3 The Physical Oceanography Processes in the Hudson River Estuary

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ABSTRACT The Hudson River has the attributes of a typical, partially mixed estuary – a moderate salinity gradient, significant vertical stratification, and a vigorous, two-layer circulation regime. Yet it also displays considerable variability, both in space and in time. In its northern reaches, the estuary becomes a tidal river, with no trace of oceanic salt but vigorous tidal currents. The salinity intrusion extends 100 kilometers (km) into the estuary during low discharge conditions, but it retreats to within 25 km of New York Harbor during the high river flows of the spring freshet. Fortnightly variations of tidal amplitude also result in pronounced variations in the estuarine regime, becoming well-mixed during strong spring tides and highly stratified during the weakest neaps. At the mouth of the Hudson is a complex network of tidal channels that link the estuarine regime of the Hudson to Long Island Sound, Newark Bay, and the Atlantic Ocean. The influence of the Hudson extends into the Mid-Atlantic Bight in the form of a low-salinity plume, which forms a coastal current and flows south along the New Jersey shore during favorable wind-forcing conditions.

Introduction

The Hudson River is one of the major watercourses of the United States East Coast. It originates on the slopes of Mt. Marcy in the Adirondack Mountains, extending nearly 600 km to New York City. Over this distance the Hudson's character changes dramatically, starting as a mountain stream, descending to become a lowland river, and then turning into a peculiar tidal river, slowly increasing in salinity to become an estuary, and finally joining the ocean as a complex network of tidal channels and bays bisecting the New York metropolitan area. These different environments are shaped by the interplay of a variety of physical processes with one element in common: the river flow. Runoff from the hillslopes coalesces to form the lakes and streams in the Adirondack highlands. The action of gravity on the accumulating water provides the driving force for this flow through the upper Hudson valley. South of Albany, the motion of the river becomes complicated by the influence of tides, which can be witnessed a remarkable 250 km from the sea. Although the tidal river flows both north and south, the net southerly river flow persists and provides the freshwater input that creates the Hudson estuary. This freshwater source is a dominant contributor to the physical regime of the estuary and harbor, as it controls the salinity structure, the vertical stratification, and the exchange of properties between the estuary, the ocean, and the atmosphere.

The Hudson River Watershed

The Hudson River watershed has two main branches, the Upper Hudson River and the Mohawk River (Fig. 3.1). The Upper Hudson extends 160 miles from Lake Tear of the Clouds in the Adirondacks to the Federal Dam at Troy. The upper Hudson is a steep-gradient river with numerous rapids, flowing through the rough terrain of the Adirondacks. Just north of the Federal Dam at Troy, the Mohawk River joins the Hudson from the west. The Mohawk follows a gentler gradient than the upper Hudson, draining the farm country between the Catskills and the Adirondacks. Although it flows through very different terrain, it contributes nearly the same discharge as the upper Hudson and comparable sediment loads.

The upper Hudson is unusual among rivers in the heavily industrialized eastern United States in that it is nearly unimpeded by dams. Although there are several dams along its course, their reservoirs are small, representing relatively little storage compared to the magnitude of the flow. Thus the seasonal flow characteristics of the river are close to their natural state.

A freshet occurs during the spring, when snowmelt from the Adirondack and Catskill Mountains

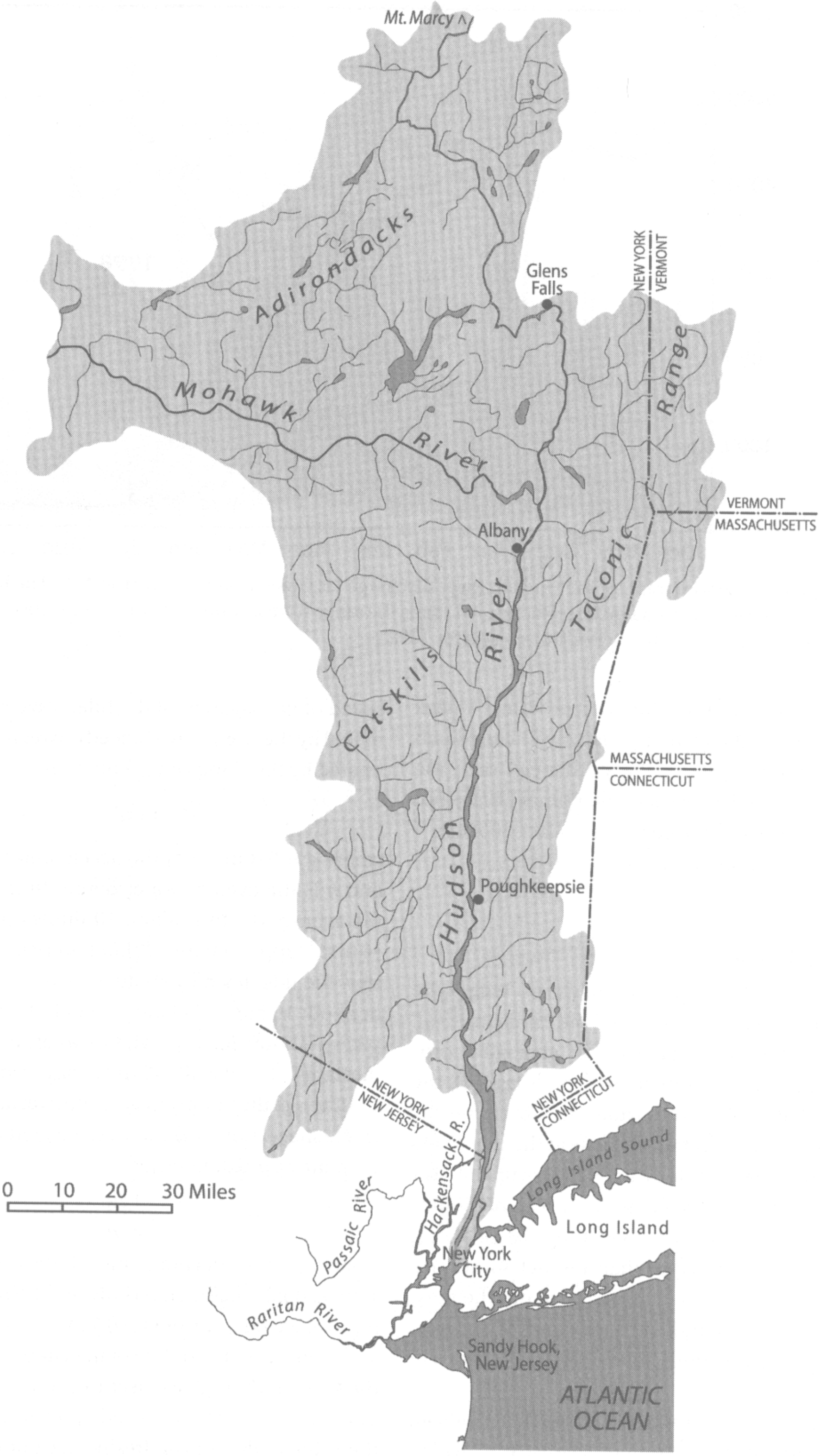

Figure 3.1. The Hudson River watershed. The length of the river, from Lake Tear of the Clouds to the Battery, is 725 km, and the area of its drainage basin is 34,700 km^2 (Howells, 1972).

