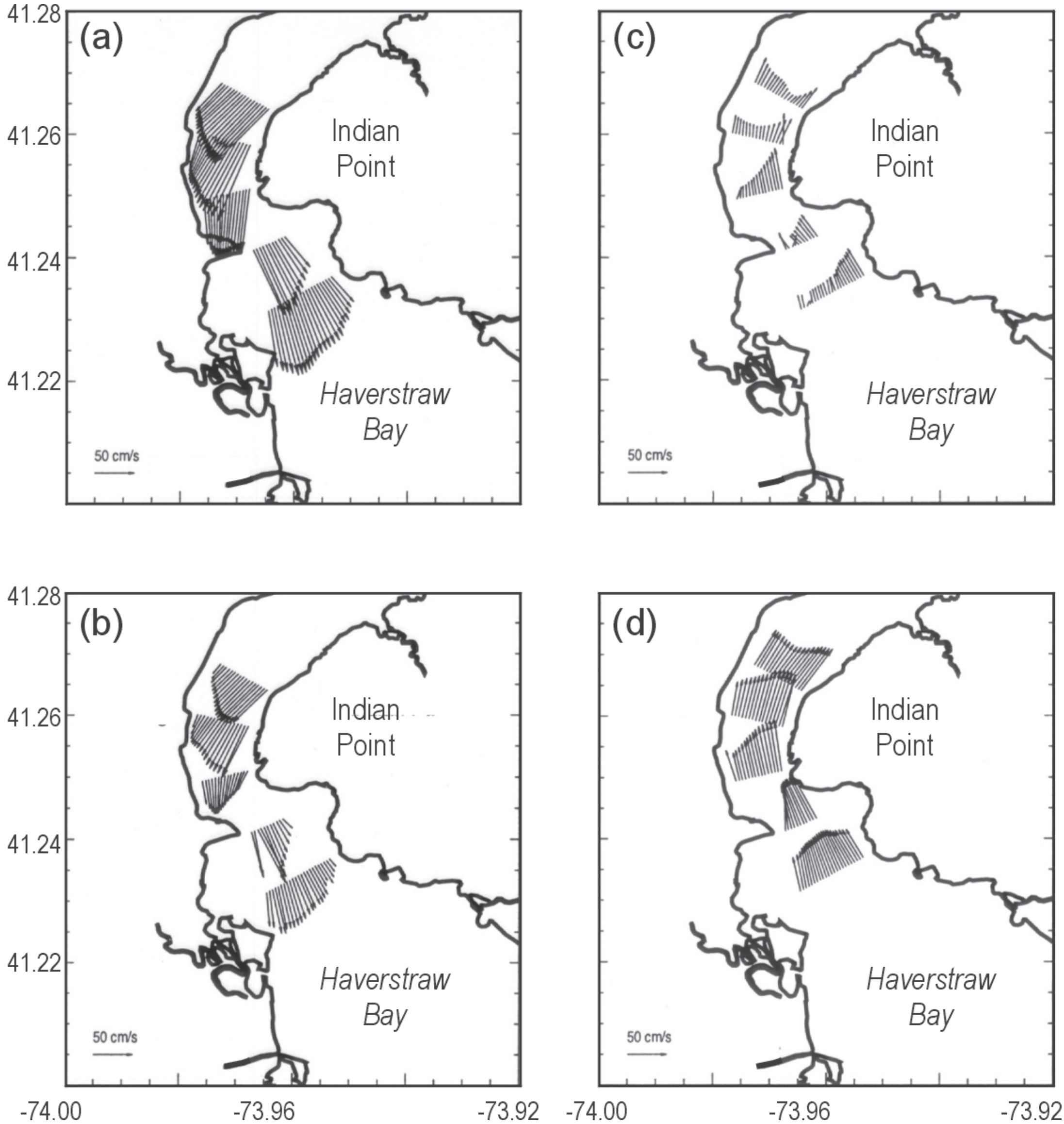

Figure 5. Observed near-surface currents in the region of the river near Indian Point at four points in the tidal cycle on 2 April 1998 (from HydroQual 1999).

the bend leading to stronger cross estuary density gradients (Chant and Wilson 1997).