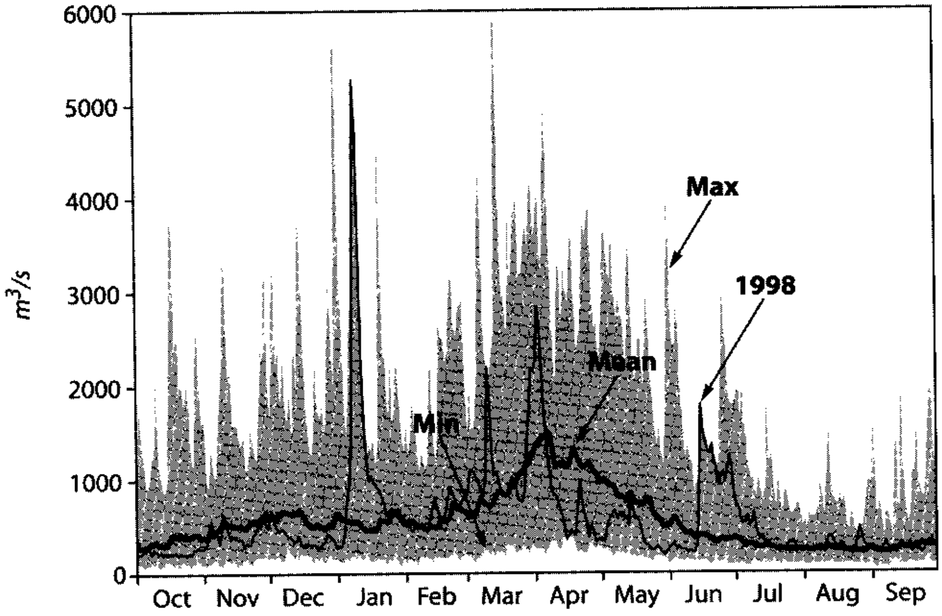

Figure 3.2. Hydrograph of the Hudson River discharge (at Green Island Dam, Troy, NY). The bold line is the mean discharge; the shaded area spans the range from minimum to maximum observed daily values; the thin line is the discharge for 1998.

combines with spring rains. This produces a peak flow of around 2,000 $m^3\ s^{-1}$, usually in late March or early April (Fig. 3.2). Big storms can raise the discharge to similar levels at other times of year, but typically the discharge decreases to 100–200 $m^3\ s^{-1}$ during the summer months.

The Tidal River

Below the dam at Troy, river flow is no longer the dominant agent of motion of the Hudson River. Owing to the particular suite of geological processes that have sculpted the landscape of the Hudson Valley, the river's shores in Albany are virtually the same elevation as those at the mouth. As a consequence, the tide extends 250 km up the river to the dam at Troy. River flow produces a net southward motion in the tidal river, but tidal velocities are usually much higher than the net southward motion due to river flow. Thus, in all but the most extreme outflow conditions, this part of the river flows in both directions, following the influence of the tides.

Through most of the length of the tidal Hudson, the tide propagates as a progressive wave, with high tide occurring later as it proceeds up the river. The speed of propagation of the tidal wave is approximated by the "long wave" speed c, which is approximated by the "long wave" equation

$$c = (gh)^{1/2}$$

where $g = 9.8\ \text{m s}^{-2}$ is the acceleration of gravity and h is the average water depth. In the Hudson the average depth is about 10 meters (m), so $c \sim 10\,\text{m s}^{-1}$ (approximately 20 knots). Frictional effects slow the tide down by about 20 percent to 8 $m\ s^{-1}$ or about 16 knots (DiLorenzo et al., 1999). Thus it takes six hours for the high tide to propagate from the Battery to Catskill, 185 km to the north. (Fig. 3.3).

Although the tide propagates upriver at an appreciable speed, tidal currents are considerably slower. They are approximated by

$$u_L = \frac{1}{2}\frac{h_T}{h}c$$

where h_T is the tidal range and h is the water depth. Typical tidal range on the Hudson is 1.5 m, producing average tidal currents of 0.7 $m\ s^{-1}$. The currents are stronger in the middle of the channel and near the surface, averaging closer to 1 $m\ s^{-1}$ or 2 knots. The tidal currents are considerably stronger than the velocity due to the freshwater outflow in the tidal river, which is on the order of 0.01 $m\ s^{-1}$ during

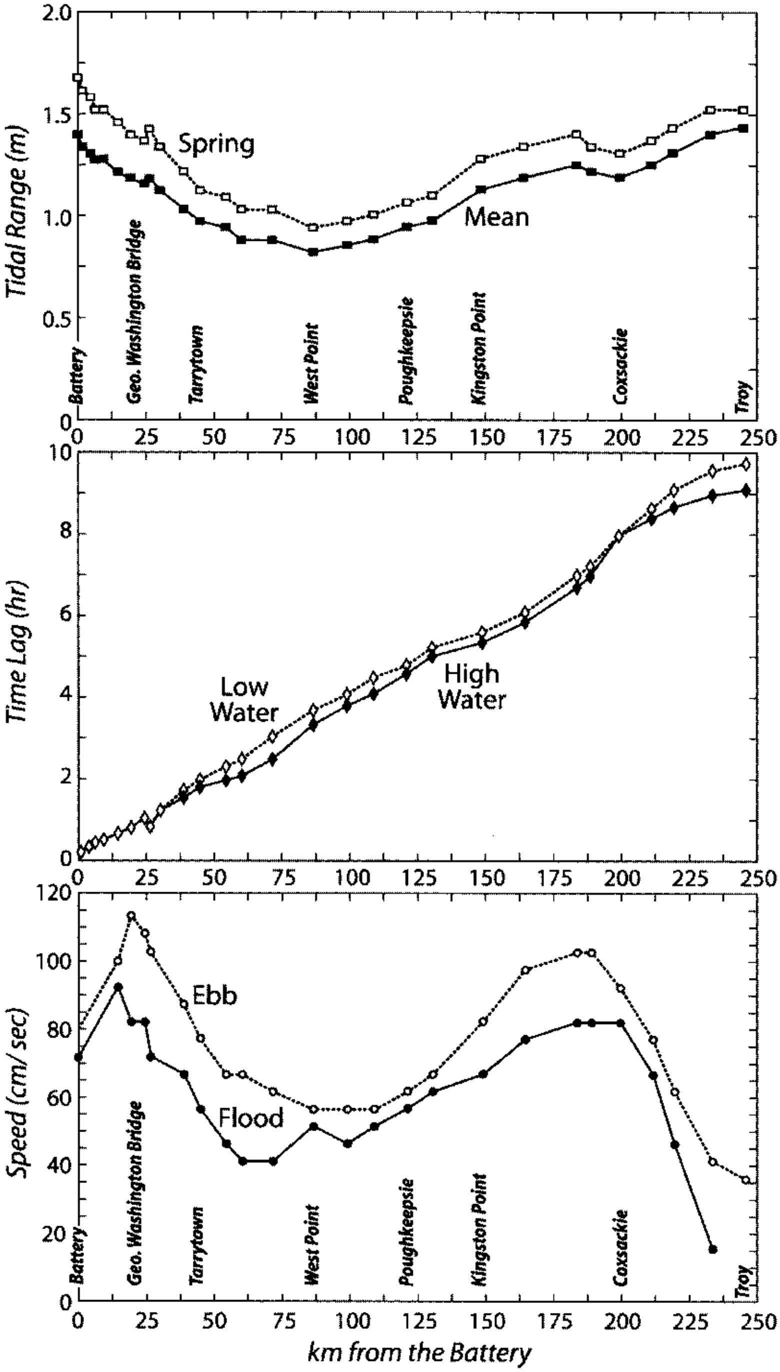

Figure 3.3. Tidal propagation conditions in the tidal portion of the Hudson River (adapted from DiLorenzo et al., 1999). The tidal wave is essentially a progressive wave, as indicated by the time delay as it propagates up the river. Tidal currents are quite energetic through most of the length of the tidal river.

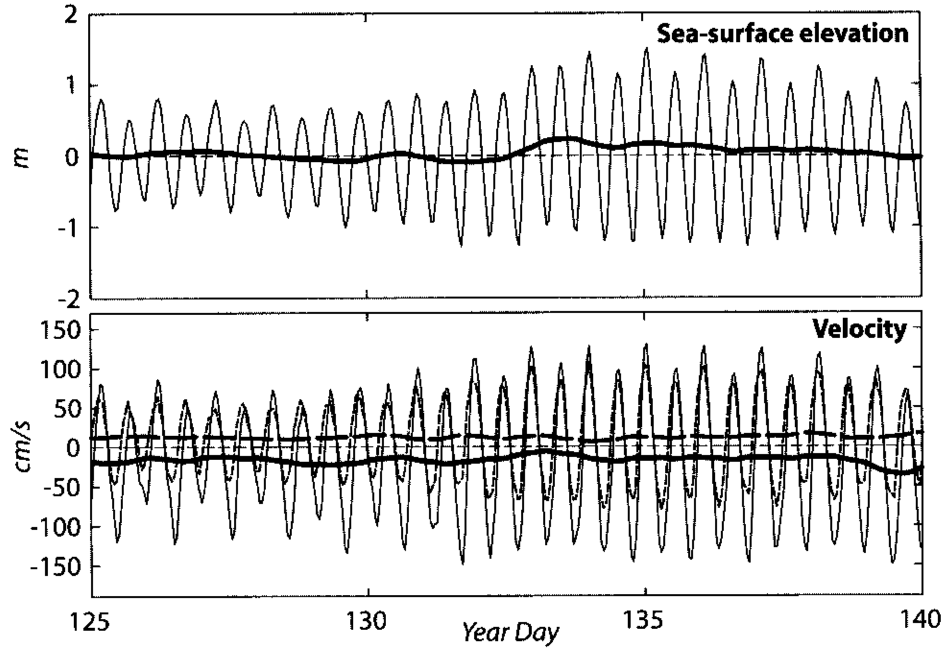

Figure 3.4. Time series of tidal elevation and currents in the lower estuary (near the Battery) in the spring of 1999. The upper panel indicates the tidal (thin line) and low-frequency (thick line) variations of sea level. The lower panel shows near-surface (solid) and near-bottom (dashed) currents, again indicating tidal (thin lines) and low-frequency (thick lines).

the dry summer months and reaches 0.2–0.5 m s^{-1} during the spring freshet. Thus, the tides provide most of the energy and fluid transport within the river below the dam at Troy.

The progressive wave character of the tide in the Hudson has an interesting influence on the phase of the currents relative to tidal height. In most tidal environments, slack water occurs close to high and low tide. However, in tidal rivers like the Hudson, maximum flood occurs within an hour of high tide, and the flood continues for the first two hours of the falling tide. As the tidal wave in the Hudson approaches the dam at Troy, it becomes more like a standing wave due to the reflection of the tidal wave at the head of tide.

The tidal forcing varies due to changes in the phase of the moon as well as other variations in the relative positions of the earth, moon, and sun. The most prominent of these occur at fortnightly and monthly time scales. These variations cause the tidal range at the Battery to vary from 1.2 m during small neap tides to almost 3 m during large spring tides (Fig. 3.4). The variations in currents are more complicated, due to changes in vertical structure of the flow as well as the influence of the salinity structure. Near-surface currents vary from around 0.7 m s^{-1} during neap tides to 1.3 m s^{-1} during the strongest spring tides. Near-bottom currents are considerably weaker, due to the frictional effects of the bottom boundary layer. These spring-neap variations in tidal flow have a profound influence on the estuarine regime, and likewise the estuarine circulation affects the strength of the currents, as explained in the next section.

The Estuary

The Hudson River estuary is an unusual hybrid of estuarine types, with elements of fjord, salt-wedge, and coastal plain estuaries. Glacial scouring of the Hudson Valley during the Pleistocene Epoch yielded a long, deep trough, which became a series of lakes dammed by glacial moraines during the retreat of the glaciers. When those moraines collapsed and sea level rose, the Hudson valley became a fjord, with depths of possibly as much as 200 m in the vicinity of the Hudson Highlands, where the bedrock was deeply gouged by ice (Worzel and Drake, 1959). Seawater filled the deep basin, and the freshwater outflow was confined to the surface layer. Tidal currents within the estuary were weaker than at present because of its great depth. This fjord environment was a nearly perfect trap for sediment, because neither the river outflow nor the tidal currents provided adequate energy to move sediment after it settled to the bottom from the turbid, surface layer. Sedimentation over the last 10,000 years has filled this glacial trough, and now the Hudson has depths of 10–15 m, more typical of coastal plain estuaries than fjords.

As the Hudson estuary got shallower, its hydrodynamics were significantly altered. Tidal currents became stronger, finally reaching their present values of around 1 m s^{-1}. As the tidal currents increased, the mixing between the freshwater outflow and the seawater increased. No longer could sea water penetrate far into the Hudson valley, due to the combination of tide-induced mixing and freshwater outflow. The present regime in the lower Hudson is a partially mixed estuary, with vigorous, tide-induced mixing between fresh and salt waters. The seawater is progressively diluted by river water as it extends up the estuary, and even during low flow conditions the water is nearly fresh at Peekskill, 70 km to the north of the harbor.

THE INFLUENCE OF FRESH WATER

The position of the salt front varies mainly due to variations in freshwater outflow (Abood, 1974). During high discharge periods in the spring (discharge exceeding 1,500 m^3 s^{-1}), the salt front is pushed south past Tappan Zee, roughly 30 km north of the Battery (Fig. 3.5). At the summertime minimum flow of around 100 m^3 s^{-1}, the salinity intrusion extends more than 90 km north to the vicinity of Newburgh. Because Poughkeepsie, at km 120, draws its drinking water from the Hudson, the upstream intrusion of salt provides a potential threat to its water supply. Twice in the last fifty years, during severe droughts in 1964 and 1995, the intrusion of salt water has come close enough to Poughkeepsie to influence its sodium content. The human health ramifications of the salinity intrusion have motivated numerous studies, including an ongoing monitoring program by the U.S. Geological Survey that documents salinity at points between Hastings and Poughkeepsie.