Tidal Excursion

An important transport concept in estuaries is the tidal excursion, which characterizes the distance a water parcel travels as a result of tidal currents. The tidal excursion is the distance between the most upstream and downstream locations occupied by a water parcel during one tidal cycle. If a parcel is released at high tide, the ebb tide will carry it downstream a distance equal to the tidal excursion. Of course, the tidal excursion varies in time (freshwater flow, spring/

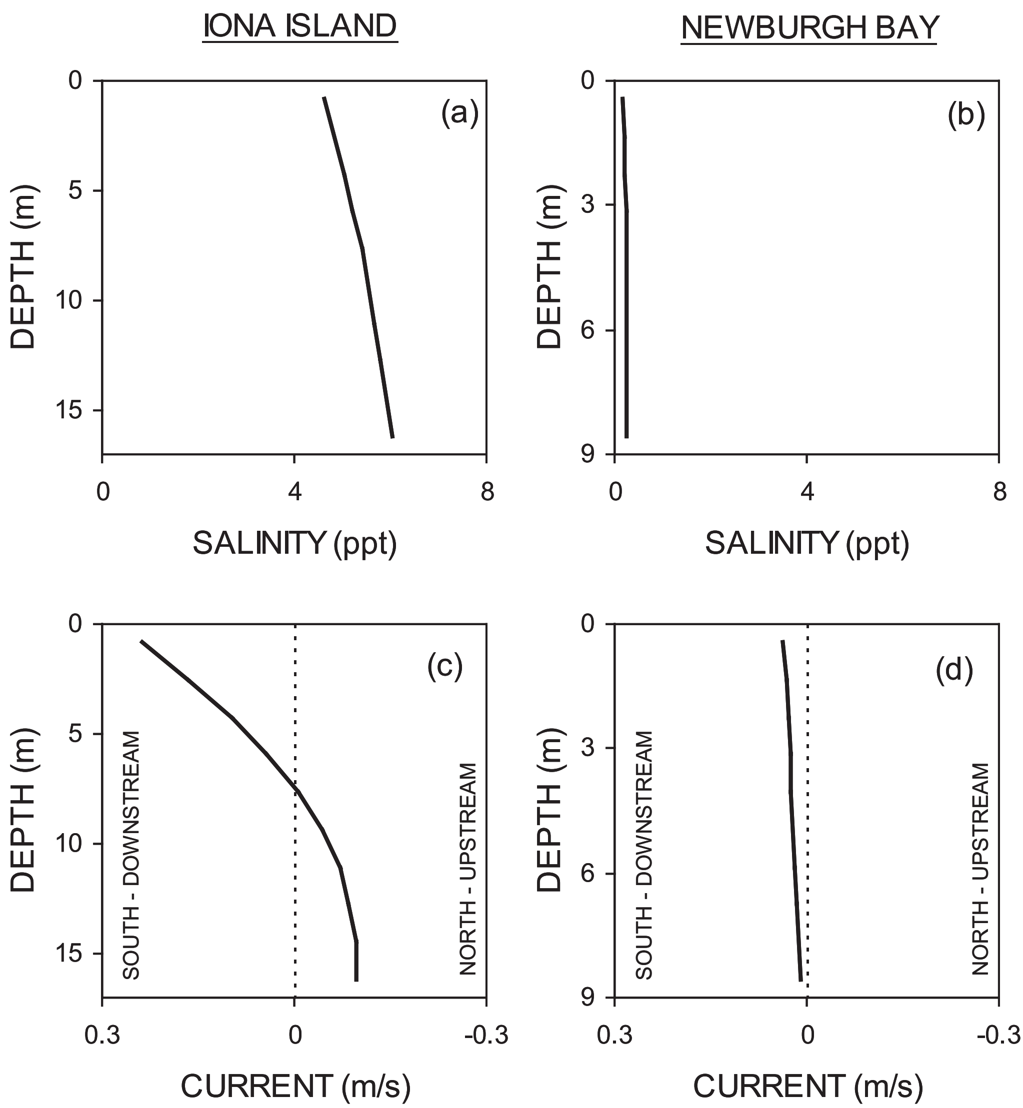

Figure 6. Vertical profiles of modeled (a, b) salinity and (c, d) net currents (a, c) at Iona Island and (b, d) in northern Newburgh Bay for 25 July 2000–8 August 2000 (from Hellweger et al. 2004).

neap) and space (location and depth). The typical tidal excursion in the Hudson River Estuary is 5–10 km, but it can be as high as 20 km.