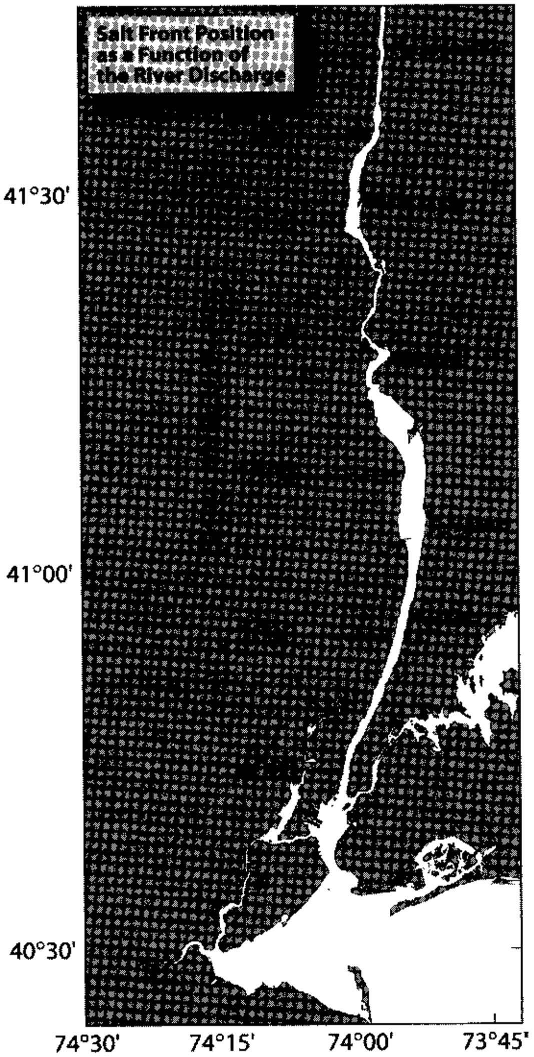

Figure 3.5. Map of the Hudson estuary showing the location of the salt front (approximately 1 psu) during different discharge conditions (adapted from Abood 1974). Discharge of 100 m^3/s typically occurs during dry summer months, whereas the typical freshet discharge is around 2,000 m^3/s. The highest observed discharge is slightly more than 4,000 m^3/s.

Although the river flow is modest relative to the tides, it has a dramatic influence on the estuary by providing a density contrast with the oceanic water. Ocean water contains about 3 percent salt by weight (or 30 parts per thousand, referred to by

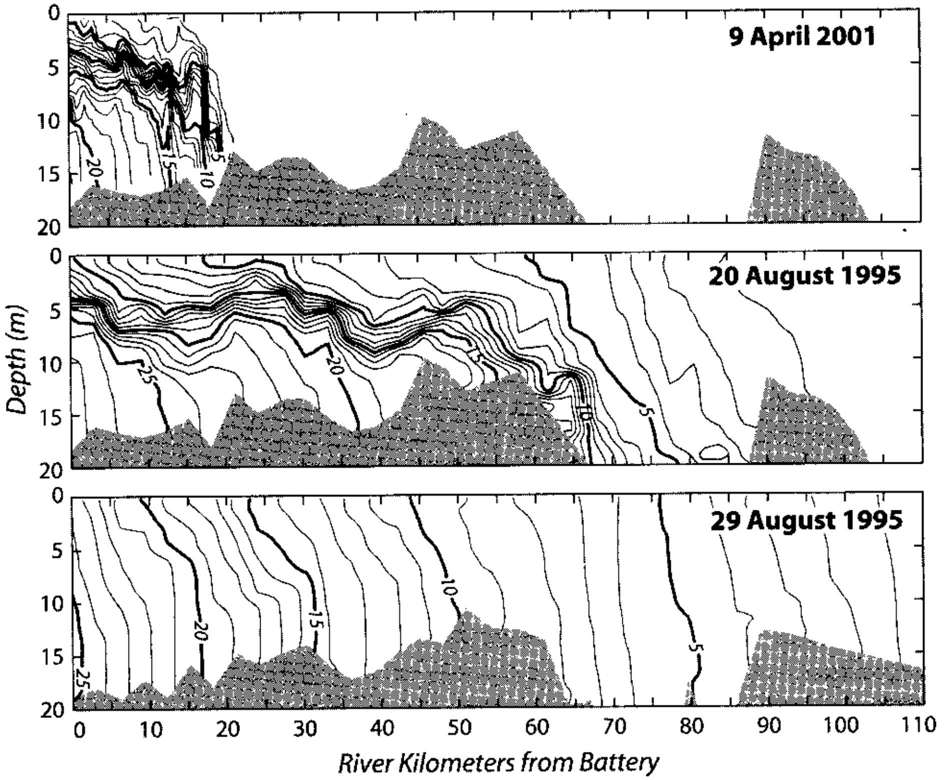

Figure 3.6. Salinity cross-sections in the Hudson estuary during different discharge and tidal conditions. Upper panel: high discharge (2,000 m^3/s) spring tide; middle panel: low discharge (100 m^3/s), neap tide; lower panel: low discharge (100 m^3/s), spring tide. (adapted from Geyer et al., 2000).

oceanographers as practical salinity units or psu), which renders salt water about 3 percent more dense than fresh water. This density contrast causes the fresh water to flow over the salt water and vice versa, leading to an estuarine "salt wedge" (Fig. 3.6). Salt wedges are most evident at the mouths of rivers with weak tidal currents relative to the river flow, such as the Mississippi. The Hudson estuary is notable in that it exhibits a salt wedge structure during neap tides, when velocities are at their fortnightly minimum, but it goes through a remarkable transition to almost well-mixed conditions during spring tides (Fig. 3.6). Other estuaries exhibit this spring-neap change in stratification – it was first noted by Haas (1977) in the Rappahannock Estuary in Chesapeake Bay. However, the Hudson exhibits a more extreme range of stratification between neap and spring tides than any estuary in which this phenomenon has been observed (Geyer, Trowbridge, and Bowen, 2000).

THE ESTUARINE CIRCULATION

Although the vertical salinity gradient varies considerably between neap and spring tides, there is always a strong horizontal salinity gradient along the estuary. This salinity gradient causes a horizontal density gradient (due to the difference in density between fresh and salt water), which in turn induces a depth-varying, or "baroclinic" pressure gradient in the estuary. The baroclinic pressure gradient drives the deep water landward, and a compensating tilt of the water surface drives the surface water seaward (Fig. 3.7). This vertically varying motion is called the estuarine circulation (Pritchard, 1952). The estuarine circulation has the strange property that, at the bottom, is directed toward land against the direction of the river flow. This tendency is counterintuitive, particularly because the estuarine circulation owes its origin to the forcing by the freshwater outflow. The explanation for this is the forcing by the

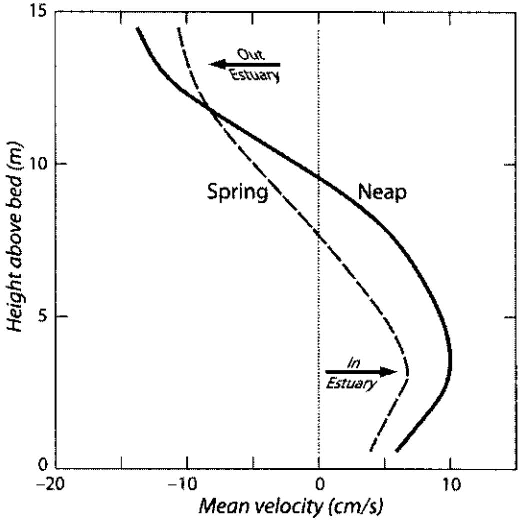

Figure 3.7. Vertical profiles of net estuarine velocity, during neap and spring tides, observed at the Battery in the spring of 1999 (from data presented in Geyer et al., 2001). Stronger estuarine currents occur during neap tides, when tidal mixing is weaker.

density contrast between seawater and fresh water, which yields a landward-directed force at the bottom of the estuary. The tilt of the water surface toward the sea provides a driving force for the surface outflow, but the density gradient is strong enough to reverse the direction of that force at the bottom.

The estuarine circulation is the mechanism that transports salt into the estuary against the outward motion of the river flow. This is accomplished by carrying high-salinity water in at the bottom and carrying out low-salinity water at the surface, resulting in a net inward motion of salt. The estuarine circulation is driven by the density gradient between fresh and salt water, thus the stronger the gradient, the stronger the estuarine circulation. The salinity distribution along the estuary is like a spring: when it is compressed during high river-flow conditions (Fig. 3.5), it exerts a greater force, driving a more vigorous estuarine circulation (Fig. 3.8). During high flow conditions, the seaward transport of salt due to the river is greater; thus a stronger estuarine circulation is required to keep salt in the estuary. When the river flow decreases, the spring relaxes, and the forcing of the estuarine circulation decreases.

STRATIFICATION

The estuarine circulation is not the only factor responsible for the salt transport in the estuary; the vertical salinity stratification is also key. The amount of salt that is transported by the two-way flow depends on the salinity difference between the surface and bottom waters. As that salinity difference increases, the amount of salt that is transported increases proportionately. Perhaps more importantly, the stratification is closely related to the amount of vertical mixing that occurs in the estuary, which in turn regulates not only most of the physical exchange processes in the estuary but also its ecology and biogeochemistry. Thus, stratification is generally considered the most important variable for the classification of estuaries.

Stratification originates from the interaction of the estuarine circulation and salinity gradient. The estuarine circulation always increases the salinity of the deep water and decreases the salinity of the surface water due to horizontal advection (Figs. 3.6 and 3.7). If there were no mixing, eventually the near-bottom water would be purely ocean water and the near-surface water just riverine, with a very strong *halocline*, or salinity gradient, between the two layers. Vertical mixing, due mainly to tidal currents, partially counteracts the stratifying tendency of the estuarine circulation. As tidal currents increase, there is greater vertical mixing and less stratification for a given amount of estuarine circulation (Fig. 3.8). Tidal mixing also has a direct influence on estuarine circulation by increasing the momentum exchange (or drag) between the incoming and outgoing water. Thus, tidal mixing affects the stratification directly, by producing vertical exchange between the upper and lower layers, and indirectly, by influencing the strength of the estuarine circulation (Fig. 3.8), which provides the source of stratification.

THE SPRING-NEAP CYCLE

The sensitive dependence of the stratification on the tides leads to large spring-neap changes in stratification in the Hudson (Figs. 3.6 and 3.8). These changes in stratification indicate large variations in vertical exchange in the estuary. Whereas stratification provides an indicator of the amount of vertical mixing, it also exerts a direct dynamical influence on turbulent motions. The vertical