Estuarine Circulation

Salt adds mass to water, and as a result, salt water has a higher density than freshwater. The density difference between fresh and salt water is small (3.5% or less) but sufficient to significantly affect the circulation in estuaries. Denser salt water enters the estuary in the deeper parts of the water column and lighter freshwater floats as a lens on top of the salt water. The result is net upstream currents in the bottom layer and net downstream currents in the surface layer, in locations where salt is present. This is commonly referred to as the "estuarine circulation." An example of this is presented in Figure 6, which shows vertical profiles of salinity and net currents at two locations. In northern Newburgh Bay, the salinity is very low, and as a result the net currents are in the downstream direction at all depths. At Iona Island, there is significant salinity, which causes an estuarine circulation. The net currents are in the upstream direction in the bottom layer and in the downstream direction in the surface layer. Steward (1958) measured weak upstream bottom currents at Riverdale. Further upstream, at West Point, upstream directed bottom currents were only present on 1 out of 4 d. At Yonkers, Hunkins (1981) measured net downstream surface and upstream directed bottom currents of 13 and 2.3 cm/s, respectively.

There are large temporal variations in the estuarine circulation patterns described above due to the spring-neap changes in the tidal forcing. The intense currents that occur during spring tides and the much smaller currents of the neap tide lead to a dramatic variation of vertical mixing intensity and to changes in vertical salinity stratification. The greater the stratification, the stronger the estuarine circulation becomes. Bowen and Geyer (2003) show an order of magnitude change in the horizontal salt transport as a result of spring-neap cycle changes in the estuarine circulation. The effect of the spring-neap cycle on salinity will be discussed in more detail later.

Temperature

The water temperature in the Hudson River Estuary varies seasonally primarily in response to changes in atmospheric heating and cooling. Tributary temperature and ocean temperature play a lesser although not insignificant role. Spatially, the temperature is influenced in part by power plant cooling water discharges.

Temporal Variability

Typical maximum (August) temperatures are 25°C and minimum (January and February) temperatures are 1.0°C (Poughkeepsie Water Works; period of record: 1951–1987; Wells and Young 1992). This seasonality is also shown in Figure 2b. There is little surface to bottom temperature difference because the water column is relatively well mixed in the vertical. Ice shells are observed floating on the water surface in the winter. Further downstream in the area of saltwater intrusion, the temperature of the waters of the Hudson River Estuary is similar to that of the Atlantic Ocean. Sometimes on a hot summer day there can be pockets of warm water confined to the near surface layers. They are typically short-lived and disappear as a result of mixing by winds or tidal currents. For a more detailed discussion on the seasonality of temperature in the Hudson River Estuary, see Wells and Young (1992) and Mancroni et al. (1992).

Spatial Pattern

Spatial differences in temperature occur because some areas (e.g., shallow banks) tend to respond faster to changes in the forcing functions (i.e., water surface heat flux). Also, the shallower tributaries tend to warm and cool faster than the deeper estuary and therefore can be a source of warm water in the spring and cool water in the fall. Maximum summer temperatures can easily exceed 30°C in these shallow regions. In the summer, the temperature is coolest in the downstream portions of the estuary. It reaches a maximum near the region where several power plants are located and then decreases slightly until the upper reaches of the estuary where the temperature again increases. There are five major power plants located along the