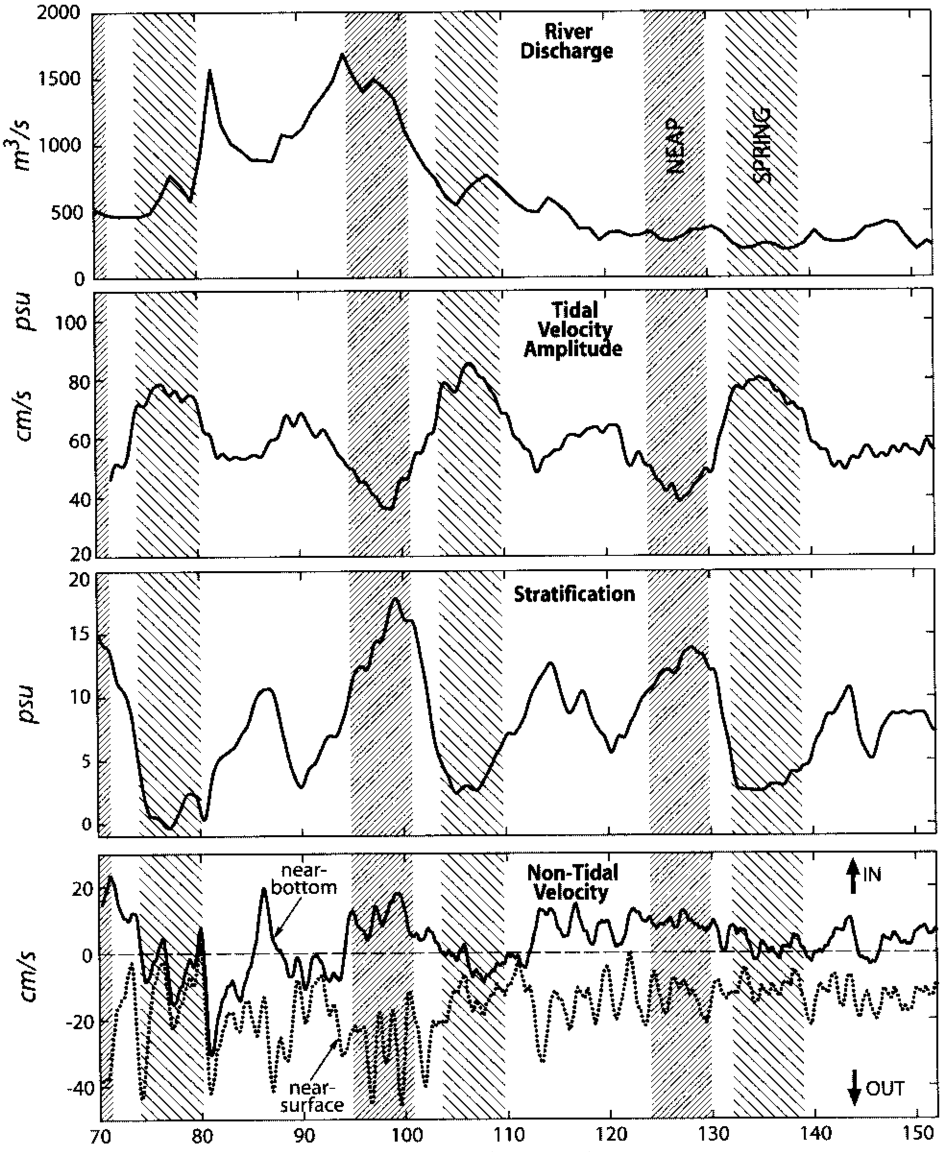

Figure 3.8. Time-series of discharge (top panel), tidal velocity amplitude (2nd panel), stratification (3rd panel), and estuarine circulation (bottom panel) from observations near the Battery in 1999 (from Geyer et al., 2001). Stratification reaches its maximum value during neap tides and its minimum during springs. The estuarine circulation also varies with the spring-neap cycle, but not as distinctly as stratification. Note that the freshwater inflow only has a modest influence on stratification.

density gradient (due to salinity stratification) acts to suppress turbulence, thus preventing the influence of tide-induced mixing from reaching the upper part of the water column during neap tides. This vertical barrier of stratification that occurs during neap tides affects the vertical transport of nutrients and oxygen, with important ecological implications.

These spring-neap variations in stratification also have important implications for horizontal transport of salt. During neap tides, vertical gradients are strong, and there is minimal vertical

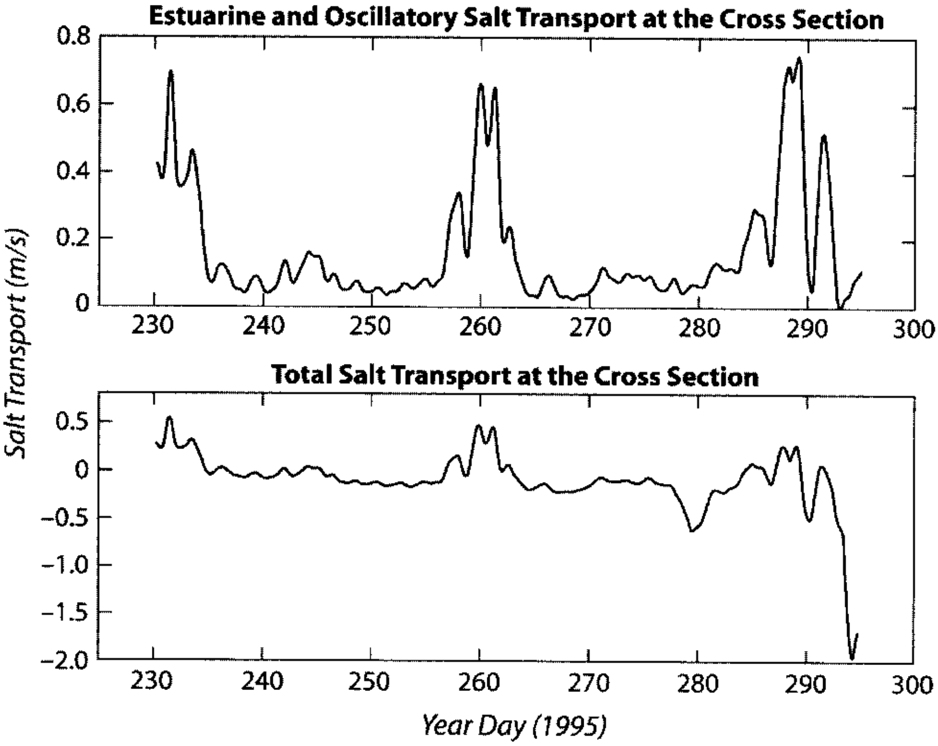

Figure 3.9. Salt flux in the Hudson estuary, during observations in 1995 (from Bowen and Geyer, 2003). The upper panel shows the landward salt flux due to the sum of the estuarine and tidal pumping transport. The lower panel indicates the net transport, including the river outflow and all of the other contributors. Large peaks in landward salt transport occur during weak neap tides, when stratification is maximal. Strong river outflow at the end of the observation period is responsible for the large negative value in the total transport.

exchange of either momentum or salt between the upper and lower layers. Thus, both the estuarine circulation and the stratification are enhanced, and the salt transport due to the estuarine circulation is maximal (Fig. 3.9). This causes salt to advance into the estuary during neap tides and to retreat during spring tides. Whereas the large variation in horizontal salt transport due to the spring-neap cycle is clearly evident, the changes in position of the salinity intrusion are not as obvious. The salinity intrusion is usually long enough that these spring-neap changes in salinity are small relative to the total length of the salt intrusion (Bowen and Geyer, 2003). In addition, variations in stratification may overwhelm the signal of the changes in the horizontal position of the salt front.

TIDAL DISPERSION

It is surprising that the estuarine circulation and river flow would have such important effects on the Hudson estuary, when the tidal currents are so much stronger. The energy provided by the tides far exceeds that provided by any other source in the estuary, and the velocities due to the tides are 5 to 10 times as great as the estuarine circulation and as much as 100 times as great as the river flow. The reason the tides do not totally dominate over these other motions with respect to the salt balance and exchange within the estuary is because of the oscillatory nature of the tidal flow. The *tidal excursion* is the distance that a parcel of water is transported by the tide in one-half cycle. It is calculated by the formula

$$L_T = \frac{T}{\pi} u_T$$

where T is the tidal period (in seconds) and u_T is the magnitude of the tidal velocity. For the tidal currents in the Hudson of 0.7–1 m s^{-1}, the excursion is 10–14 km. The reason that tides are not dominant in

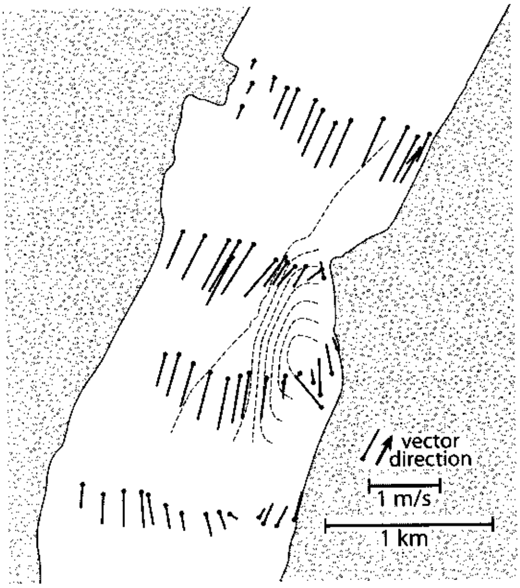

Figure 3.10. An eddy in the tidal stream due to deflection of the ebbing flow by the headland at the George Washington Bridge (from Chant and Wilson, 1997). The sticks indicate the direction and magnitude of the depth-averaged current (with dots at the origin). The eddy results in a salinity anomaly of 3 psu due to trapping in the core of the eddy.

the horizontal exchange in the estuary is that in the other half of the tidal cycle, the water parcel will be transported back roughly the same distance. What makes tides important is their net influence over a tidal cycle, which is due to nonlinearities (i.e., processes that depend on u_T^2).

Tidal dispersion is the net transport accomplished by the asymmetry between the flood and ebb motions that results in net displacement of water parcels over a tidal cycle. Tidal dispersion arises from a number of different mechanisms, most of which are associated with differences in the strength of the tidal current across the estuary. These processes can collectively be regarded as shear dispersion. Shear dispersion occurs both due to lateral and vertical variations in tidal velocity. Its magnitude is dependent not only on the cross-sectional variations of velocity; it is also dependent on the rate of mixing either in the vertical or transverse direction. The flow around headlands can produce eddies that enhance the transverse shears and thus increase the tidal dispersion (Fig. 3.10). Rarely, however, does tidal shear dispersion reach the magnitude of exchange induced by the estuarine circulation (Zimmerman, 1986).

Other, more complicated types of dispersion can occur due to interactions between the tides and the estuarine circulation. The estuarine circulation and its associated salt flux are defined based on tidal averages of the flow and the salinity, but there can be correlation between variations in velocity and salinity that lead to net salt transport. In regions of irregular topography, these transports can exceed the strength of the estuarine circulation (Geyer and Nepf, 1996).

TIDE-INDUCED MIXING

As discussed in context with the estuarine circulation, one of the most important nonlinear processes accomplished by the tides is the generation of turbulence. The generation of turbulence at the bottom of the estuary is well understood: the flow over the rough bottom produces eddies that diffuse momentum and water properties in the vertical dimension. The turbulence problem becomes more complicated farther up in the water column, where stratification is stronger. Stratification tends to suppress turbulence associated with bottom-generated turbulence, but as that turbulence is suppressed, the shears tend to increase. Once the shears get high enough relative to the strength of the stratification, a new source of turbulence, shear instability, can start mixing within the stratified water column (Peters, 1997). Shear-induced mixing is important in the Hudson during neap tides and times of high flow, when stratification is strong. The complex interactions between tidal currents, shear-induced mixing, and internal waves are not yet fully understood, and these interactions represent an important aspect of estuarine dynamics that limits our ability to model estuarine physical processes.

NEW YORK HARBOR

The character of the estuary changes at the Battery, where the Hudson River joins the East River at New York Harbor. In contrast to the simple morphology of the Hudson, the Harbor has a complex geometry, with interconnections between several adjacent embayments through a series of tidal straits (Fig. 3.11). The flow in these straits, among the swiftest in the harbor complex, are driven primarily

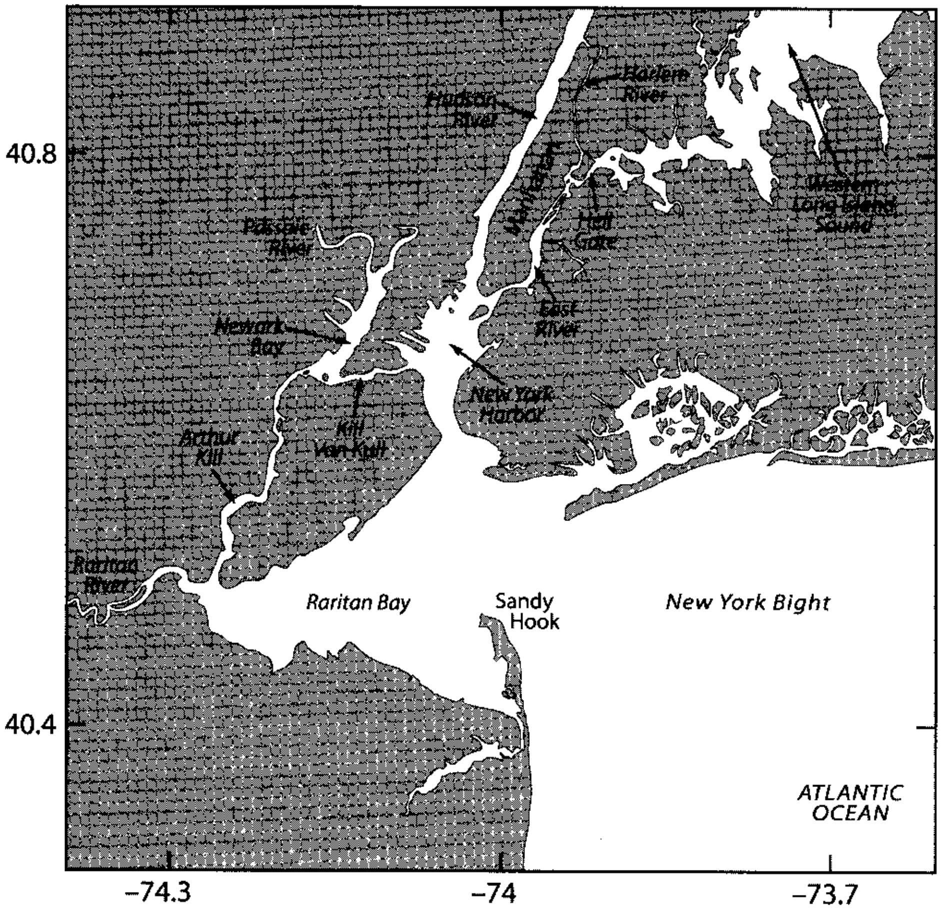

Figure 3.11. Map of New York Harbor complex and western Long Island Sound. The Battery is at the southern tip of Manhattan.

by sea level differences between different water bodies due to tidal and meteorological forcing. Weaker, but persistent two-layer flow is also driven by a salinity difference at the ends of the straits and by non-linear tidal dynamics.

THE EAST RIVER

Despite its name, the East River is not a river but rather a tidal strait, for it has no significant natural direct source of fresh water (in fact sewage outflows are the largest direct source of "fresh water" to the East River). Tidal currents in the East River are among the strongest in the region because of a remarkable difference in the amplitude and timing of the tide between Long Island Sound and New York Harbor. Tides in western Long Island are nearly 70 percent larger than those in the Harbor, and the time of high and low water occurs over 3 hours later in western Long Island Sound than in the Harbor (Fig. 3.12). This oscillating sea level slope drives 2 m s^{-1} tidal currents in the East River, and the notorious tidal currents at Hell Gate (at the junction of the East and Harlem Rivers) can exceed 3 m s^{-1}.

Weaker but lower frequency flows in the East River are also driven by sea level slopes set up by a difference in the wind-driven response of the Harbor and Western Long Island Sound (Wilson, Wong, and Filadelfo, 1985). These flows fluctuate with winds that typically vary at a 2–5 day time scale. While these flows are an order of magnitude weaker than the tidal currents, they are more persistent and may significantly contribute to the exchange between the Sound and the Harbor.