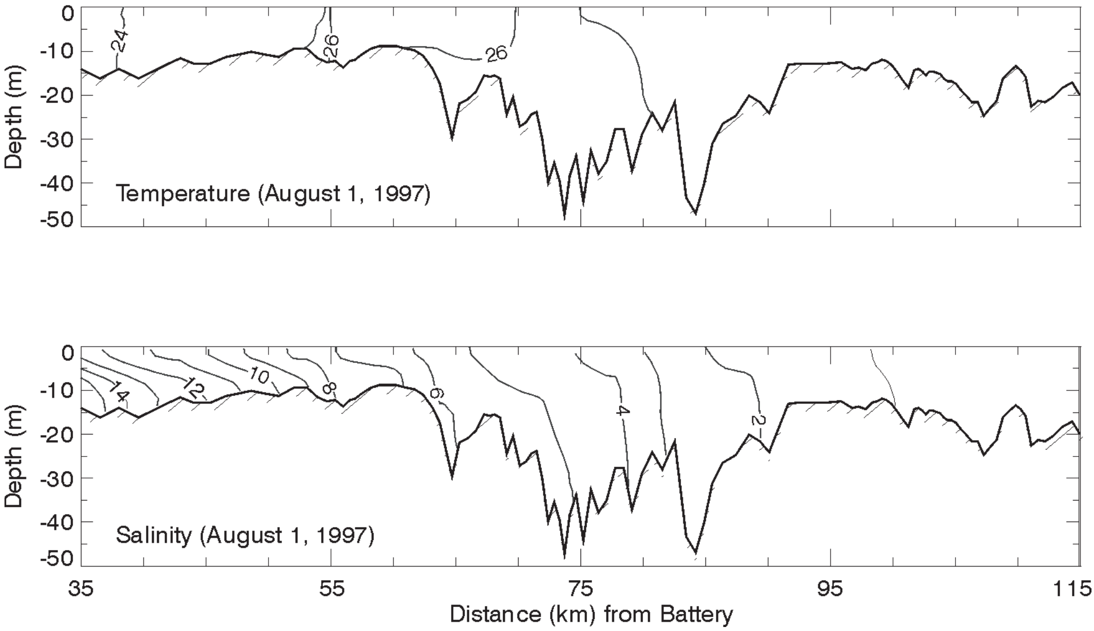

Figure 7. Longitudinal profile of observed temperature (a, top) and salinity (b, bottom) on 1 August 1997. Distance is measured from the Battery (from HydroQual, 1999). Indian Point is located at the 55 km mark.

river and they are an important factor influencing the temperature in the Hudson River Estuary (e.g., Wrobel 1974; HydroQual 1999). These power plants withdraw large volumes of water for cooling and discharge back to the river at an elevated temperature. The maximum cooling water flow from the Indian Point nuclear power plant is 110 m^3/s (Hutchison 1988), which is comparable in magnitude to the summer low flow of the largest natural freshwater input into the estuary (Green Island, August, 160 m^3/s). Withdrawal and discharge locations for power plants are typically located close enough so that their effect on currents is localized and constrained to the immediate area of the power plant. However, the increase in temperature can be evident over a larger area. The maximum temperature increase for the Indian Point nuclear power plant is between 8°C and 9°C in the vicinity of the discharge (Hutchison 1988) and about 1°C or so midriver (HydroQual 1999). The effect of the plant discharge can be seen in the spatial temperature profile shown in Figure 7a.

Salinity

Freshwater enters the estuary at the upstream end, and salt water mixes upstream from the ocean. The result is a mixture of fresh and salt water throughout much of the estuary. The horizontal and vertical distribution of salt varies dynamically at various time scales in response to changes at the upstream and downstream boundaries.

Spatial and Temporal Pattern

The salinity in the Hudson River Estuary varies along the length of the estuary, as illustrated in the longitudinal profile presented in Figure 7b. The Figure shows that salinity is higher at the downstream end and lower at the upstream end. Also, the salinity tends to be higher in the bottom layer than at the surface. The "S" shaped salinity contours are quite typical of estuarine environments (vertical stratification is discussed in the next section).

The salinity distribution varies temporally. At the tidal time scale, the salinity can change rapidly at a certain location due to the tidal water movement, especially in the area of strong longitudinal salinity gradient (salt front, see below). At longer time scales, salinity intrudes further into the estuary during neap tides and retreats during spring tides (Bowen and Geyer 2003). At even longer time scales, the salinity responds to the freshwater input, which varies at seasonal and shorter time scales. Short time scale pulses of freshwater are known as freshets. Freshwater tends to push the salt water out of the estuary during high flows and permits salt water to intrude during low flow periods. The freshwater flow variability is evident in both the surface and bottom salinity at Croton-on-Hudson and at the Lincoln Tunnel as shown in Figure 2c, d. The salinity is generally higher during the low-flow summer and lower during the high-flow spring. Also, the effect of each of the freshets (Figure 2a) is evident in the salinity (Figure 2c, d). During the spring high-flow period, the salt front (the upper limit of saltwater intrusion, defined here as a salinity of 0.1 ppt) is located between Yonkers and Tappan Zee and during the summer low-flow period it moves north and is typically located south of Poughkeepsie, just south of the City of Poughkeepsie drinking water intake (Wells and Young 1992). For a more detailed discussion of the estuarine circulation and the movement of the salt front in the Hudson River Estuary, the reader is referred to Geyer et al. (2000) and de Vries and Weiss (2000).