A mean salinity gradient exists along the East River with bottom waters in Western Long Island

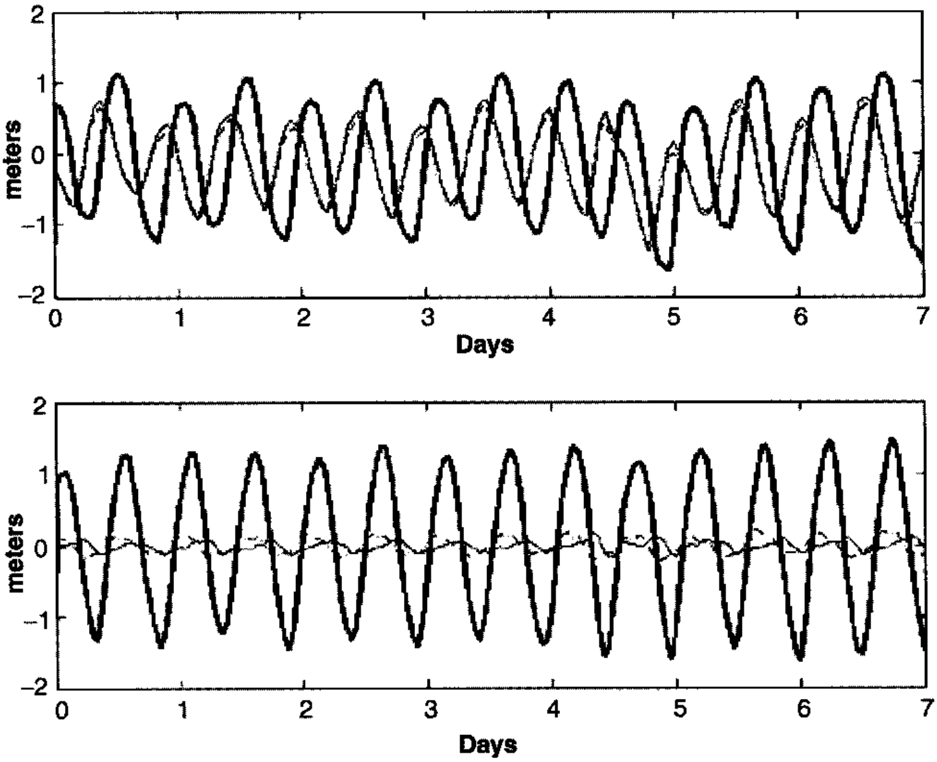

Figure 3.12. A) Upper panel, hourly sea level from Western Long Island Sound at Willets Point (thick line), The Battery (dashed line), Sandy Hook (dotted line) and the western Kill Van Kull at Bayonne (thin line). B) Lower panel, Sea level difference between the Battery and Willets Point (thick line), Sandy Hook and the Battery (dashed-dotted line), and Bayonne and the Battery (thin line).

Sound on average 4 psu more saline than those in the Harbor (Blumberg and Pritchard, 1997). Strong tidal currents in the lower East River maintains a well-mixed water column, while salinity stratification in the upper portions of the strait near Willets Point tends to be about 2 psu in the vertical dimension (Blumberg and Pritchard, 1997). Mixing is strong enough that the mean flow tends to be unidirectional throughout the water column. Yet there is debate on both the magnitude and even the direction of the mean flow. A number of investigators (Blumberg, Khan, and St. John, 1999; Blumberg and Pritchard, 1997; Jay and Bowman, 1975) estimate a mean flow of about 300 m^3 s^{-1} from the Sound into the Harbor. However, Filadelfo, Wilson, and Gomez-Reyes (1991) report persistent flow in the opposite direction.

THE KILLS, NEWARK BAY AND RARITAN BAY

Like the East River, the Kill Van Kull and Arthur Kill are tidal straits. Maximum tidal currents reach m s^{-1} in the narrowest reaches of these straits and attenuate to less than 0.5 m s^{-1} in Newark Bay. Tidal excursions in the Kill Van Kull are greater than the length of the channel, thus tidal motion is effective in mixing water between Newark Bay and New York Harbor, particularly during spring tides. In contrast, tidal excursions in the Arthur Kill are significantly shorter than the length of the tidal strait; thus, tides are an ineffective agent driving exchange between Newark Bay and Raritan Bay.

The Raritan River and Passaic River are the major direct sources of fresh water to Raritan and Newark Bays, respectively – both with mean annual discharges of 50 m^3 s^{-1}, with peak flows in the spring of 100–300 m^3 s^{-1}. This fresh water drives a two-layer exchange in the tidal straits and Newark Bay. Salinity stratification and two layer exchange is persistent in the southern reaches of the Arthur Kill.

Meteorological forcing drives flow through the Kills both by the direct action of the wind on the water's surface and by producing a difference in the water levels at the ends of the tidal straits. Similar to what occurred in the East River, these meteorologically forced flows tend to last for several days, and potentially are an effective means to exchange water fluid between Newark Bay and Raritan Bay (Blumberg et al., 1999; Chant, 2002).

The New York Bight and the Coastal Current

In total, the Harbor system discharges annual mean flow amounting to over 700 m^3 s^{-1} of fresh water to the New York Bight. Approximately 600 m^3 s^{-1} of this is from the Hudson, Raritan, and Passaic Rivers, with an additional 100 m^3 s^{-1} from sewage outflows. In addition, the estimated 300 m^3 s^{-1} transported through the East River augments this flow and yields a mean volume transport leaving the Harbor complex through the Sandy Hook-Rockaway transect of approximately 1,000 m^3 s^{-1} – nearly double the discharge of the Hudson.

The flow through the Sandy Hook-Rockaway transect, like the flow throughout much of the Harbor system is two layered, with the surface layer flowing seaward and the lower layer flowing landward. Thus, the transport of fluid in the upper layer must also compensate for the inflow in the lower layer. Based on salt conservation at the Sandy Hook-Rockaway transect, an annual mean outflow of approximately 3,500 m^3 s^{-1} of estuarine water enters the New York Bight, with approximately 2,500 m^3 s^{-1} of saline waters from the Bight entering into the Harbor. These transports significantly vary at weekly, monthly, seasonal, and interannual time scales.

Once past the Sandy Hook–Rockaway transect, the estuarine water from New York Harbor forms a coastal current that flows south along the New Jersey shore. The tendency for the current to head southward is due to the effect of the earth's rotation, or the "Coriolis effect," which turns the fluid to the right in the absence of other forces. Winds also play a major role in defining the structure and direction of the outflow. Southerly winds (winds *from* the south) spread the plume offshore causing it to thin and may arrest its southward flow. Northerly winds compress the plume against the coast and augment its flow to the south. As the plume is transported south along the New Jersey coast it continuously mixes with the more saline shelf waters in the coastal ocean. The mixing is primarily wind driven, while the weak tidal currents, that tend to be less than 15 cm/s along the New Jersey inner shelf, play a secondary role. The Hudson's coastal current has been observed along southern New Jersey near Cape May, more than 150 km south of the Battery (Yankovsky et al., 2002). Eventually, the signature of the Hudson's freshwater flow is lost south of Cape May, New Jersey, where its plume becomes obscured when it mixes with waters from Delaware Bay.

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