Vertical Stratification

The saltiest water resides at the bottom with fresher water at the surface, and hence there is a vertical salinity gradient. This vertical salinity stratification exists throughout the estuary (in the presence of salt) and can be seen in the longitudinal salinity profile shown in Figure 7b. The vertical distribution of salinity is characterized by an upper layer of low salinity, which very slowly increases with depth, an intermediate layer of more rapid salinity increase, called the halocline, and a deep layer in which the salinity increase with depth is small. This "S" shape salinity distribution is the result of the interplay of factors that "want" to keep the water column stratified and those that "want" to mix it. The vertical salinity gradient serves to stabilize the water column and inhibit vertical mixing. A lighter water parcel from the fresh surface layer will resist mixing into the heavier salty bottom layer. This "stability" tends to keep the water column stratified. The water column can become destratified through two mechanisms, tidal straining and turbulence mixing. Both tend to mix the water column and erase the vertical stratification.

Tidal straining (Simpson et al. 1990; Nepf and Geyer 1996) arises from the vertical variation in tidal currents in the presence of a longitudinal salinity gradient. The maximum stratification typically occurs at low water after the ebb flow has moved fresher water in the upper layers seaward over the saltier water in the deeper layers. During the flood, this process is reversed with tidal straining acting to reduce the stability of the water column, which results in a mixed water column close to high water.

Turbulence is produced at the water surface due to wind stirring and at the bottom, where tidal currents move back and forth over the sediment bed. The turbulent energy is highest during the spring tide, and it is then when the vertical stratification often breaks down. This is evident in the

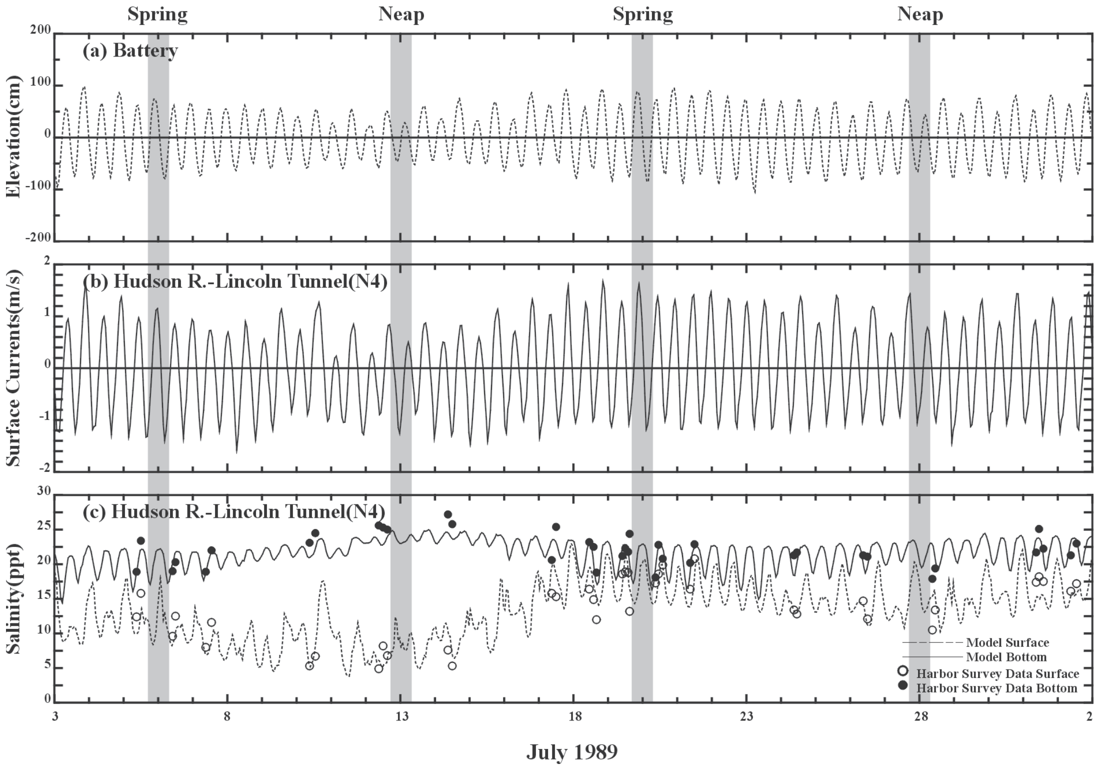

Figure 8. A 30 day time series beginning July 3, 1989 of (a) water surface elevation at the Battery, (b) currents at the nearby Lincoln Tunnel (N4) station and (c) surface and bottom salinity at the nearby Lincoln Tunnel station (from Blumberg et al. 1999).

salinity at the Lincoln Tunnel (Figure 8c). Throughout most of the period shown, the bottom salinity is significantly higher than at the surface. However, during spring tide (e.g., Day 292, see tidal amplitudes of Figure 8a and currents of Figure 8b), the energy is sufficient to overcome the stabilization and the water column mixes vertically causing the surface and bottom salinities to become the same. In general, however, the vertical stratification, which is observed in the water column, is the result of the interplay between both the straining and mixing, both of which vary in time.

Dispersive Processes

Various processes, operating at different spatial scales, contribute to the horizontal spreading of salinity and other constituents that have been introduced into the water column. The main processes are shear dispersion and tidal trapping, both of which will be described below.

Shear Dispersion

The currents in the Hudson River Estuary have significant lateral (Figure 5) and vertical (Figure 6) structure. This structure in currents spreads out pollutants that are in the water column by a process termed shear dispersion (Pritchard 1954). Pritchard considered the vertical current structure of the estuarine circulation, in what turns out to be the dominant, but not only mechanism for shear dispersion in the Hudson River Estuary. The following examples illustrate the mechanisms. An initial vertically mixed instantaneous release (“slug release”) of a substance (e.g., tracer) near Iona Island will travel upstream and downstream with the tide, but the difference in net currents between the surface and bottom layers is 20 cm/s (see Figure 6), which means that after 1 d, the slug at the bottom is almost 20 km further upstream than the slug at the surface. The estuarine circulation will have created this slug dispersion. The effect of this net estuarine circulation is evident in the longitudinal tracer concentration profile presented in Figure 10b. The tracer plume shows much higher spreading downstream of the 90 km location, which is the area where the salinity is high enough to cause an estuarine circulation.

Consider a second example where lateral variations in currents exist and a hypothetical instantaneous spill (slug) release by Indian Point at a time when the current is switching from ebb to flood (Figure 5c). At this time, water flows upstream and downstream on the west and east sides of the river, respectively. These currents would spread out the slug by stretching it in the upstream and downstream directions. This variation in currents is extreme and specific to this time in the tidal cycle only. Most of the time water either flows upstream or downstream throughout the cross section. However, the upstream component is stronger on the west side and the downstream component is stronger on the east side, and this structure causes dispersion as well. For example, the strong flood currents shown on Figure 5 d have significant lateral (across river) structure and would also lead to significant longitudinal spreading of a slug in the upstream direction.

Tidal Trapping

The geometry of the Hudson River Estuary can be very complex, containing many irregularities (coves and inlets). The interplay of the tidal flow with these geometric

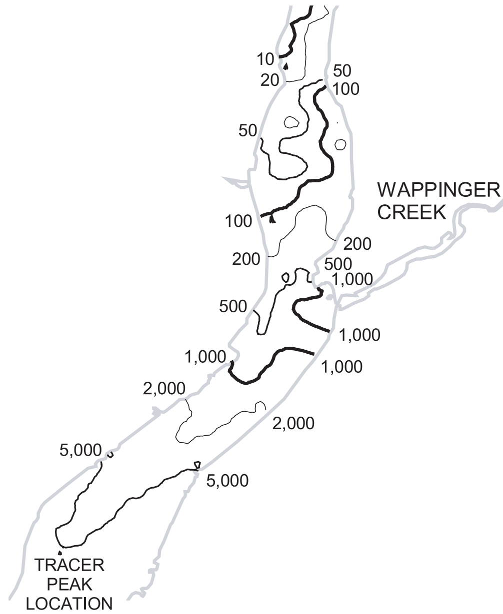

Figure 9. Horizontal distribution of modeled SF_6 tracer concentration in fmol/L in northern Newburgh Bay by Wappinger Creek 2 d after a slug release (from Hellweger et al. 2004).

irregularities enhances longitudinal dispersion by a process called "tidal trapping" (Okubo 1973). The basic concept of tidal trapping is that geometric irregularities can temporarily trap a water parcel as it passes by and then release it at some later time. This effectively removes a small amount of water from the original main channel water mass and then adds it back later to a new main channel water mass. For example, a water parcel with low salinity is removed from its original low salinity main channel water mass and then added later to a new main channel water mass with higher salinity.

The process of tidal trapping is illustrated in Figure 9, which presents tracer concentrations in northern Newburgh Bay near Wappinger Creek. In that experiment, a slug of tracer was released further south in Newburgh Bay. The Figure shows the hori-

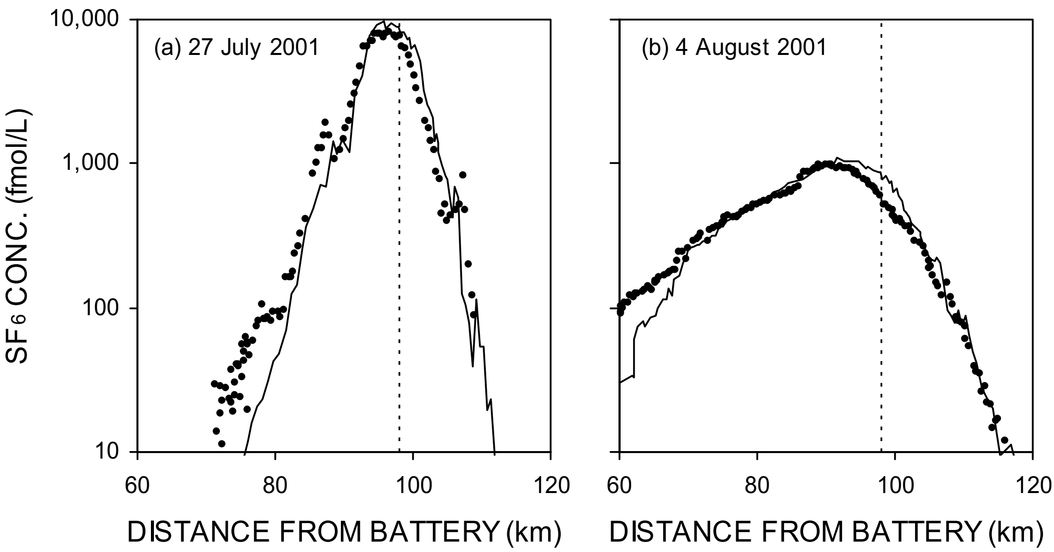

Figure 10. Longitudinal profiles of tracer concentration 2 and 10 d after a slug release in Newburgh Bay. Symbols are data and lines are model (from Hellweger et al. 2004).

zontal tracer distribution 2 d after the release. The concentration at the mouth of Wappinger Creek is higher compared to that in the main channel (1,000 versus 500). A water mass with a high tracer concentration became trapped there. The tidal trapping by Wappinger Creek is also evident in longitudinal profiles of tracer concentration, presented in Figure 10a (Wappinger Creek is located at the 105 km point in that Figure). At the upstream and downstream end of the plume, there are smaller scale "secondary peaks," which are also the result of tidal trapping (located at the 90, 105, and 110 km positions).

Longitudinal Dispersion Coefficient

The combined effect of all the dispersive processes is to cause a spreading of constituents, which can be characterized using a longitudinal dispersion coefficient. The longitudinal dispersion coefficient has been estimated using tracer studies, with values ranging from 3 to 270 m^2/s at various locations from Troy to Newburgh (Hohman and Parke 1969; Clark et al. 1996, 1997; Ho et al. 2002). Longitudinal dispersion coefficients are useful for characterizing the dispersion in an estuary, but it is important to realize that the dispersion changes in space and time and that there is not one coefficient that applies to all locations of the estuary or even one location at all times. This variability is evident in the estimated dispersion coefficients for the Hudson River Estuary given above.

Residence Time

Another important concept related to currents and dispersion is residence time or flushing time. This concept is defined as the average time a water parcel spends in the estuary or a certain part of the estuary (e.g., freshwater reach). For the freshwater region of the Hudson, the flushing time is simply the volume divided by the upstream inflow. During the springtime high inflow periods, the flushing time in the estuary is less than 40 d. However, it is on the order of 200 d during the summer low flow periods. In the estuarine portions of the Hudson River, the flushing time is defined as the average volume of estuary divided by the seaward rate of outflow. For a salinity intrusion of 80 km and a typical estuarine surface current of 10 cm/s (see Figure 6), the residence time is about 8 d. This is obviously much shorter than the residence times in the freshwater regions, again demonstrating the impressive dispersive characteristics of the estuarine circulation.

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